SEDIMENTOLOGICAL AND STRUCTURAL EVOLUTION OF TERTIARY BASINS

OF ALPES-DE-HAUTE-PROVENCE, S.W. ALPS

by

ALUN HUW DAVIES B.Sc. (CARDIFF)

CONTAINS PULLOUTS

Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor in Philosophy by Alun Huw Davies, December 1988.

This thesis is dedicated to Dilys

ABSTRACT

The Tertiary Durancian basins of Alpes-de-haute Provence lie to the immediate west of the S.W Alpine thrust belt and to the east of the Rhone Graben. They provide an opportunity to study the degree and nature of tectonic control on sedimentation in the autocthonous S. W alpine foreland.

The Tertiary basins comprise a series of erosional remnants which range in size from a few to some 400 sq. km. The basins are predominantly filled by continental and shallow marine clastic successions and are of up to 2000m thickness. They developed upon a pre-folded Mesozoic substrate which had an inherited framework of high-angle Hercynian - Mesozoic crustal faults.

Detailed facies analysis of the basin-fill successions highlighted the facies patterns and sediment transport paths in the basins. This was integrated with compositional petrographies and structural analaysis to enable the evolution of the basins to be considered in the context of contemperaneous structural events.

Lateral thickness variations show that the Durancian basins developed within a system of differentially subsiding fault blocks defined by reactivated inherited faults. Facies analyis of the basin fills shows that they had a three stage fill which may be related to a progressive change from an extensional to transpressional regime in the alpine foreland as the alpine thrust belt migrated southwestward toward the Western Mediterranean rift system.

An initial Priabonian - Aquitanian continental stage was characterised by the extensional reactivation of inherited faults by the Western Mediterranean rift system. Deposition of extra-basinal (alpine) and intrabasinal sourced terminal alluvial fans occured within a system of NE-SW trending horsts and grabens. In the north-east of the Durancian basin alpine sourced, mixed-load alluvial fans terminated in extensional grabens adjacent to the alpine thrust front. In contrast intra-basinally sourced mass-flow alluvial fans developed in NE-SW trending grabens in the south-west of the Durancian basin.

An Aquitanian - Vindobonian marine stage involved the

development of tide-dominated seaways and gulfs confined within transtensional fault blocks as the Mediterranean flooded northward in response to a Burdigalian eustatic transgression accompanied by regional subsidence. Alpine sourced, tide and wave influenced, siliciclastic shorelines developed adjacent to the alpine thrust belt which passed westward into bioclastic shelf systems.

A Messinian-Pliocene continental stage was coincident with a eustatic sea level fall and the translation of the alpine thrust belt to the eastern margins of the present day foreland basin. It was marked by marine regression and the development of an alpine sourced braided river system confined within a transpressional basin adjacent to the alpine thrust front.

ACKNOWLEDGEMENTS

Perhaps the hardest part of any thesis is getting the acknowledgements right and remembering all those who made its completion possible.

Firstly I would like to thank Trevor Elliott for supervising this study and in particular for his patience in editing the draft versions of the thesis.

For enlightening discussions on alpine geology, thanks also go to Drs. R. H. Graham and J. Hossack.

Martin Evans and Gill Apps were a big help in the 'field' and also provided early ideas on alpine geology and how to survive field seasons.

A 'merci beaucuop' to Guy and all at the field centre, Digne for hospitality during field-work and Alex for the loan of his bike !

At Liverpool:

Thanks to Steve Crowley and Colin Howard for general discussion. Special thanks must go to the 'back-up' crew who helped in so many ways with the production of this book, and who were; Nick, Nick (tell us a joke), Rowser, John, Rina, Steve, Greg and especially to the "girls"; Cli, Debbie, Hannah and Rowena. In addition, thanks to the "late-night club" who made the writing up such a team affair.

In the Department of Earth Sciences, thankyou to all of the postgraduates past and present whom I have known and who have made my stay in Liverpool such an enjoyable one.

My extra special thanks go to my Dad, for his encouragement and support during my years at college, both in Liverpool and Cardiff.

Finally I would like to give a big thankyou to Rowy for her untiring help throughout the days working up to this thesis and for invaluable discussions on the Barreme Basin.

CONTENTS

GRID REFERENCES FOR KEY FIGURE LOCALITIES.

Grid references are detailed as international latitude and longitude as given on I.G.N 1:25 000 maps.

FIGURE.	GRID REFERENCE.	
	Deg min sec	Deg min sec
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6.8	06 03 55	44 13 10

Aims of Study

This thesis is an attempt to elucidate the sedimentological and structural evolution of autochthonous Tertiary basins, informally termed the Durancian basins, of Alpes-de-Haute-Provence in the South-Western Alps.

The specific aims of the thesis are:

1) to provide descriptions and interpretations of the facies associations of the Tertiary (upper Eocene - upper Miocene) in terms of their depositional processes and environments of deposition.

2) to further provide descriptions and interpretations of the lateral variability in facies associations and to erect depositional system models for the Tertiary.

3) to integrate the above data with lateral thickness variations in the Tertiary basin fills, and structural features within the Mesozoic and Tertiary formations and erect a linked structural and sedimentological model for the Durancian basins. Specifically the degree of control of alpine thrust tectonics and extensional foreland tectonics on the sedimentary history of the autocthonous alpine foreland will be considered.

CHAPTER 1.

INTRODUCTION, REGIONAL AND TECTONIC SETTING

1.1 Regional Setting

The South East of France contains a number of remnant Tertiary basins which vary in size from a few sq. km to some 1800 sq. km in area, and which form a part of a system of Western European Tertiary Basins extending northward from the Rhone Graben, into the Swiss Molasse and Bresse Graben (Figs. 1.1 & 1.2). One of the largest Tertiary basins in S.E France is the autochthonous, clastic-dominated, **Durancian Basin** (Figs. 1.1 & 1.2 (i-iii). This thesis is concerned with the evolution of a series of small remnant basins which form the northern margin of the Durancian basin, and which lie in the Alpes-de-Haute-Provence region of the External South-Western Alps (Fig. 1.2 - 1.3 (i)).

1.2 Structural Zonation of S.E France

The structure of S.E France can be considered in terms of two regional systems namely the **South-Western Alps,** and the **Rhone Graben** (Figs. 1.1 - 1.3).

1.2.1 The South-Western Alps.

The Alps can be classified into a series of major Alpine structural zones which are essentially coincident with the palaeo-geographic domains of the Mesozoic continental passive margin (Figs. 1.3 - 1.4) (after Lemoine 1978). Internal and External Zone's are recognised which are structurally sub-divided

(Fig. 1.4) into:

(i) The **Internal Zone** which comprises:
(a) the Liguro-Piemont Zone - developed over oceanic and highly attenuated continental crust.
(b) Brianconnais (including Sub-brianconnais) zones developed over weakly attenuated continental crust.

(ii) The **External Zone** of the South - Western Alps is structurally sub-divided (after Haccard *et al.*, 1980) into:
(a) the allocthonous sub-alpine chain;- defined as the allochthonous Mesozoic - Tertiary sequences lying east of the Digne thrust front and west of the Internal Zone.
(b) the autochthonous sub-alpine chain;- defined as the autochthonous / parautocthonous Mesozoic - Tertiary cover sequences lying west of the alpine thrust front and east of the Rhone Graben.
(c) the autocthonous / parautochthonous crystalline massifs - the Argentera, Barrot and Pelvoux massifs.

1.2.2 The Rhone Graben and Alpine Foreland.

The dominant structures of this region are a series of major NE - SW trending, high-angle crustal faults which extend for some 100 - 300 km (Figs. 1.2 & 1.3). The principal faults are the Cevennes, Nimes, Lamanon, and Durance faults, which are interpreted in this thesis to define four regional crustal blocks (Fig. 1.3 section (iii)). The latter three faults may be traced north-eastward into the autochthonous sub-alpine chain, and crystalline massifs suggesting that they are also fundamental structures in these alpine areas.

1.3 Tertiary Tectonic Regimes.

1.3.1 Introduction.

During the Tertiary the S.E of France was characterised by the coeval existence of major compressional and extensional deformation belts (Fig. 1.3).

To the east, the **South-Western Alps**, a classic thrust - fold belt, developed during the Alpine compression of the passive European continental margin of the Ligurian Ocean (Mesozoic

Tethys) in the late Cretaceous - Tertiary (Debelmas *et al.*, 1970; Lemoine 1980; Debrand-Passard 1984 and Lemoine *et al.*, 1986). To the west the Alps were fringed by the Rhone Graben, which formed a part of the Western Mediterranean Rift System, initiated during the Oligocene (Ziegler 1982).

1.3.2 S.W Alpine Deformation History

Alpine compressional tectonics began during the early Late Cretaceous (about 80my), with the westward thrusting of the Ligurian - Piemont oceanic crust onto the Piemont domain, the eastern margin of the European continental margin (Tricart 1984) (Figs. 1.4 & 1.5). The late Cretaceous was subsequently marked by the 'Eoalpine' phase of compressional deformation affecting the whole Piemont zone and its cover of Ligurian - Piemont nappes. Alpine compressional deformation translated into the External Alps in the early Tertiary. The S.W alpine thrust system was south westward directed with propagation being of a pulsed, discontinous nature (Tricart 1984 and Graham 1985) (Fig. 1.5). The tectonic style was that of "decollement" tectonics with Mesozoic and Tertiary sedimentary sequences detached from the Hercynian basement along Triassic evaporitic horizons.

The first phase of alpine deformation predated the Upper Lutetian (47 - 45 Ma) and ended in the late Middle Eocene (Graham 1985). A 4 Ma period of tectonic stability in the External Zone separated this from the ensuing Oligocene, Sannoisian deformation phase (Graham *op. cit*). A third deformation phase occurred in the late Oligocene. Timing constraints on subsequent alpine deformation are poorly defined, but a fouth deformation phase resulted in the overthrusting of the eastern margin of the autochthonous Digne-Valensole basin (eastern margin of the Durancian Basin) (Fig. 1.3) during the late Pliocene - Quaternary (Gigot *et al.* 1974).

1.3.3 Western Mediterranean Rift System Extensional History

The Western Mediterranean Rift System (Fig. 1.6) was initiated in the lower Oligocene, in response to a shift in the trajectories of the African and European plates (Ziegler 1980) and consequent develop ment of an extensional regime between the European plate and the Corsica - Sardinia micro-plates (Fig. 1.7

from Rehault 1984). The rift system had a general extension direction of NW - SE (Bessis 1986).

The extensional history of the system was twofold (Fig. 1.8) (Burrus 1984, and Bessis 1986), namely:

(i) An **active rift phase** :- This was initiated at the base of the Rupelian (35 Ma) and extended through to the lower Aquitanian (23 - 24 Ma).

(ii) **Thermal subsidence phase**:- Active rifting was succeeded in the lower Aquitanian (23 - 24 Ma) by oceanic accretion and associated regional thermal subsidence.

Subsidence across the region during rifting was not uniform as Hercynian age basement faults were reactivated. No major system of extensional faults was initiated during the Tertiary (Goguel 1944 & 1948) rather the lithosphere more efficiently accomodated the Tertiary extension through the reactivation of inherited high-angle lithospheric faults (see chapter 2). As a consequence a system of Tertiary horsts-grabens was developed, primarily defined by NE-SW trending high angle faults, and discussed in detail in chapter 2.

1.4 The Durancian Basin

The autochthonous Durancian Basin (Fig. 1.1 - 1.2) is developed immediately north of the Mediterranean Sea, and is defined to the east by the front of the S.W Alpine thrust-fold belt. This thrust front structurally separates the autochthonous Durancian basin from contemporaneous, allochthonous thrust-sheet top basins developed to the east. To the west the basin is defined by the major NE-SW trending Lamanon Fault (Fig. 1.2 & 1.4), which structurally separates it from the Tertiary basins of the Rhone Valley. To the north the margin of the basin is defined by the termination of Tertiary outcrop.

Internally the basin is transected by a series of high-angle faults which subdivide it into a series of fault blocks (see section 2.4), upon which sub-basins of the order of 50-800 sq km scale were developed. The faults form two sets which trend NE-SW and E-W, and which together with the Lamanon Fault, are interpreted as reactivated crustal faults (see chapter 2).

As a consequence of differential movement on these faults during the Tertiary, the geometry of the sub-basins within the

Durancian Basin changed. This can be related to changes in the tectonic regime superimposed on the Durancian Basin, with three distinctive sub-basin configurations recognised (Fig. 1.9), corresponding to the:

(i) upper Eocene - Oligocene period.

(ii) lower and middle Miocene period.

(iii) upper Miocene - Pliocene period.

1.5 Previous Work.

1.5.1 Structural Interpretations of the Durancian Basin.

The Tertiary basins of S.E France have been interpreted as belonging to either an extensional basin system, termed the **Rhone Graben** (Rhodanian basins), or to a foreland basin system, termed the **South-West Alpine foreland basin** depending upon their geographical position (Fig. 1.2). The margins of these systems are however poorly defined, and an area of interpretational overlap exists in the region of the Durancian basin (Fig. 1.2).

The Tertiary basins of the External Zones of the S.W Alps have long been considered to be related to the development of the alpine mountain chain and have been described as **flysch** and **molasse** basins (de Lapparent 1938). The refining of this interpretation by considering the basins as part of the one larger **South West Alpine foreland basin** (Fig. 1.2) which evolved in conjunction with the alpine thrust - fold belt has been taken by Elliott *et al.* (1985). The foreland basin is considered in terms of a number of allocthonous and autochthonous remnant sub-basins which sequentially acted as depositional centres during its south-westward migration. In a westerly sense these sub-basins are the allochthonous Gres d'Annot and Barreme basins, and the autochtonous Digne - Valensole basins (Figs. 1.2, Fig. 1.3 (ii), and 1.10). In this interpretation the Digne-Valensole basin, which forms the north-eastern margin of the regional Durancian Basin, is considered to have been the late Tertiary depocentre of the foreland basin.

The principal evidence supporting a foreland basin interpretation for these remnant basins is (1) that the Tertiary basins become younger in a south-westward direction parallel to the principal direction of alpine thrusting (Fig. 1.10), thus suggesting that the depositional centre of the foreland basin migrated in concert with the thrust belt (Fig. 2.8). (2) individual formations which are traceable between some the remnant basins show a westward younging. (3) the sedimentary fill within individual basins shows an intimate relationship with the thrust and fold belt develop ment as evidenced by a multitude of syn-tectonic depositional features (Evans 1988).

In contrast, the Tertiary remnant basins of the Rhone Valley (Fig. 1.2) have traditionally been interpreted as extensional basins which developed within the **Rhone Graben rift system** (Goguel 1944, 1948) through the Tertiary reactivation of Hercynian age crustal faults (Goguel *op.cit*). This extensional basin interpretation was extended to the basins of Provence by Montenant (1969) who considered the Manosque-Forcalquier basin (a part of the regional Durancian basin: see Fig. 1.9) to have been an extensional basin controlled by the reactivation of a major "Mesozoic" fault, the Durance fault. Subsequent work, notably by Gubler *et al.*, (1975), and Cavelier *et al.* (1984), has supported this extensional basin interpretation of the Manosque-Forcalquier basin.

1.5.2 Previous Sedimentological Work.

The northern Durancian basins have not previously been the subject of any detailed sedimentological analysis. Work has been published on sections of the basins fills and this is detailed in the preface to the appropriate chapters.

1.6 Stratigraphy

On a detailed scale the stratigraphy of the Durancian basin is complex in that it involves the fills of a series of differentially subsiding sub-basins. However in terms of its gross fill the regional basin has a basic three-stage stratigraphy, namely an Eocene - lower Miocene continental succesion, a lower-middle Miocene marine succession, and an upper Miocene - Pliocene continental succession.

In the northern Durancian basins, the absence of any recent biostratigraphical work means that the simple bio-stratigraphy erected by Goguel (1964) (see Fig 1.11) is adopted:

3) Messinian - Pliocene: - **Valensole Formation**; comprising massive continental conglomerates.
2) Burdigalian - Vindobonian: - **Marine Molasse Formation**; comprising marine sandstones, mudstones and conglomerates.
1) Priabonian - Aquitanian: - **Molasse Rouge Formation**; comprising continental conglomerates, sandstones, mudstones and evaporites.
The biostratigraphy is refined through the recognition of lithostratigraphic members (Fig 1.11) which are discussed in the relevant chapters.

A Tertiary stratigraphic column is given in Fig. 1.12.

1.7 Techniques

This project took the form of a field-work based study. Sedimentological data was collected at various localities and recorded by a standard graphic logging technique. Field grain size estimates were made with reference to the Wentworth scale chart. The colour of rocks was described with a Munsell colour chart. Structural data was collected by field mapping. Structural and sedimentological observations were supplemented with data obtained from the interpretation of B.R.G.M geological maps, namely:

(i) 1:250 000 - Valence sheet, Nice sheet, Marseille sheet, Gap sheet
(ii) 1:80 000 - Digne sheet, Sederon sheet, Gap sheet
(iii) 1:50 000 - Digne sheet, Forcalquier sheet, Seyne sheet.
Laboratory techniques were limited to the petrological analysis of thin sections under a Vickers microscope.

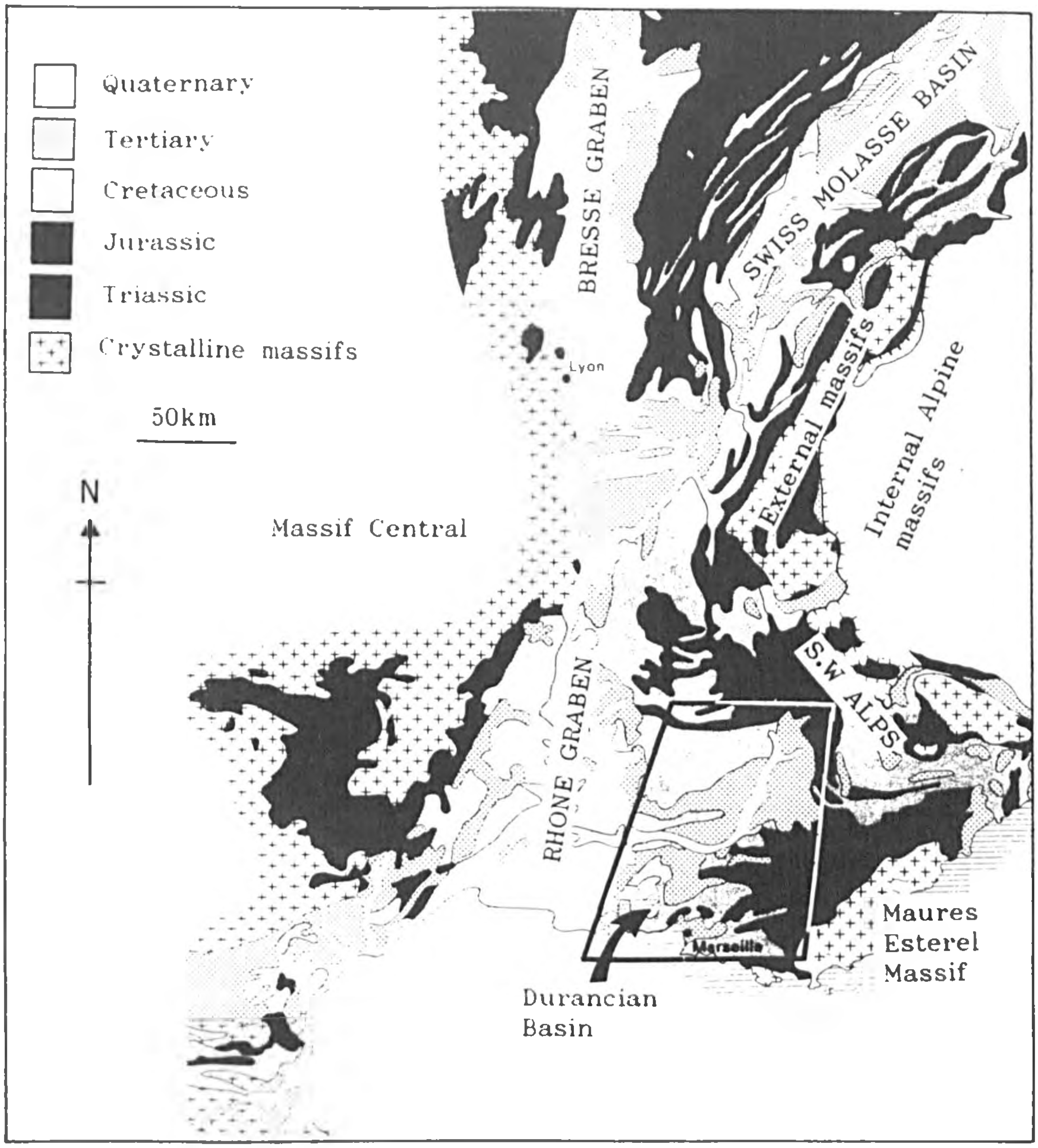


FIG 1.1 Simplified geological map of S.E France.

Alpine thrust front

Hercynian crystalline basement

Tertiary remnant basins

Durancian Basin limits

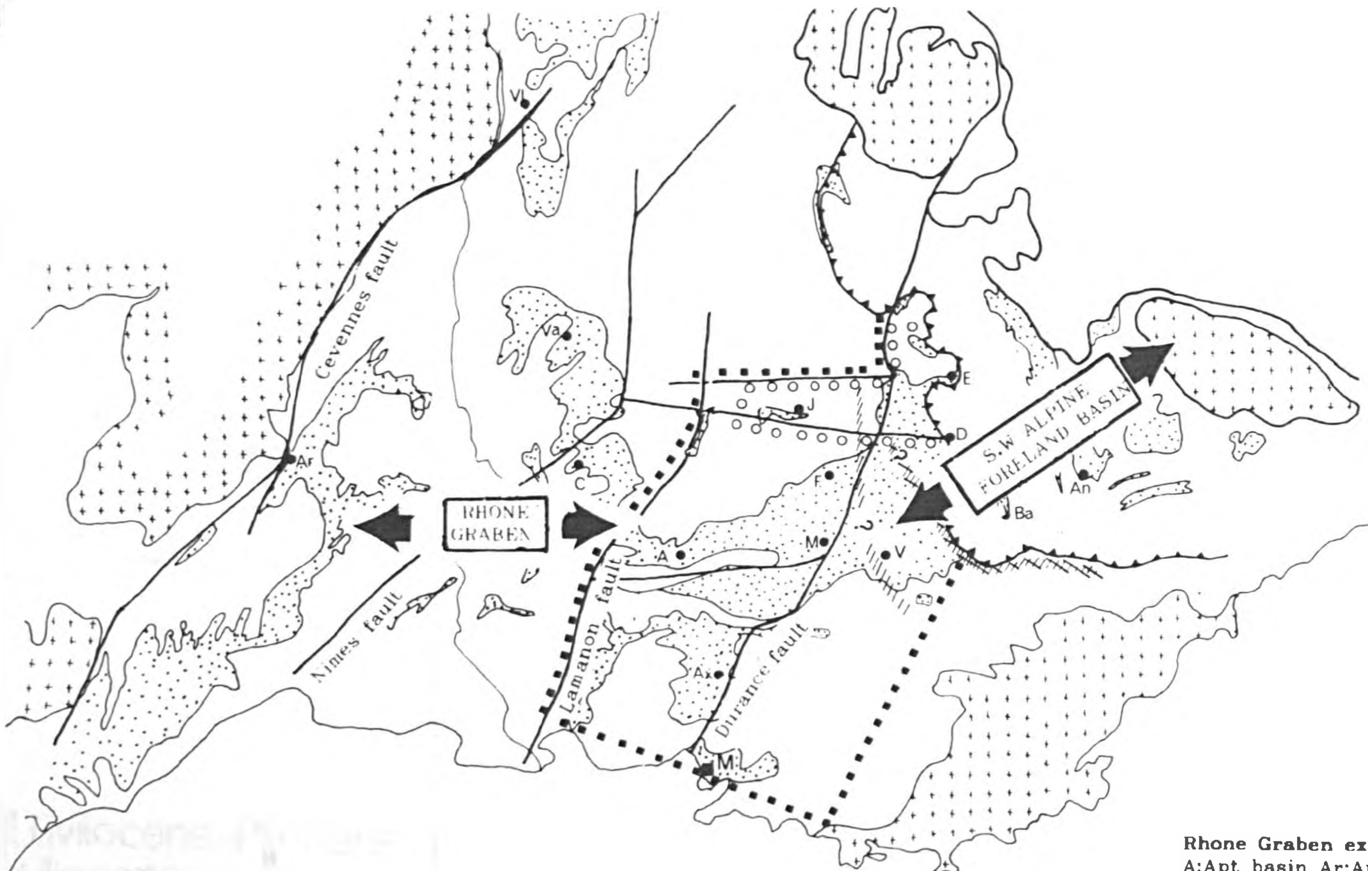
Northern Durancian basins

Major high-angle faults.

Eastern limit of interpreted extensional basins

Western limit of interpreted foreland basin

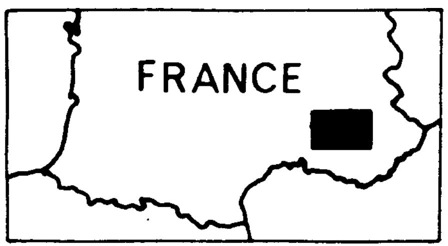
i

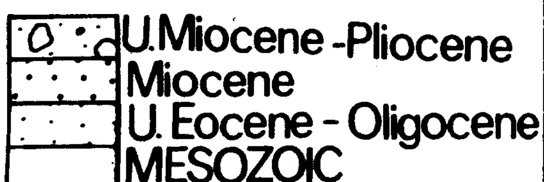


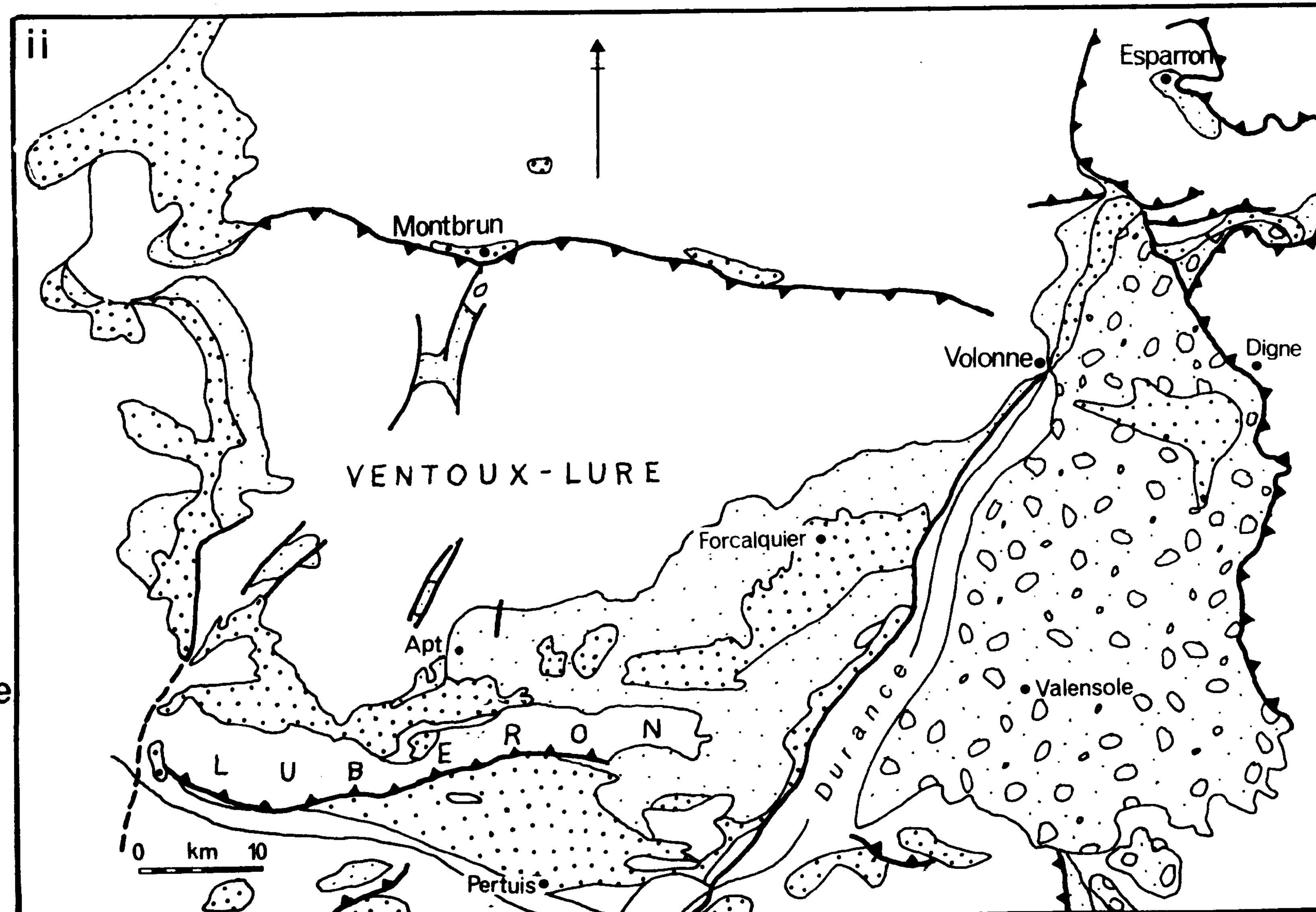
Rhone Graben extensional basins:-
A:Apt basin Ar:Arles basin Ax:Aix basin C:Carpentras basin
M:Marseille basin M-F:Manosque Forcalquier basin Va:Valreas basin
Vl:Valence basins (after Cavelier *et al.* 1984).
S.W Alpine foreland basins - An:Gres d'Annot basins
Ba:Barreme basin D-V:Digne Valensole basin (after Elliott *et al.* 1985)
Uninterpreted:
(J):Jabron basin

FIG 1.2 Tertiary outcrop of S.E France.

Fig 1.2A Geological map of the Durancian basin.







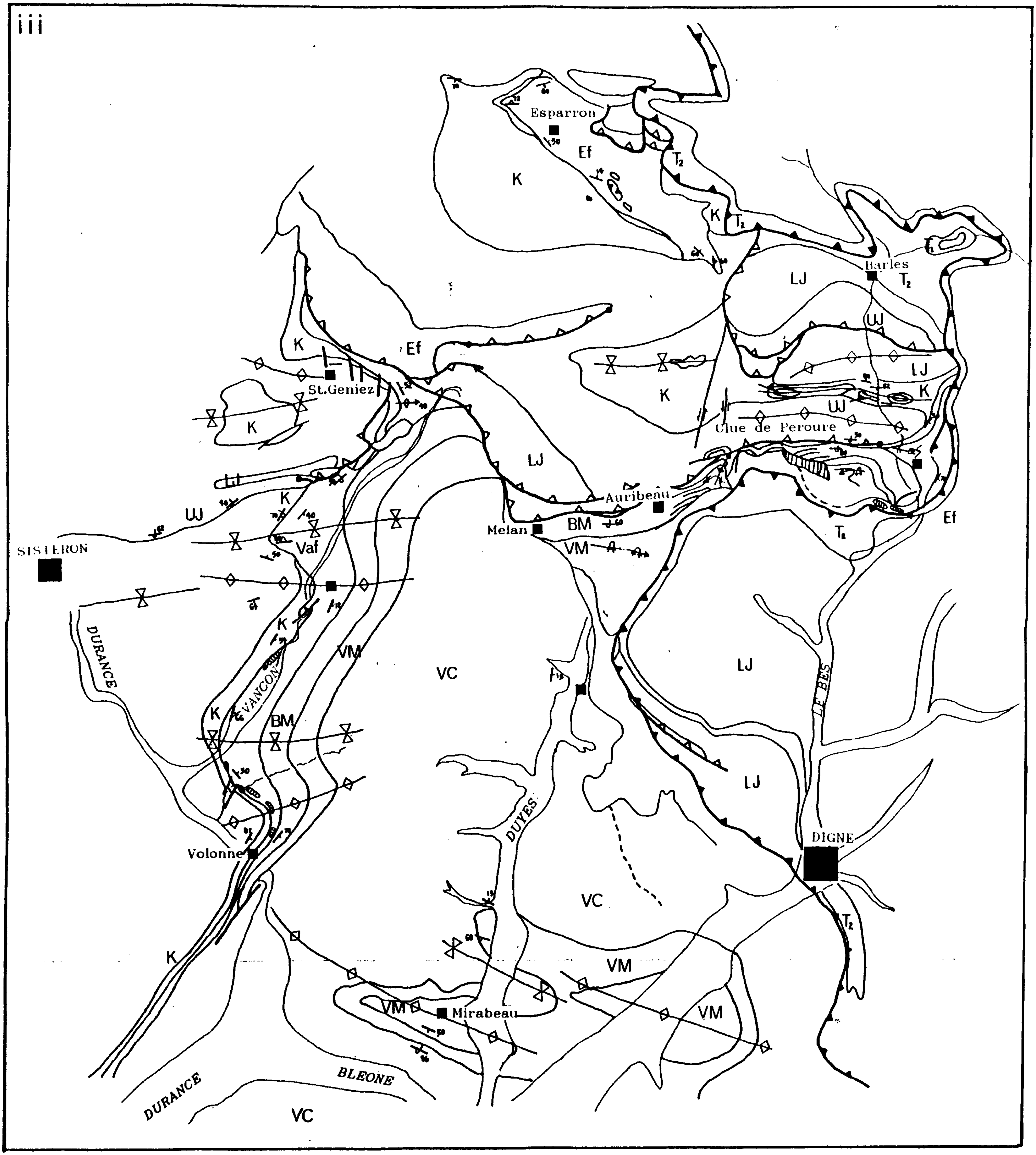
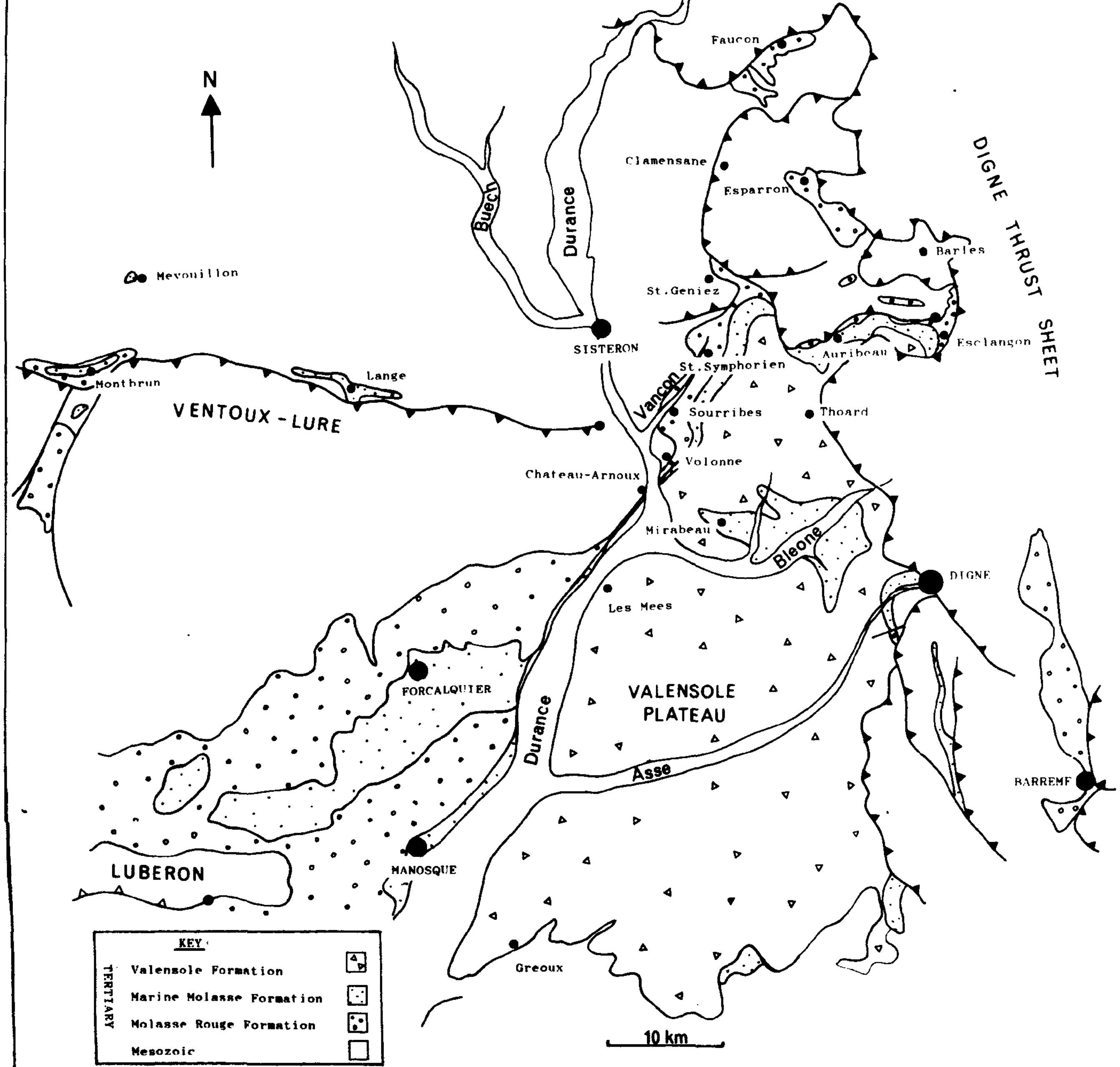


fig 1.2 GEOLOGICAL MAP OF THE DIGNE AREA (adapted from Goguel 1964).

Figure 1.2 (iv). Locality Map - a simplified geological map of the Durancian basin showing the location of the principal sites of study.



Section Lines: Section (i) Internal Zone -> Autochtonous Tertiary basin (Digne-Valensole basin) (from Debelmas 1980)

Section (ii) Allocthonous External Zone -> Autocthonous Tertiary basin (Digne Valensole basin (DV)) Atf: Alpine thrust front (from Graham 1985).

Section (iii) Rhone Graben and autochthonous alpine foreland (from ECORS Preliminary Report *in. prep. C.Howard pers comm*)

Internal Zone Ligurian-Piemontais (L-P)
Brianconnais (inc. Sub-brianconnais) (B)

External Zone (E) Allocthonous- east of alpine thrust front
Autchonous - west of alpine thrust front.

Rhone Graben (R)

Alpine thrust front

Minor thrusts in autochthonous external alps

Limit of Durancian Basin

Major high angle faults

Tertiary outcrop

Hercynian crystalline basement

Di: Digne M:Marseille Ba: Barreme basin GA: Gres d'Annot basin.

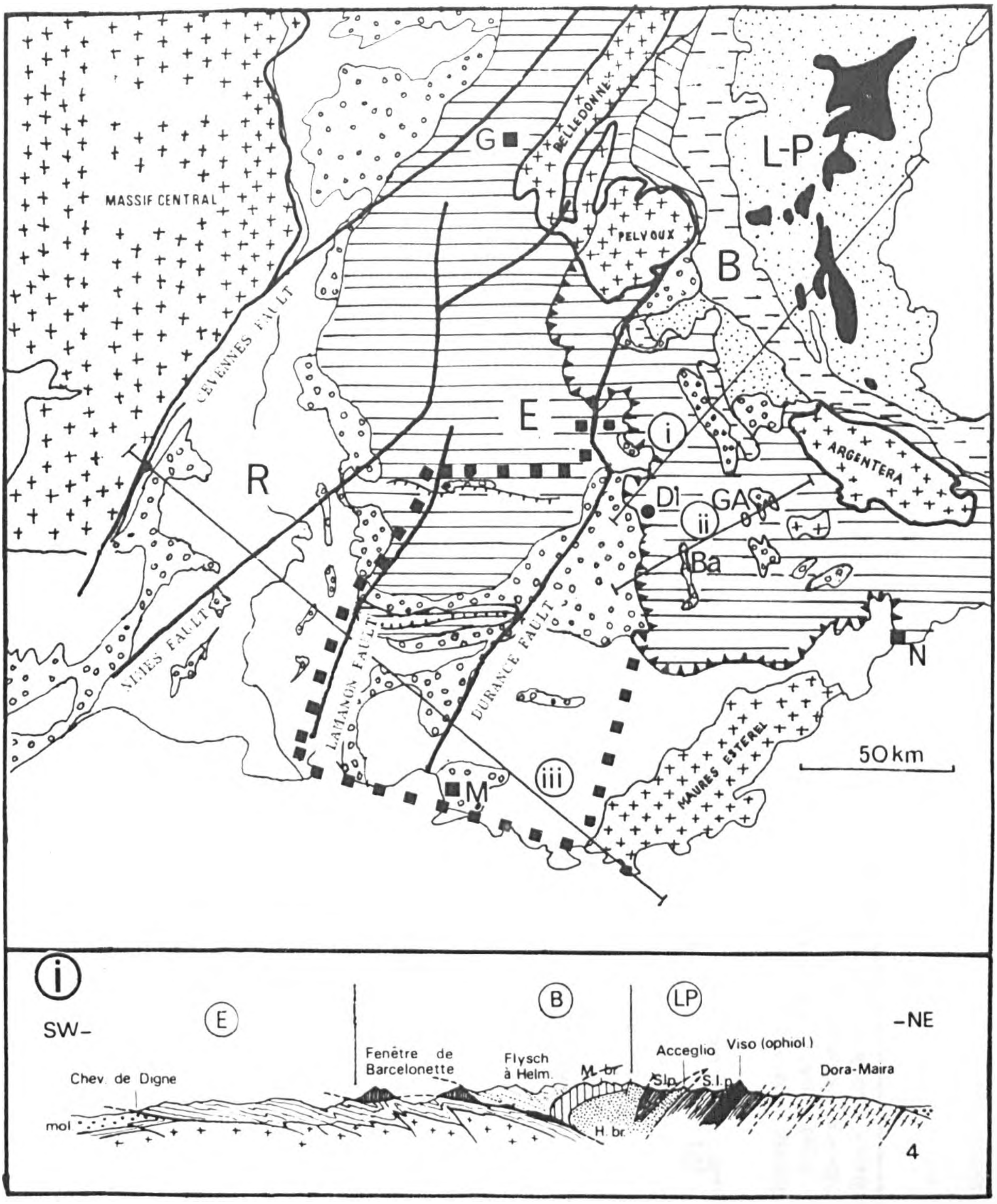
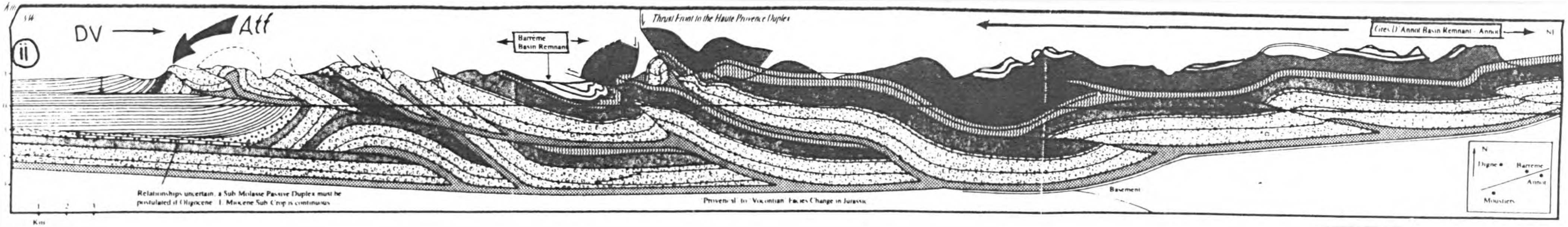


FIG 1.3 Geological map and structural cross-sections of the South-Western Alps and Rhone Graben, S.E France (after Debelmas 1980, and Lemoine 1980)



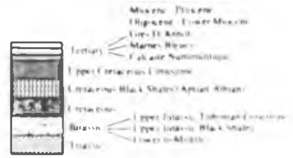


FIG 1.3 Section (ii) Allocthonous External Zone -> Autocthonous Tertiary basin (Digne Valensole basin (DV)) Atf: Alpine thrust front (from Graham 1985).

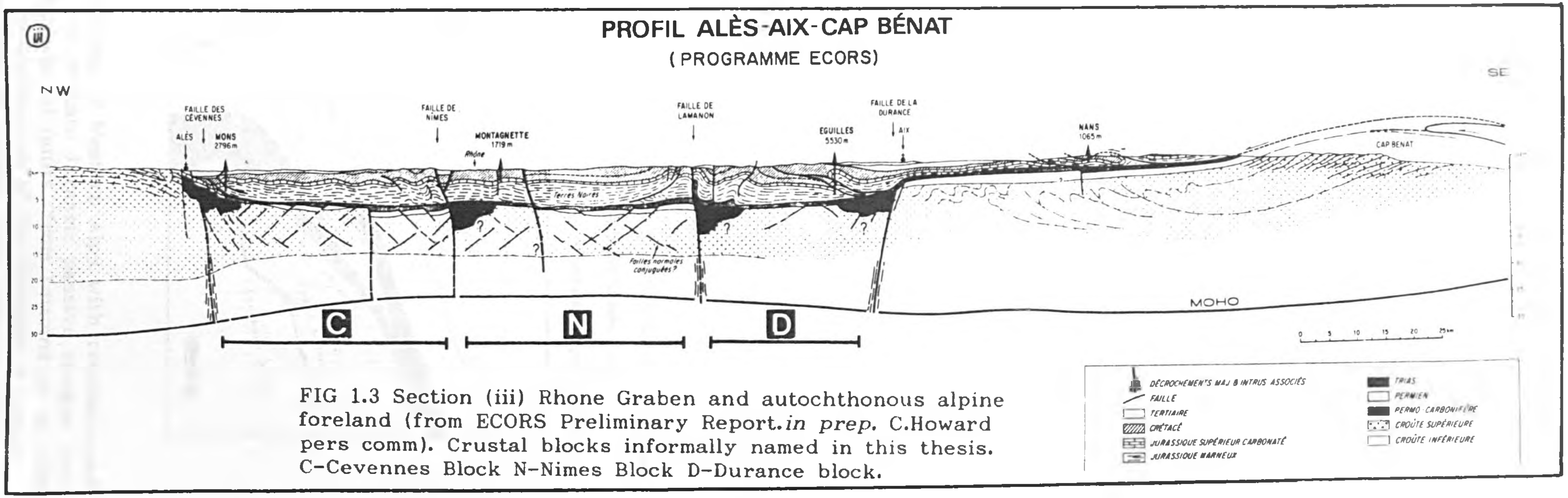
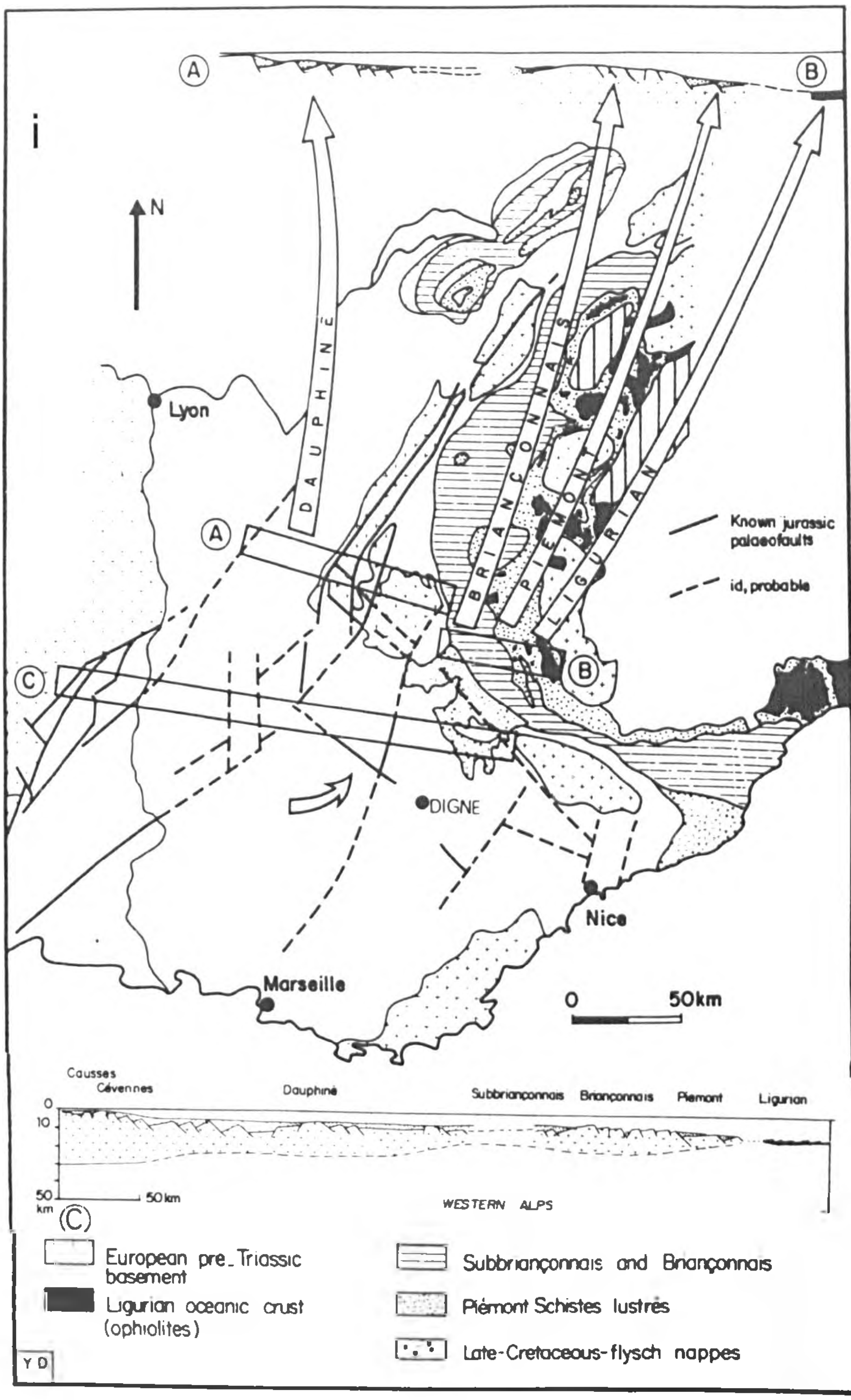


FIG 1.3 Section (iii) Rhone Graben and autochthonous alpine foreland (from ECORS Preliminary Report.*in prep.* C.Howard pers comm). Crustal blocks informally named in this thesis. C-Cevennes Block N-Nimes Block D-Durance block.



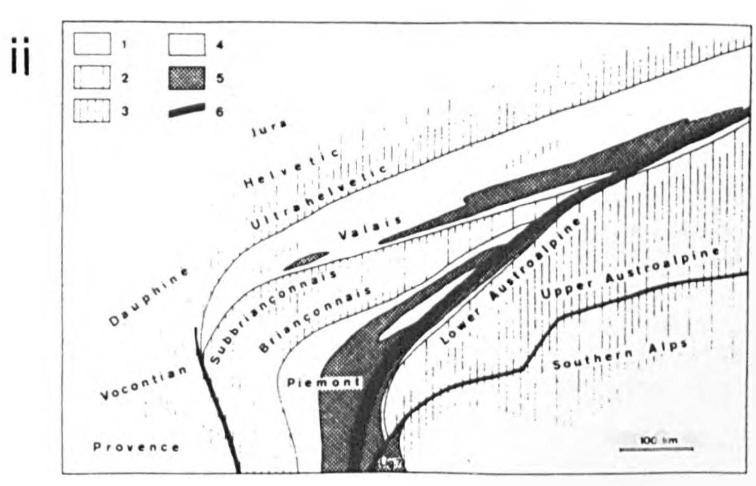


FIG 1.4 (i) Sketch map of Western Alps with reconstruced cross-sections (A-C) of Late Jurassic passive margin. Note the major Jurassic extensional faults (from Lemoine *et al.* 1986). (ii) palinspastic reconstruction of Mid-Jurassic - Lower Cretaceous alpine passive margin (from Lemoine 1980)

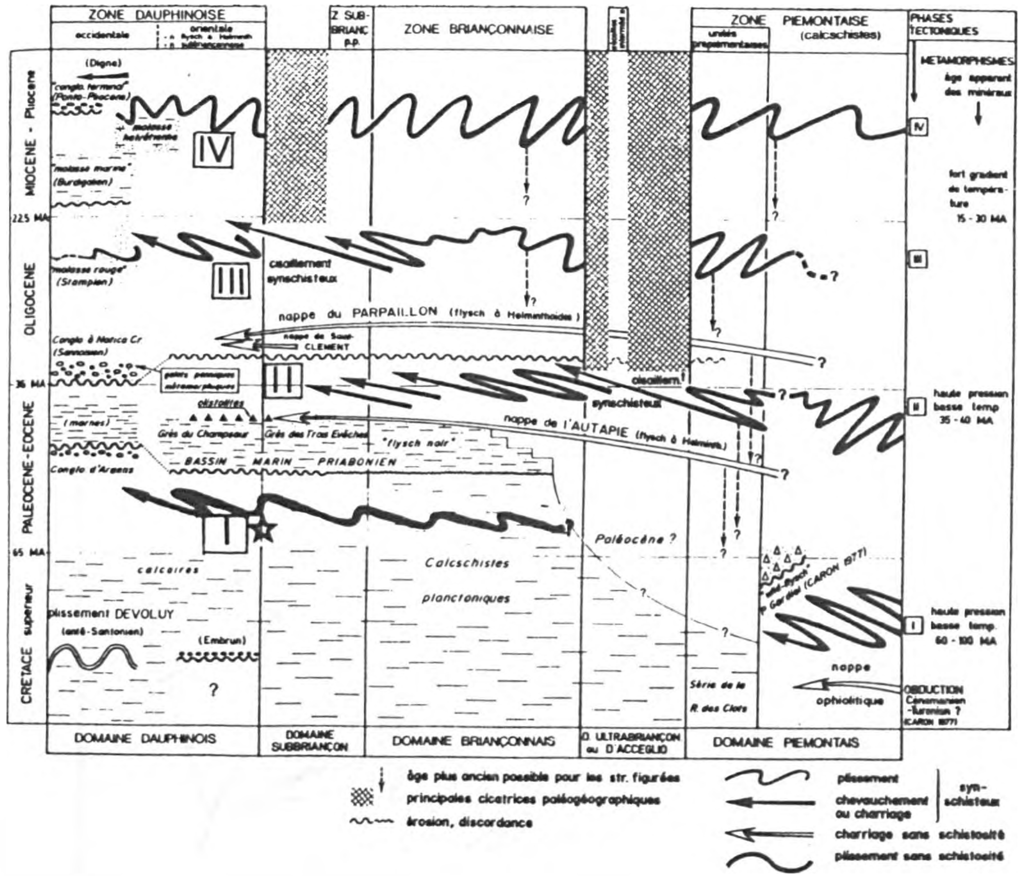


FIG 1.5 Principal tectonic phases recognised in the S.W Alpine thrust-fold belt (after Lemoine 1980 (B)) *- after Graham 1985).

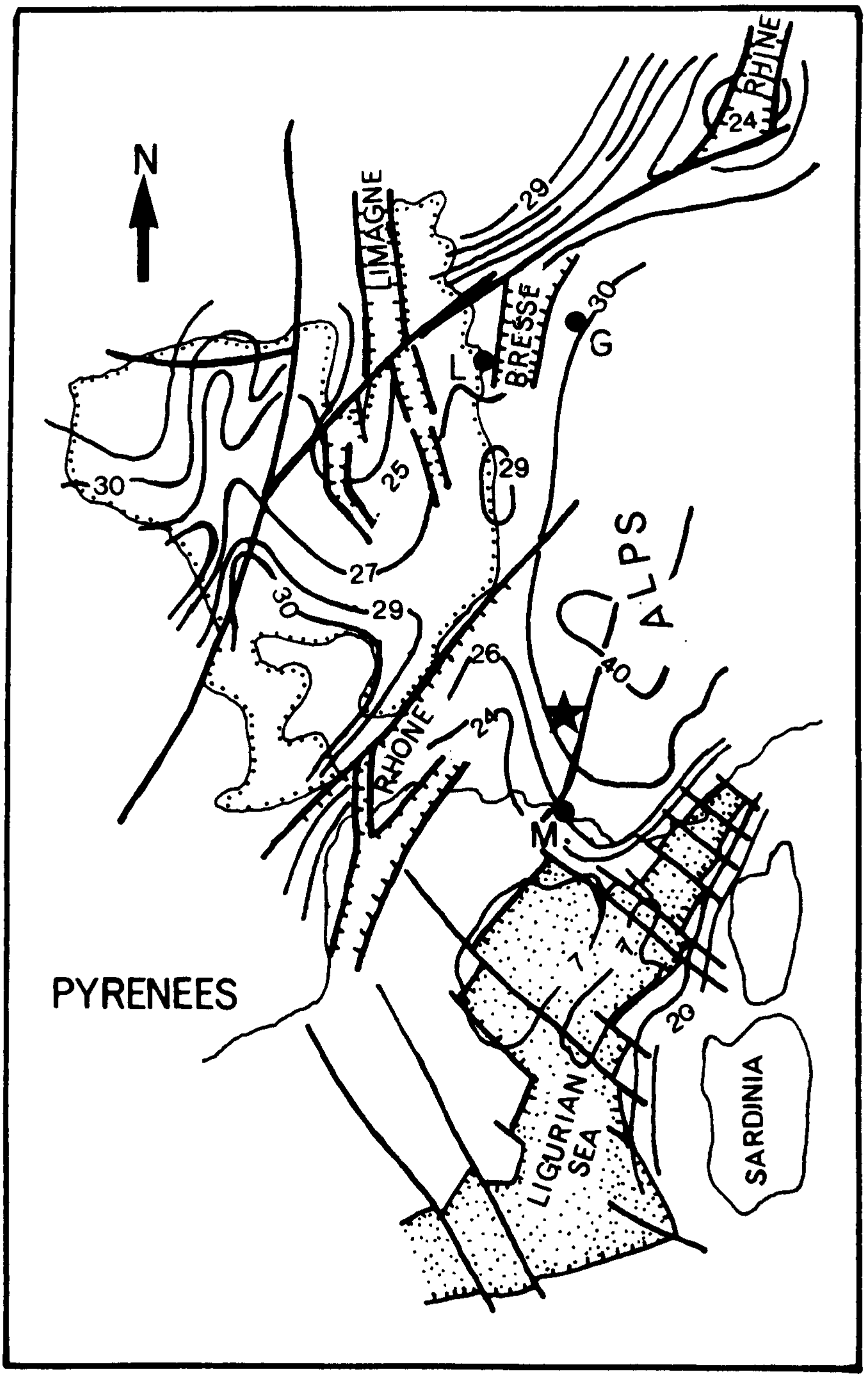
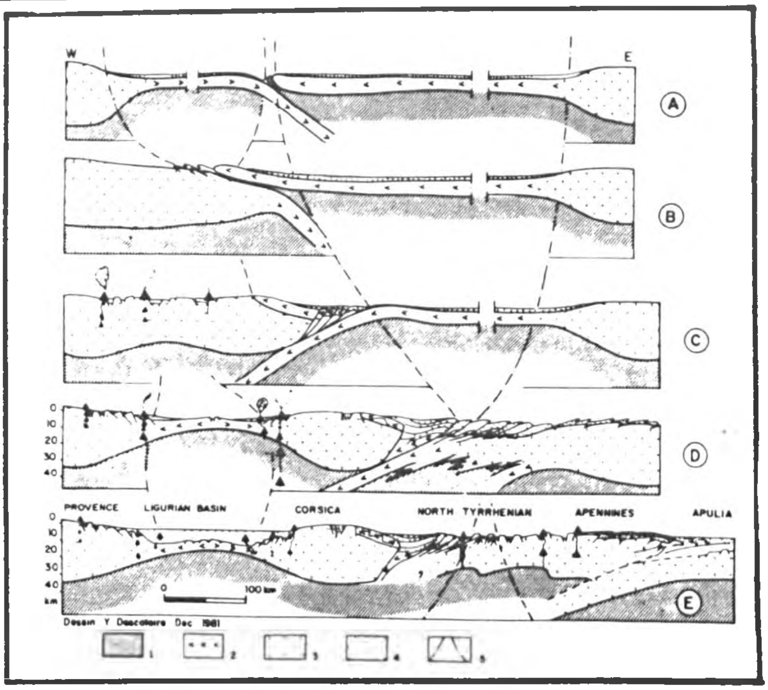


FIG 1.6 Schematic map of the Western European Rift system, showing the Rhone, Bresse, and Rhine grabens, superimposed on mantle-crust isobaths (after Lorenz *et al.* 1980, and Burrus 1984, Rehault *et al.* 1984). Maximum extension in Western Mediterranean Rift System occured in Ligurian Sea where oceanic crust developed. Note the attenuated crustal thicknesses in the Rhone Graben, relative to the thickened crust beneath the Alpine thrust belt. Position of study area marked by *.



(A) Upper Cretaceous: eastward European plate subduction; (B) Upper Eocene: Alpine collision and Corsican obduction; (C) Middle Oligocene: after the subduction "flip" (westward to northwestward Apulian subduction) occurred, resulting in rifting in the Western Mediterranean area, and calc-alkaline volcanic activity in western Sardinia and Provence; (D) Upper Burdigalian: after drifting in the Ligurian marginal basin, collision occurred between the Corsican and Apulian margins; (E) Middle Pliocene: subsidence continued in the Western Mediterranean: westward African (Apulian) subduction beneath the Calabro–Sicilian arc, and the Tyrrhenian Sea rifting and opening. *1* = Upper mantle; *2* = oceanic crust; *3* = continental crust; *4* = sediments; *5* = oceanic basin boundaries.

FIG 1.7 Geodynamic evolution of the Western Mediterranean along an E-W line crossing Provence-Languedoc, Corsica, and the Appenines (from Rehault *et al.* 1984).

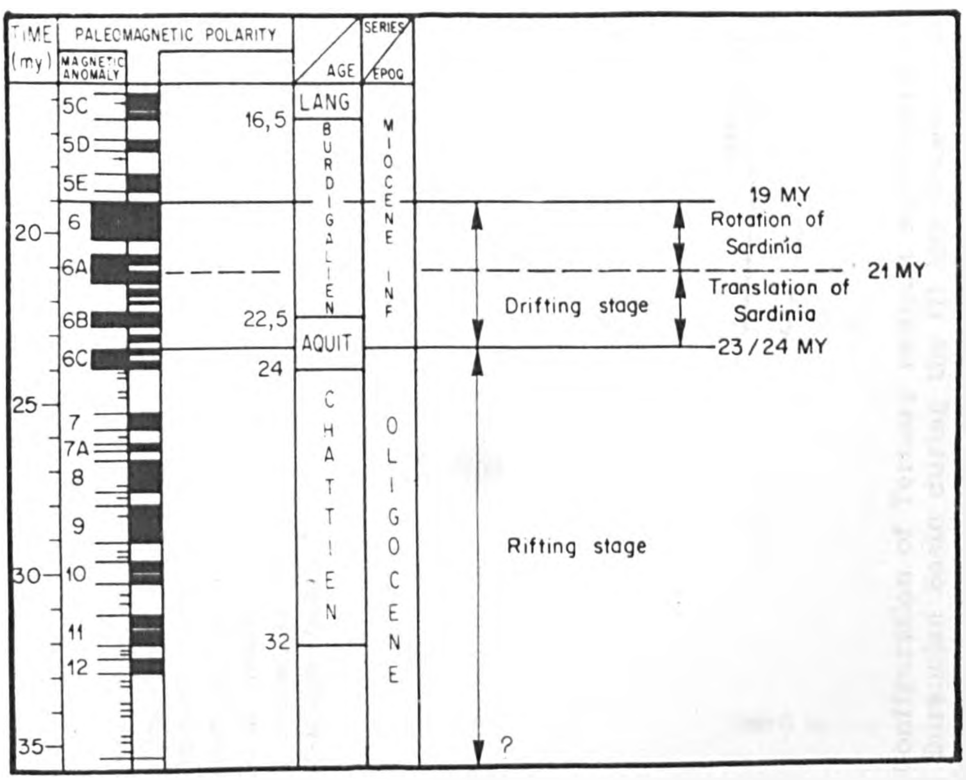


FIG 1.8 Magnetostratigraphy and dating of the opening of the Western Mediterranean Rift System (from Burrus 1984).

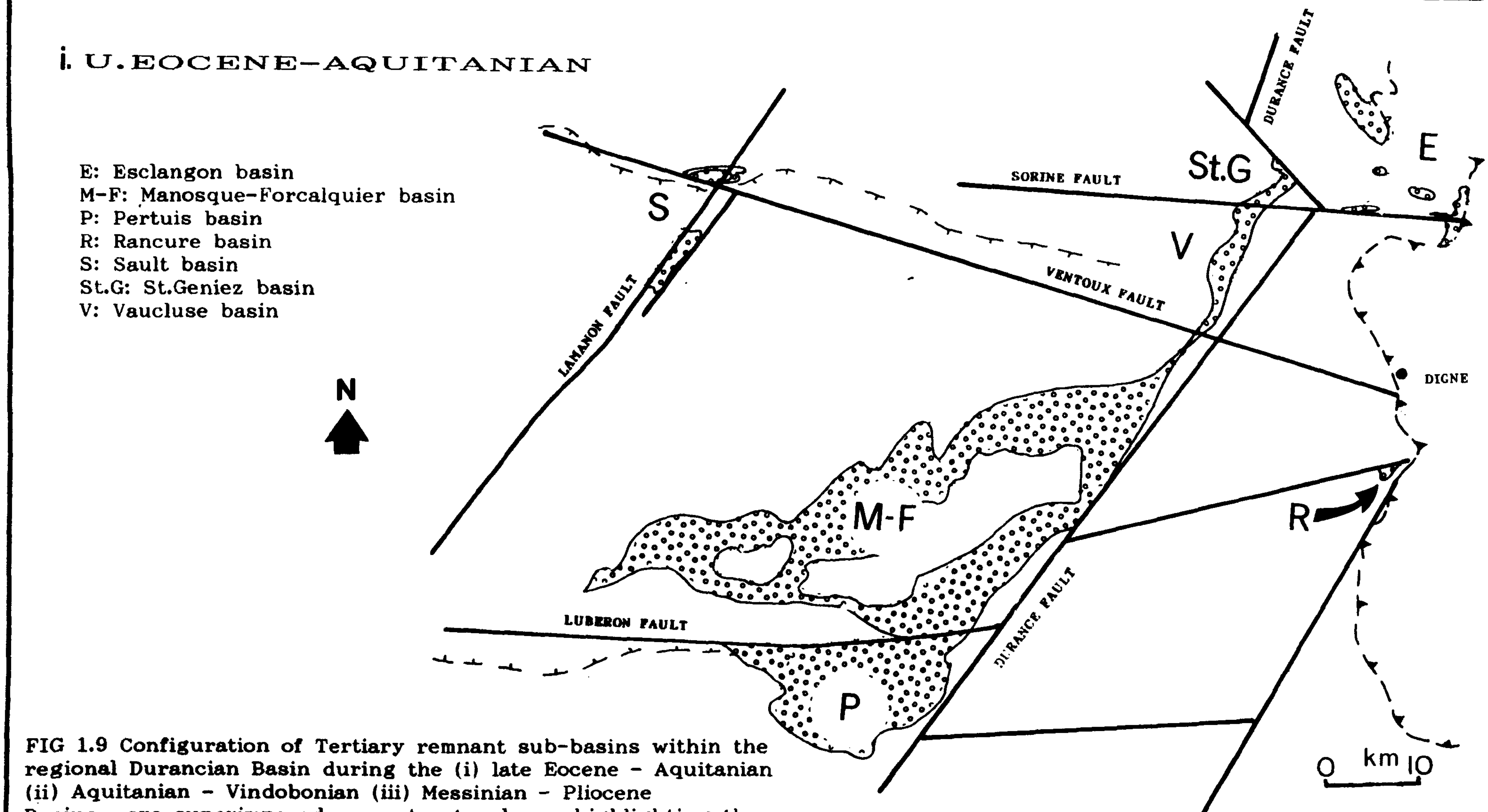
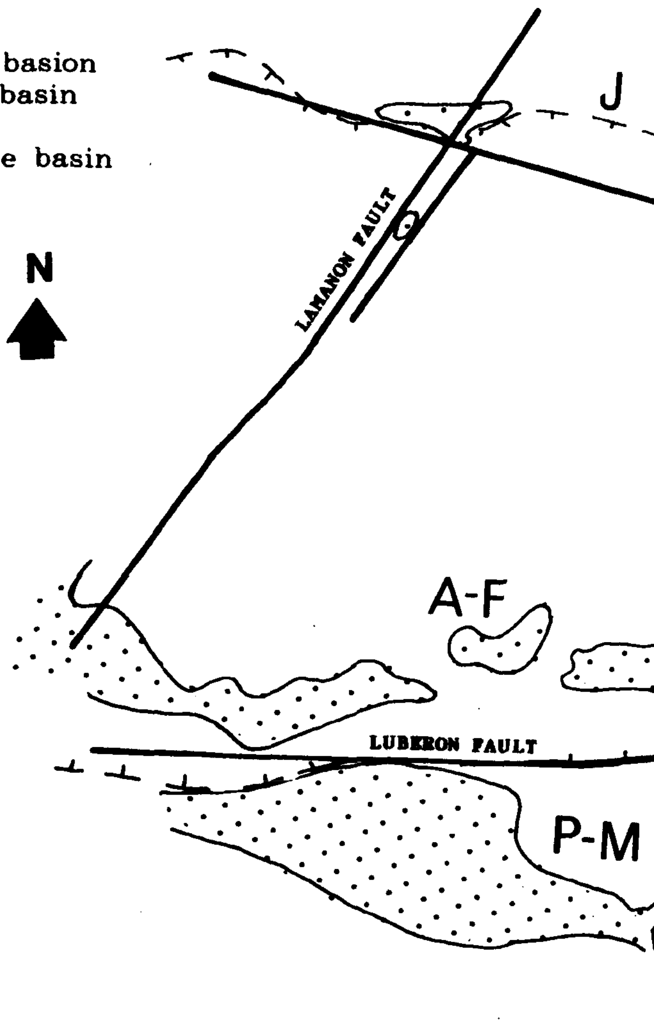


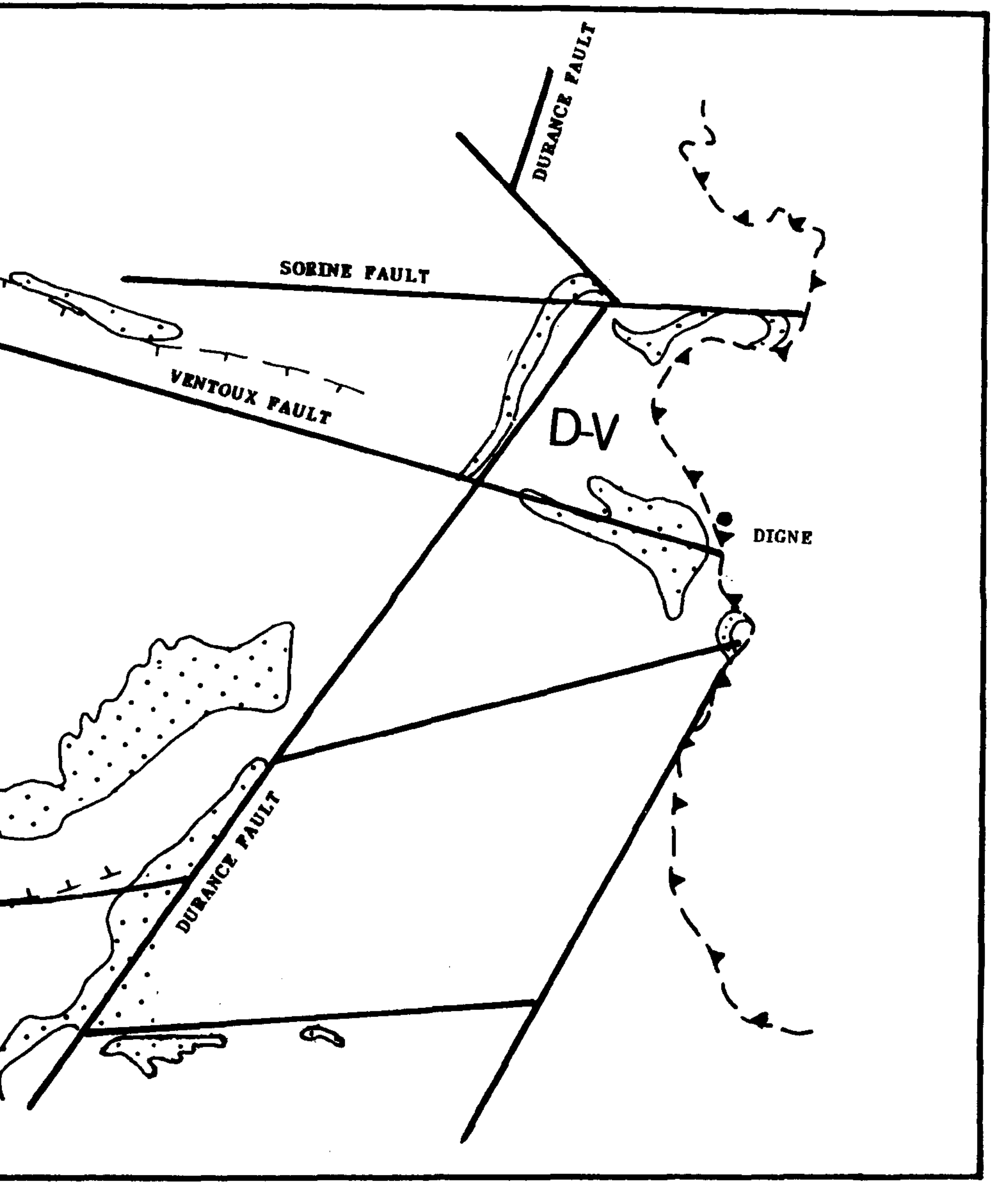
FIG 1.9 Configuration of Tertiary remnant sub-basins within the regional Durancian Basin during the (i) late Eocene - Aquitanian (ii) Aquitanian - Vindobonian (iii) Messinian - Pliocene
Basins are superimposed on a structural map highlighting the position of the inherited high-angle fault framework.

FIG 1.9 (ii)

BURDIGALIAN–VINDCBONIAN

A–F: Apt-Forcalquier basion
D–V: Digne-Valensole basin
J: Jabron basin
P–M: Pertuis-Manosque basin



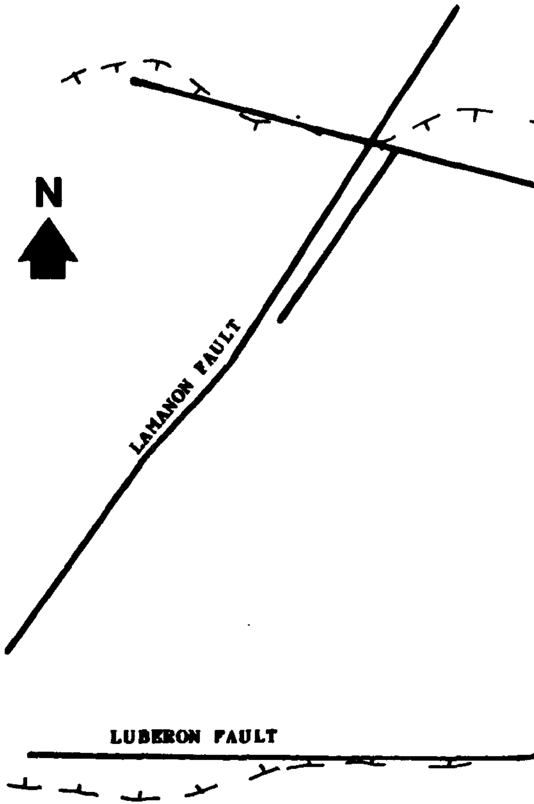


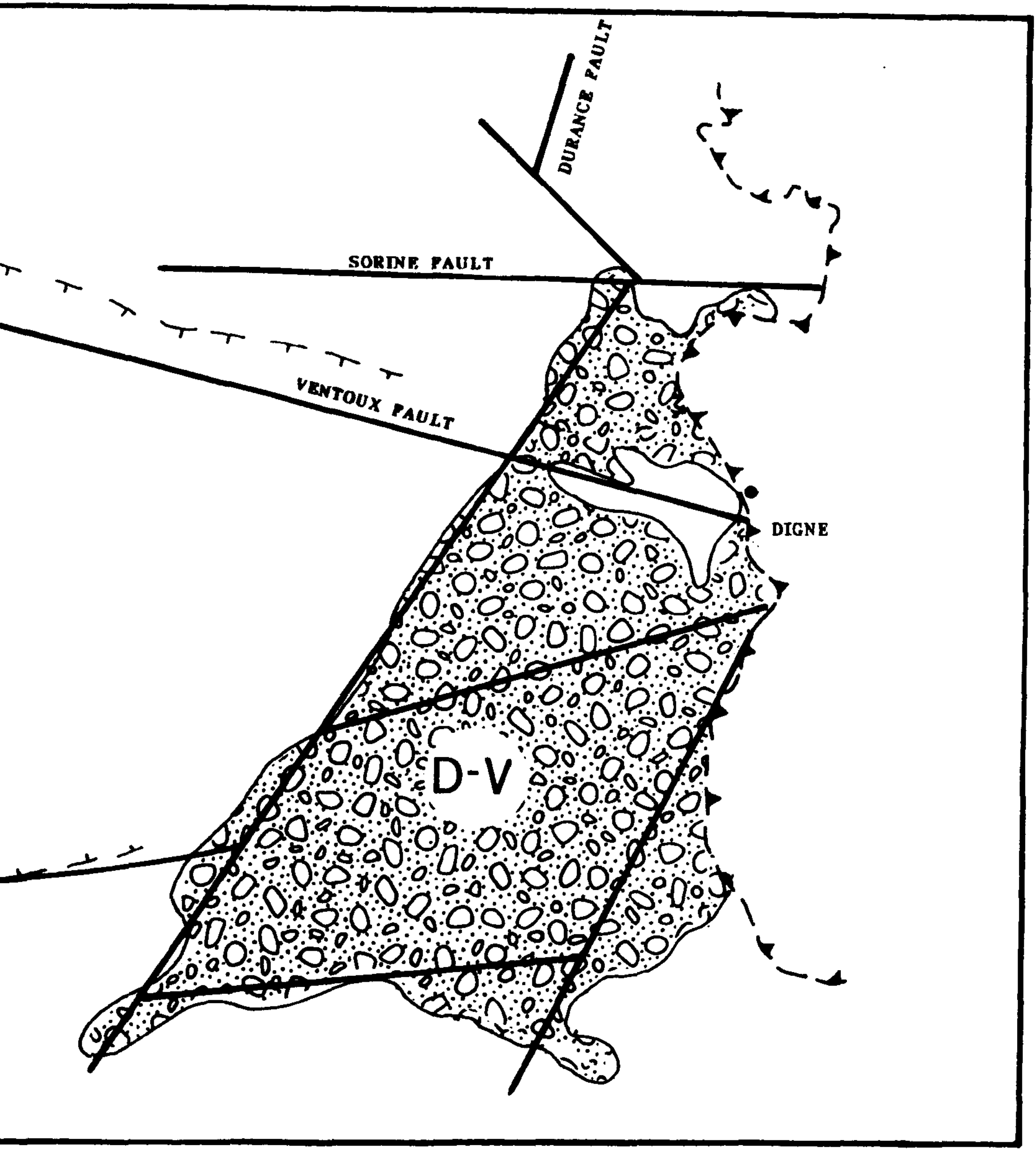
DURANCE FAULT
SORINE FAULT
VENTOUX FAULT
D-V
DIGNE
DURANCE FAULT

FIG 1.9 (iii)

MESSINIAN–PLIOCENE

D–V: Digne–Valensole.

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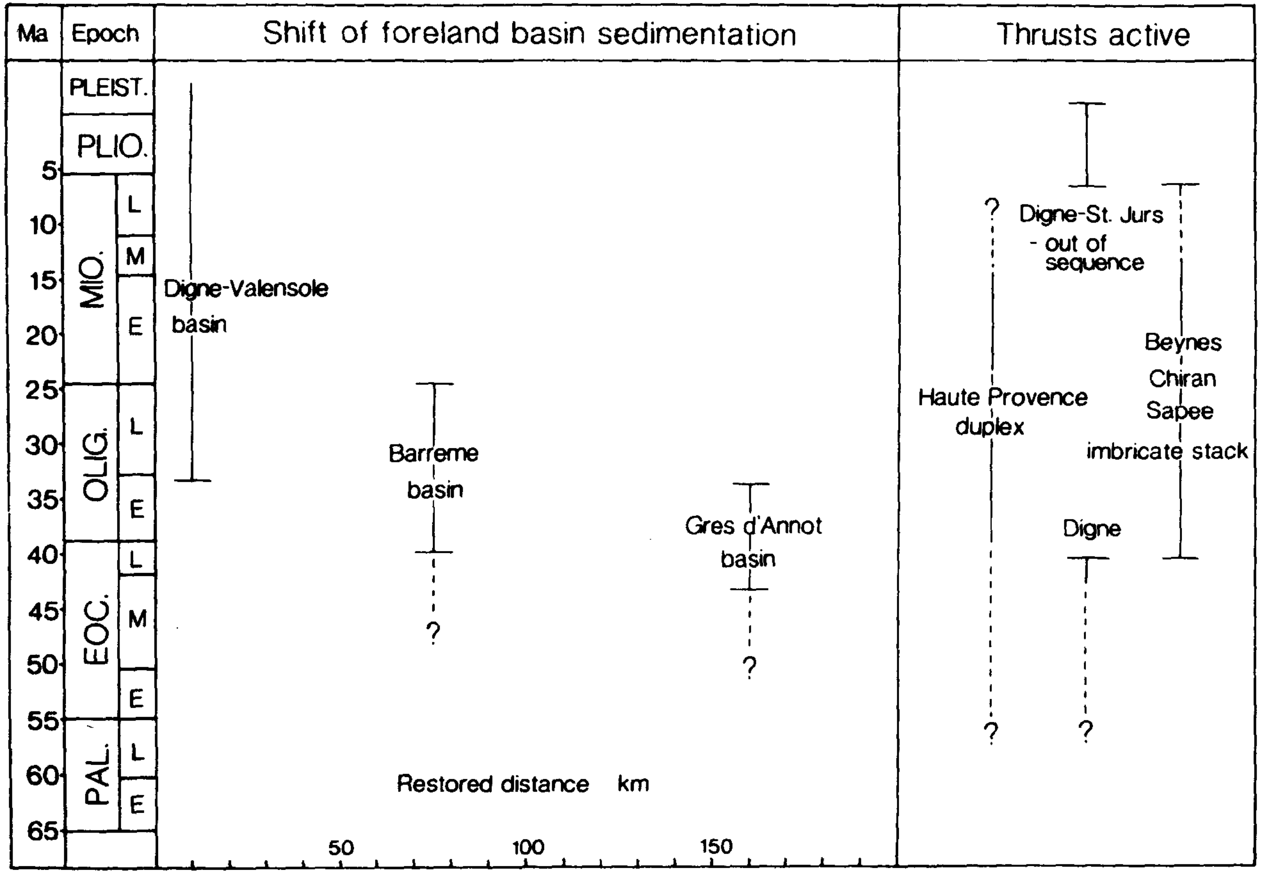
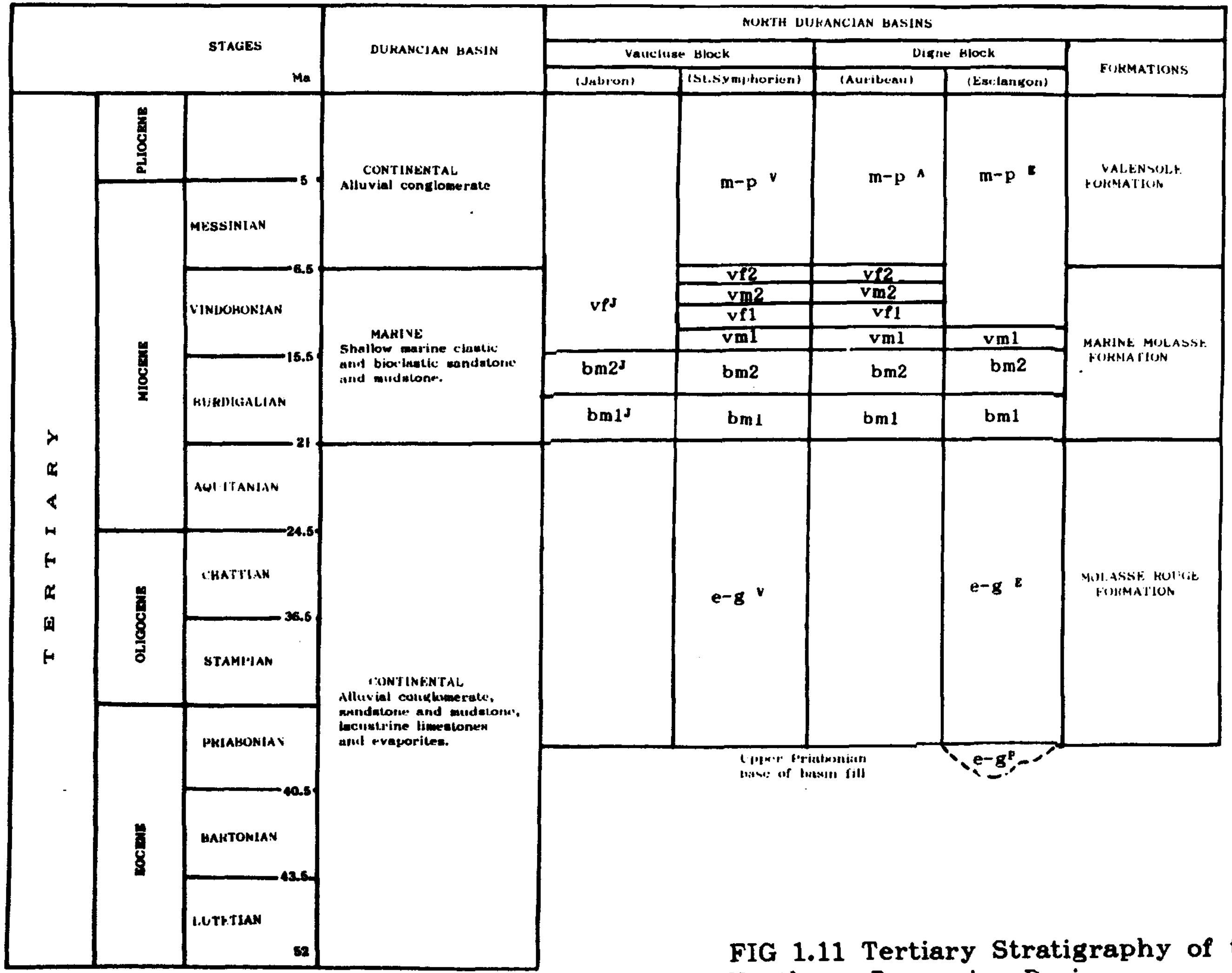


FIG 1.10 Stratigraphical relationships of Tertiary basin deposits of the S.W Alpine Foreland Basin (after Elliott *et al.* 1985)



m-p V - Valensole Conglomerate
m-p A - Auribeau Conglomerate
m-p E - Esclangon Conglomerate.

vf2 - Upper continental Vindobonian
vm2 - Upper marine Vindobonian
vf1 - Lower continental Vindobonian
vm1 - Lower marine Vindobonian
bm2 - Upper marine Burdigalian
bm1 - Lower marine Burdigalian

e-g V - Vancon System.
e-g E - Esclangon System
e-gP - Palaeo-valleys

FIG 1.11 Tertiary Stratigraphy of the Durancian, and specifically Northern Durancian Basins.

SERIES		STAGES
MIOCENE	UPPER	MESSINIAN
		5.2
		TORTONIAN
		6.3
	MIDDLE	SERRAVALLIAN
		10.2
		LANGHIAN
		15.2
	LOWER	BURDIGALIAN
		20
		AQUITANIAN
		25.2
OLIGOCENE	UPPER	CHATTIAN
		30
	LOWER	RUPELIAN
		36
EOCENE	UPPER	PRIABONIAN
		39.4
	MIDDLE	BARTONIAN
		42
		LUTETIAN
		49
	LOWER	YPRESIAN
		54
PALEOCENE	UPPER	THANETIAN
		60.2
	LOWER	DANIAN
		66.5

Fig 1.12 Tertiary chronostratigraphy (after Haq *et al.* 1987)

CHAPTER 2.

THE INHERITED STRUCTURAL FRAMEWORK OF SOUTH-EAST FRANCE AND ITS REACTIVATION DURING THE TERTIARY.

2.1 Introduction.

The repeated tectonic reactivation of major fault zones in the continental crust ("resurgent tectonics") is now a widely supported concept (Watterson 1975; White *et al.,* 1986, and Etheridge 1986). Reactivation of these faults is likely in both a high-level brittle regime where the faults have lower cohesive strengths and sliding frictions than the surrounding rock, and, in a deeper, ductile-regime where flow strength is lower because of their finer grain sizes or distinctive mineralogy (Watterson 1975, Etheridge 1986)

It has been widely suggested that the Tertiary evolution of S.E France was controlled, to variable degrees, by the reactivation of an inherited structural framework of high-angle crustal faults.

This chapter addresses the problem by:

(i) reviewing the evidence for an inherited (pre-Priabonian) structural framework (section 2.2).

(ii) reviewing the evidence for the Tertiary (post-Priabonaian) reactivation of these inherited structures (section 2.3).

(iii) integrating the reviewed data to generate a map of the inherited faults and the fault blocks *1 which they define within the Durancian basin (section 2.4).

Footnote *1

'Fault blocks' (after Sylvester *et al.,* 1976) are distinguished by (i) marked differences in lithostratigraphy and stratigraphic thicknesses of their cover (i) having margins characterised by narrow zones of compressional, extensional or strike-slip deformation with the fault blocks themselves being relatively undeformed areas.

A tectono-sedimentary model of the Tertiary evolution of the Durancian basin, which integrates this data with structural and sedimentological data collected during fieldwork, is presented in Chapter 6.

2.2 Inherited Structural Framework of S.E France

2.2.1 Introduction

Numerous studies on the late Palaeozoic, and particularly Mesozoic sequences of S.E France have revealed that the region is transected by a series of high-angle crustal faults. The consensus of opinion, and the contention in this thesis, is that the principal inherited faults were a series of NE-SW trending, faults, namely the Cevennes, Nimes, Lamanon and Durance faults (Fig. 1.2). These faults can be demonstrated to have had a multiphase history of activity prior to the Priabonian (L. Tertiary) with three major phases of utilisation recognised (Fig. 2.1), namely:

i) a Hercynian (Carboniferous) wrench phase.
ii) a Mesozoic extensional phase.
iii) a late Cretaceous - lower Tertiary compressional phase.

2.2.2 Hercynian phase - basement shear faults

The major crustal faults of S.E France are interpreted to have been initiated as wrench faults during the Carboniferous and to have formed a part of the Hercynian deformation system within the European plate (Fig. 2.2 (i)) (Lorenz *et al.,* 1980)

Two principal sets of Hercynian shear faults have been recognised by Vialon (1974) and Boudon *et al.,* (1976) namely a Variscan Set, striking N50°E, and an Amorican Set, striking N140°E. These can be demonstrated to have been active and controlling Carboniferous-Permian sedimentary basins in the Massif Central (where pre-Mesozoic basement is exposed) (Fig. 2.2 (ii)), and it is presumed that similar fault bound basins are buried beneath the Mesozoic cover of S.E France.

2.2.3 Mesozoic phase - extensional faults.

During the Mesozoic, rifting of the European continental margin associated with the opening of the Central and Northern Atlantic (Fig. 2.3) was accomodated in the region by movement on three principal fault sets, namely i) NE-SW / NNE-SSW ii) NW-SE i) E-W / WNW-ESE, trending sets (Lemoine *et al.*, 1986)

The similarity in orientation of these faults and the earlier Hercynian faults clearly suggests that the latter were reactivated in an extensional regime during the Mesozoic.

Isopach maps of the total (preserved) Mesozoic thickness superimposed onto a structural map of the External Zone and Rhone Valley region of S.E of France (Fig. 2.4 after Menard 1980) show the important role played by NE-SW and E-W crustal faults in accomodating up to 11 kms (compacted) thickness of Mesozoic sediment. The faults defined a series of major depocentres, separated by structural highs above which reduced sequences were developed. Of particular interest to this study is the Mesozoic Durancian 'High' which separated the Provence and Vocontian Basins, and above which the Tertiary Durancian basin developed (Fig. 2.4).

Two principal phases of rifting accompanied by active faulting have been recognised within the Mesozoic (Lemoine *et al.*, 1986), namely:

(i) Late Triassic - Liassic rift phase.

This, the major rift phase, was initiated in the Late Triassic / Liassic with the opening of the Central Atlantic and Ligurian Sea (Fig. 2.3). The eastern margin of the Massif Central defined the hinge line of the extended European passive continental-margin, and to the east of which developed a horst and graben system. This is clearly demonstrated by isopach maps for the total Jurassic (Fig. 2.5 after Hossack *pers. comm* from data of Debrand-Passard 1984) which demonstrates that elongate regional grabens were developed defined by NE-SW trending faults (Cevennes, Nimes, Durance), and dissected by E-W, and NW-SE trending 'transfer faults' (Ventoux, and Luberon). Regional isopach studies are supported by detailed localised studies of the geometry of the faults and the basin fill at the margins of these fault-bound basins (See Fig. 2.5 (ii) after Arnaud *et al.*, 1978, and Lemoine *et al.*, 1986).

(ii) Lower Cretaceous rift phase.

A second phase of Mesozoic rifting which re-utilised the inherited fault framework was initiated in the Lower Cretaceous in response to the opening of the North Atlantic and Bay of Biscay (Lemoine *et al.,* 1986). Isopach maps for the total Cretaceous (Fig. 2.6 after Hossack *pers. comm.* from data of Debrand-Passard *et al.,* 1980) show that in response to this stress regime the E- W / WNE-ESE orientated extensional faults, particularly the Ventoux and Luberon faults, were preferentially reactivated and defined an E-W orientated regional depocentre (Vocontian basin) (Fig. 2.6).

Detailed analysis of the Cretaceous lithostratigraphy of this basin (Fig. 2.6 (ii) from Porthault 1982, and Cotillon *et al.,* 1984) supports this interpretation and confirms that the E-W trending Ventoux and Luberon faults were active. Detailed facies analysis of basin fill (Remane 1970, Ferry *et al.,* 1979, and Cotillon *et al.,* (1984) also show that the facies belts and sediment transport paths within the basin were orientated parallel to the E-W trending faults of the graben margin.

2.2.4 Late Cretaceous-lower Eocene phase - thrust-fold belt.

During the late Cretaceous - lower Eocene period the South of France lay in the northern foreland to the Pyrenean orogenic belt and was subjected to northward directed compressional deformation (Fig. 2.7), termed the Pyrenean - Provencal deformation phase. As a result, East - West orientated high-angle basement faults were positively inverted. East-West striking, and north and southward verging compressional structures developed in the Mesozoic cover sequences directly above these major faults, with the regions lying beween the faults remaining essentially undeformed. In the Durancian basin study area this deformation took the form of low amplitude folds which developed above the E-W trending Ventoux, Mirabeau, Rancure and Luberon faults (see Fig. 1.2 (i-iii).

2.3 Tertiary Reactivation of Inherited Faults.

To date, work on this problem has taken two approaches, namely:

(i) compressional reactivation in the alpine thrust belt.

(ii) extensional reactivation in the region of the Rhone Graben.

2.3.1 Compressional reactivation.

The western alpine belt has provided a number of detailed examples of the compressional reactivation (inversion) of crustal faults during the Tertiary (Davies 1982, Barfety & Gidon 1983, Tricart & Lemoine 1982, see Gilchrist *et al., 1988: in prep.* for review).

These studies emphasise the importance of the inherited, fault framework in defining the geometry of the alpine thrust sheets. As reviewed by Gillchrist *et al.,* (*op. cit.*) this control is twofold. Firstly, throughout much of the allocthonous external zone the alpine thrusts affected only the Mesozoic and Tertiary "cover" sequences (detached from crystalline basement along Triassic evaporites) (Beach 1981; see also Fig. 1.3 sections i-ii) with the high-angle faults primarily acting as stress risers causing the localisation and positioning of thrust sheet ramps. This can be demonstrated for the frontal alpine thrust, the Digne thrust sheet (See Fig. 2.8 after Siddans 1979). Secondly, in the external massifs control is demonstrable at a lower level as the high level thrusts are joined, folded or breached by lower level thrusts which have reactivated earlier high-angle crustal (crystalline basement) faults.

2.3.2 Extensional reactivation.

The concept of inherited faults being reactivated during the Tertiary was proposed by Goguel (1944, 1948) from his study of the Rhone graben. An isopach map of the Tertiary of S.E France (data from Debrand-Passard *et al.,* 1984) superimposed onto a structural map of the region (Fig. 2.10) supports this interpretation and demonstrates that the Tertiary depocentres were primarily controlled by NE-SW trending faults (a similar situation to the Jurassic).

Notable differences to the Mesozoic subsidence pattern (Fig. 2.4 - 2.6) are the absence of any significant Tertiary sequences in the Vocontian region which had been a regional depocentre in the Mesozoic, and also the development of thick Tertiary sequences across the 'Durancian Block' which had been a persistent Mesozic high.

In the study area of Provence (Durancian Basin), work on the extensional reactivation of inherited faults has concentrated on the Manosque-Forcalquier basin (one of the Duracian basins: see Fig. 1.9) which was interpreted to have developed adjacent to a reactivated Mesozoic fault, the Durance fault (Goguel 1959, and Montenant 1968). Subsequent work (Cavelier *et al.*, 1984) has provided detailed descriptions of the lithofacies and lateral thickness variations within the basin fill to support this interpretation (See Fig. 3.61).

An appreciation of the regional importance of inherited 'Mesozoic' faults during the Tertiary has also been shown by Triat & Truc (1983) who produced a map of the NE-SW high-angle faults of S.E France (Fig. 2.11). They used a case study basin (Carpentras basin) to demonstrate the active extensional role of these faults during the Tertiary.

However the most direct evidence of the Tertiary reactivation of crustal faults comes from the Durancian Basin itself, where a seismic-reflection profile, and bore-holes (Dubois & Curnelle 1978) demonstrate that small Neogene basins were developed above the site of high-angle faults, the Mirabeau, and Rancure fault zones, which offset the Hercynian basement (Figs. 2.12 - 2.13)

2.4 Inherited faults of the Durancian Basin.

The inherited high-angle faults of the Durancian Basin are detailed in Fig.ure 2.12. The faults have two principal orientations, namely NE-SW, and E-W/ENE-WSW, and sub-divide the basin into a series of discreet 'fault blocks'.

The faults of the NE-SW set are the most laterally continous and include the major Durance and Lanamon faults which have been detailed in Fig. 1.3 (iii). The Durance fault lies buried beneath Tertiary sequences of the Durancian basin (Fig. 1.9) but its position is inferred from a series of en-echelon folds developed along its length. In the west of the basin a NE-SW trending fault, the Valensole fault is tentatively suggested after Dubois & Curnelle (1978). Its presence is supported by the fact that if extended north-eastwards into the allocthonous external alps it defines the lateral ramp of the Digne thrust sheet (Fig. 2.9). Furthermore, thickness variations in the lower Jurassic sequences of the thrust sheet show that it is an inverted Mesozoic half-graben and that the Valensole fault acted as an extensional fault defining its southern margin prior to acting as an alpine (Tertiary) lateral ramp (Fig. 2.9).

The principal faults of the E-W set are the major Ventoux, and Luberon faults which have already been shown to have played a major role in the Cretaceous history of the area (section 2.2.3).

The present expression of these faults is as E-W trending anticlines, with thrust breached cores, which are interpreted to have developed through the compressional inversion of the faults during both the Pyrenean, and subsequently Alpine, deformation phases. The trace of the Luberon thrust fold is noted to curve into the Durance fault, and to intersect with it at the point of development of a buried Triassic diapir. The position of the Rancure fault is derived from its map expression, and the development of a small, Oligocene filled graben on its southern margin (Graziansky *et al.*, 1982). A little further to the north borehole and seismic date shows the development of a zone of faults at Mirabeau which appear to be an eastward extension of the Ventoux fault.

The Sorine fault (described in detail in section 6.6; see also Fig. 6.6) is considered to extend eastwards to the Digne thrust front, and to have defined the position of the lateral ramp of the Melan- Clamensane sheet thrust.

Summary.

There is widespread evidence for the repeated reactivation of Hercynian age, high-angle lithospheric faults under both extensional and compressional regimes during the Mesozoic and Tertiary periods. The potential of these faults having been active during the late Eocene-Pliocene evolution of the northern Durancian basins appears to be high, and as will be demonstrated in the succeeding chapters they were to play a critical role in controlling these sedimentary basins.

Period	Carboniferous	L.Triassic-Mid Jurassic	Low. Cretaceous Mid Cretaceous		L.Cretaceous-M.Eocene	L.Eocene-Oligocene	L.Oligocene-Aquitanian	L.Miocene-Pliocene	L.Pliocene-Quaternary
Deformation Phase	Hercynian	Tethyan rifting	N. Atlantic rifting.		Pyrenean - Provencal Phase	Western-Mediterranean Phase		Alpine Phase	------------------->
Deformation	W	E	E		C	E	C	C	C
AREA									
S.E France	Conjugate fault sets N50°E N140°E	NW-SE Fault sets Principal NE-SW NW-SE, E-W	N-S Fault sets Principal E-W NE-SW, NW-SE	Luberon[1]	003°	097°	043°	008°	-
				Vaucluse Plateau[1]	003°	114°	056°	-	-
				North Digne-Valensole Basin / Digne Thrust Sheet.[2]	000°	-	050°	350-020°	020-030°

1 - Bergerat 1987. 2 - Gigot *et al.* 1974.

C - principal compressive strain axis. E - principal extensional strain axis.

FIG 2.1 Deformation phases of S.E France.

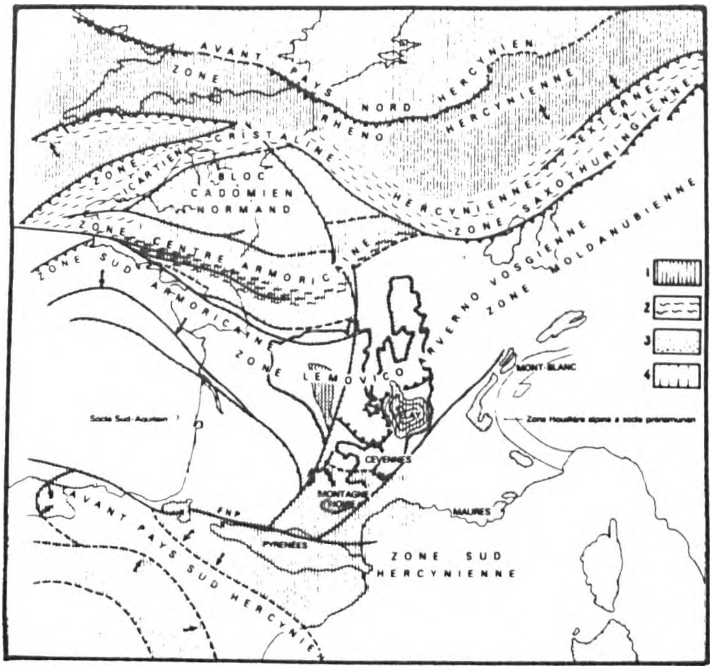


FIG. 2.2(i)Hercynian Structures of S.E France - Major structural zones of the Variscan chain. 1: Upper Carboniferous deformation zone 2: Dinantian deformation zone 3: Internal zone of Variscan orogeny 4: Unaffected pre-Variscan basement (from Lorenz *et al.* 1980)

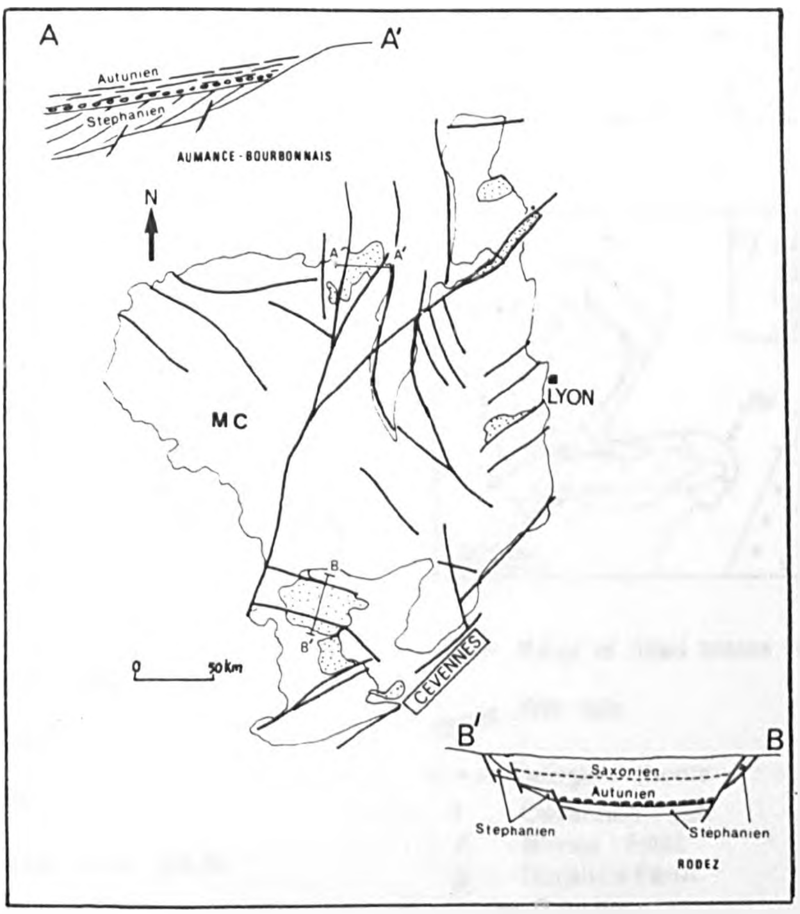


FIG 2.2(ii) Hercynian trans-tensional wrench faults active during the Upper Carboniferous-Permian in the Massif Central controlling sedimentary basin develop ment (after Lorenz *et al.* 1980).

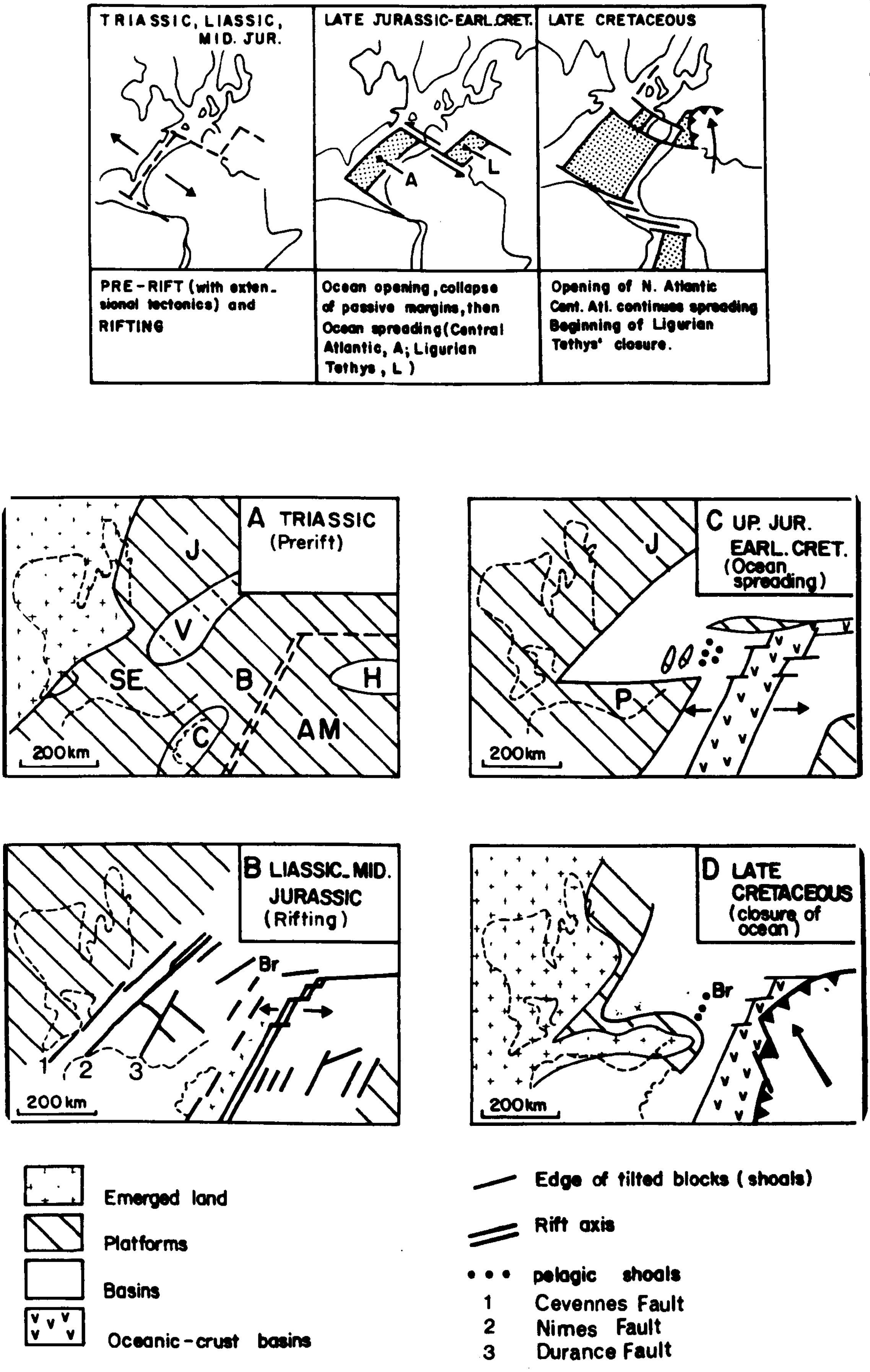


FIG 2.3 Main stages in (i) the developement of the Ligurian Tethys and Atlantic oceans, and, (ii) the palaeogeographic developement of the Ligurian ocean and adjacent margins (from Lemoine *et al.* 1986).

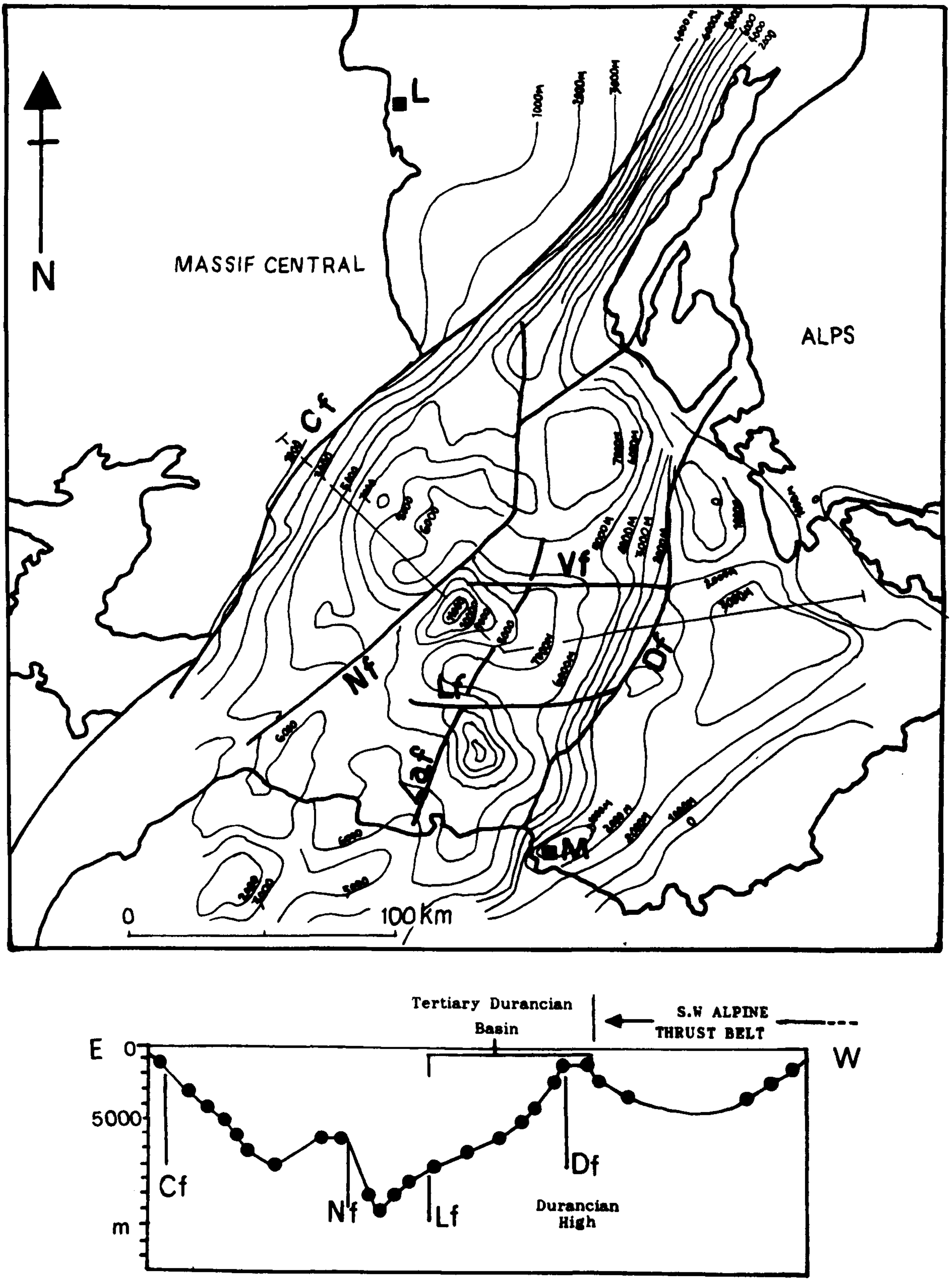


FIG 2.4 (i) Present day isopach map of the total preserved Mesozoic thickess superimposed on a structural map of S.E France (after J. Hossack *pers. comm* from data of Debrand-Passard *et al.* 1984). Isopachs in metres.

(ii) A cross-section (E-W) of (i), high-lights the active control of NE-SW lithospheric faults on Mesozoic basin developement. The position of the "Durancian High" which was a structural high throughout the Mesozoic is marked on the section, together with the site of developement of the Tertiary Durancian Basin.

Cf: Cevennes fault D: Durance fault Laf:Lamanon fault L: Luberon fault N: Nimes fault V:Ventoux fault.

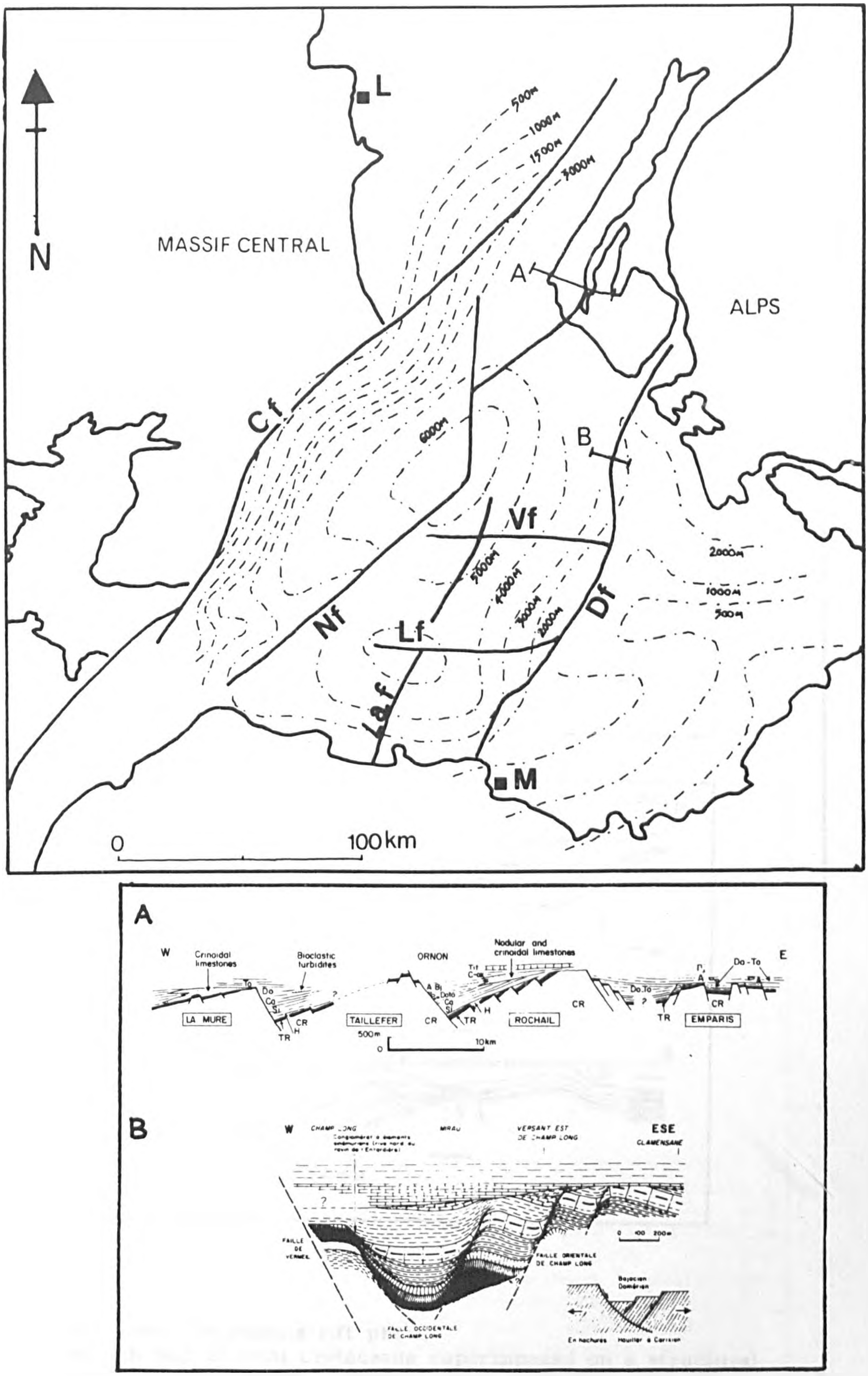


FIG 2.5 Late Triassic - Liassic rift phase.
(i) Isopach map of total Jurassic superimposed on a structural map of S.E France (after J.Hossack *pers. comm* from data of Debrand-Passard *et al.* 1984) Faults as in Fig 2.4 (ii) Reconstructed geometries of syn-rift graben-horsts in (A) the Pelvoux (from Lemoine *et al.* 1986) and (B) Alpes-de-Haute Provence (Clamensane) (from Arnaud *et al.* 1977). Positions of sections A & B is marked on (i).

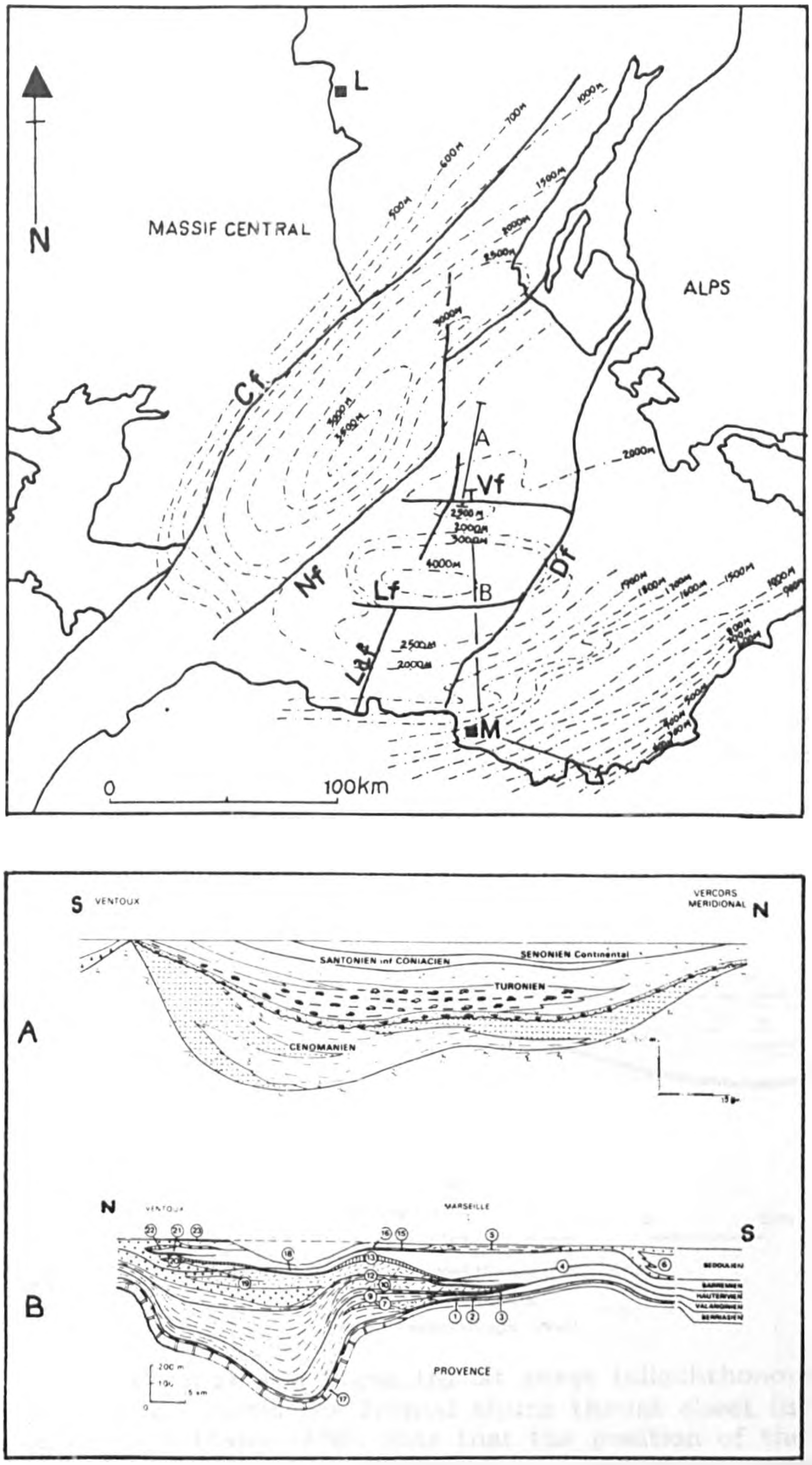


FIG 2.6 Lower Cretaceous rift phase.
(i) Isopach map of total Cretaceous superimposed on a structural map of S.E France (after J. Hossack *pers. comm* from data of Debrand-Passard *et al.* 1984). Note that the regional depo-centre (4,000m) is defined by E-W faults. Faults as in Fig. 2.4. (ii) Reconstructed geometries of major half-grabens defined by E-W trending faults within the Vocontian Basin.

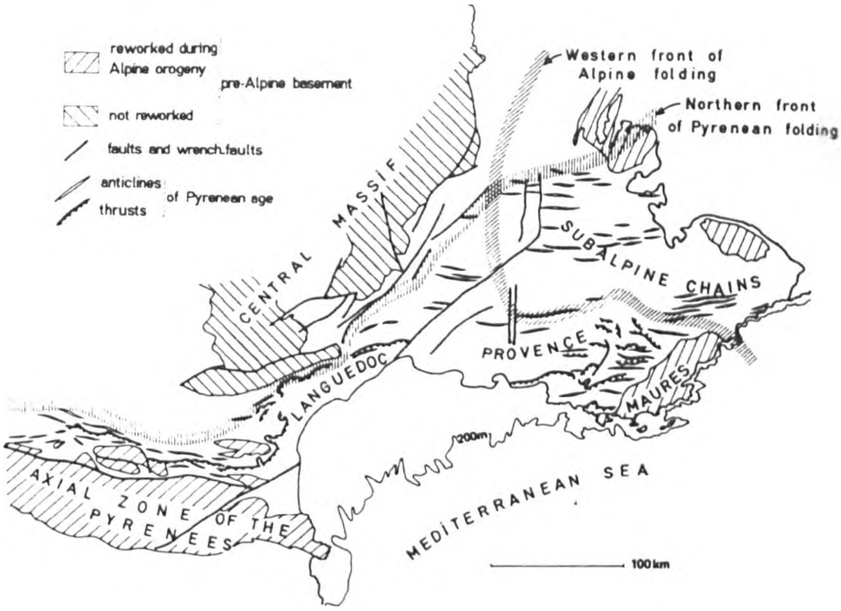


FIG 2.7 Schematic structural map showing the limits of the North Pyrenean and Western Alpine thrust-fold belts (from Lemoine 1980).

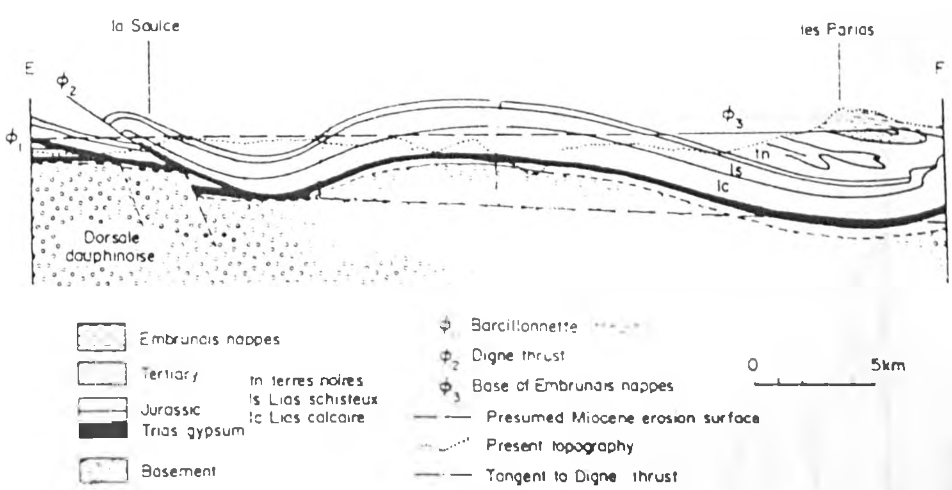
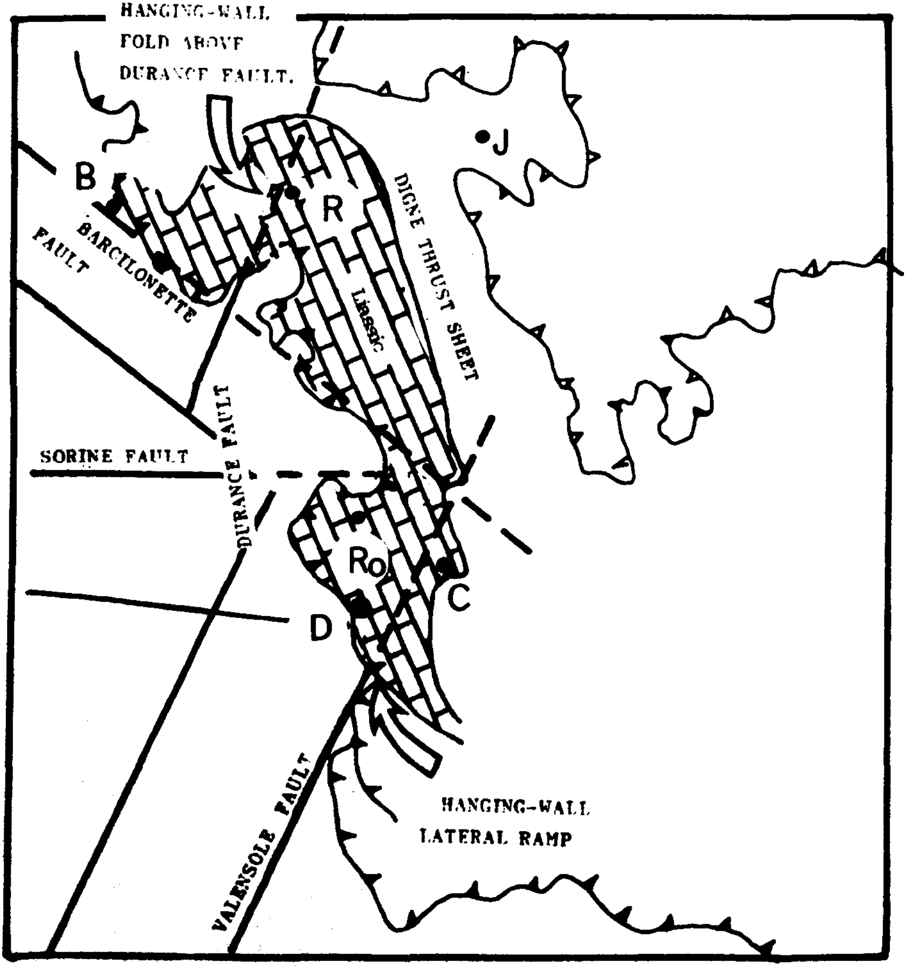


FIG 2.8 Section through the Digne thrust sheet (allochthonous external alps) which forms the frontal alpine thrust sheet in the study area (from Siddans 1979). Note that the position of the thrust sheet's frontal ramp is controlled by a Hercynian basement fault zone (Barcillonettte fault zone of Fig. 2.12) - this offsets the Hercynian basement creating a basement horst (Dorsale Dauphinois) which acts as a stress riser causing the ramping of the thrust sheet. Also note that the "Dorsale Dauphinois" is another term for the "Durancian High" which was noted to be a structurally high fault block throughout the Mesozoic (see Fig 2.4 (ii)). Siddans section is oversimplified in showing the thrust sheet as having a "layer-cake" stratigraphy, in fact it takes the form of a half-graben fill and the thrust sheet may be interpreted as an inverted half-graben (see Fig 2.9).



B:Barcillonette
C:la Cine
J:St. Julien
R:Remollon
Ro:La Robine

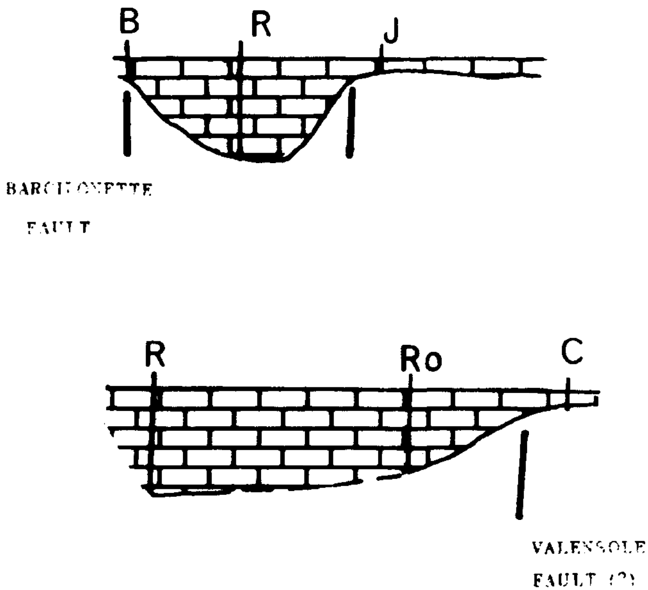
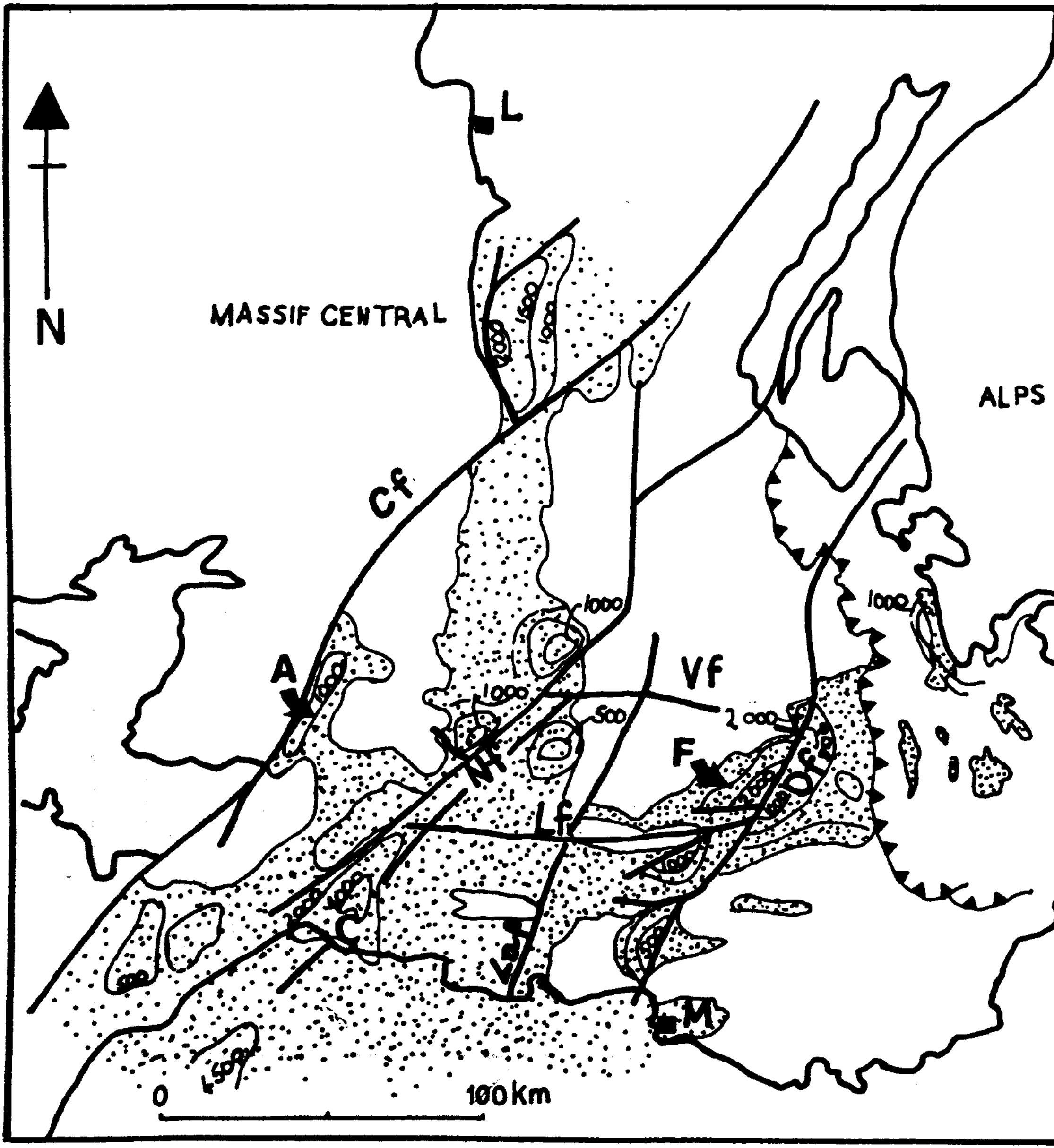


FIG 2.9 Liassic thickness variations within the Digne thrust-sheet clearly show its half-graben form. The position of the known and inferred Hercynian basement faults are indicated (Liassic thicknesses after Mouterde *et al.* 1984). Note how the NE-SW basement faults are coincident with the position of lateral ramps and lateral hanging-wall folds in the Digne thrust sheet, whilst the NW-SE fault apparently. The developement of Triassic diapirs, and Hercynian basement (Barles) along the NW-SE fault support its interpretation as a major basement fault defining the eastern margin of the Durancian High (see Fig's 2.4 & 2.8)



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L
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1000
1500
2000
500
2 000
100 km
0

FIG 2.10 Isopach map of total Tertiary thickness (preserved) superimposed on a map of the inherited fault framework of S.E France (Isopach data from Debrand-Passard *et al.* 1984) (faults as in Fig 2.4). The S.W Alpine thrust front is marked on (the east of) the map. Note that west of the alpine-thrust front (i) the Tertiary depo-centres are coincident with the position of the inherited faults (ii) maximum thicknesses were achieved in the centre of the Rhone valley, and offshore (Gulf of Lions).

A:Ales Graben F:Forcalquier Graben
C:Camargue Graben

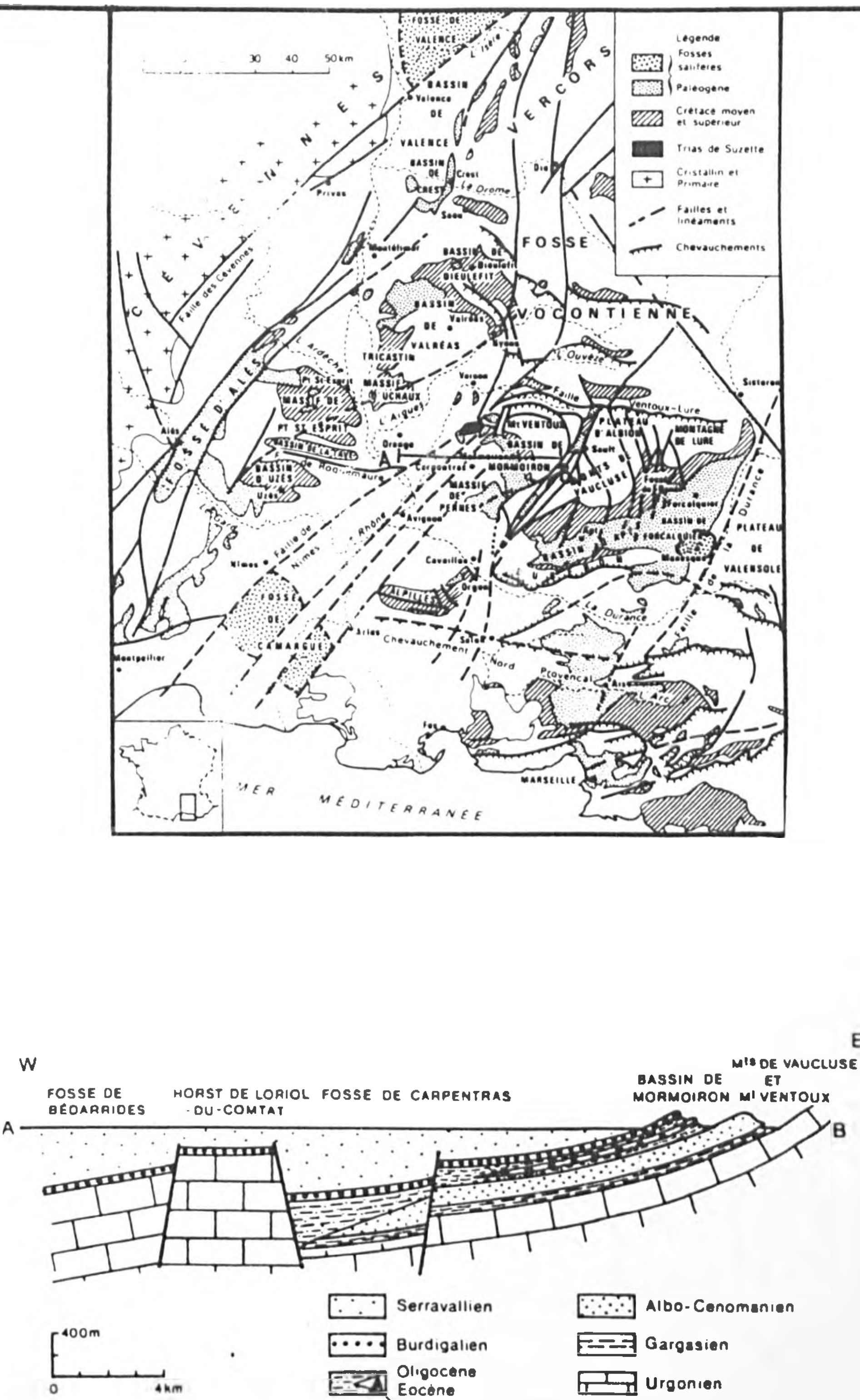


FIG 2.11 (i) Map of the NE-SW (N50E) high-angle faults and Tertiary extensional basins of S.E France (from Triat & Truc 1983). The Durancian basin forms the eastern part of this map, south of the Ventoux and Lure mountains and west of a line through Sault, Cavaillon, and Orgon. (ii) Reconstructed cross-section through the Carpentras graben showing that the NE-SW faults actively defined Tertiary basins (from Triat & Truc (*op. cit.*).

FIG 2.12 Inherited faults and fault blocks of The Durancian Basin. Note how (i) the position of the buried faults is marked by the developement of Triassic diapirs (ii) thrusts and folds in both the allocthonous and autochthonous zone are coincident with the position of the inherited high-angle faults (iii) the degree of inversion of these high-angle faults decreases south-westwards into the alpine foreland.

Allochthonous alpine thrust front
Parautochthonous thrusts of the autochthonous external zone
High-angle "inherited" faults
Triassic evaporite diapirs
Hercynian crystalline basement (Barles)

B-Barles D-Digne S-Sault
Section A-A' - Fig 2.13
Section B - Fig 2.8

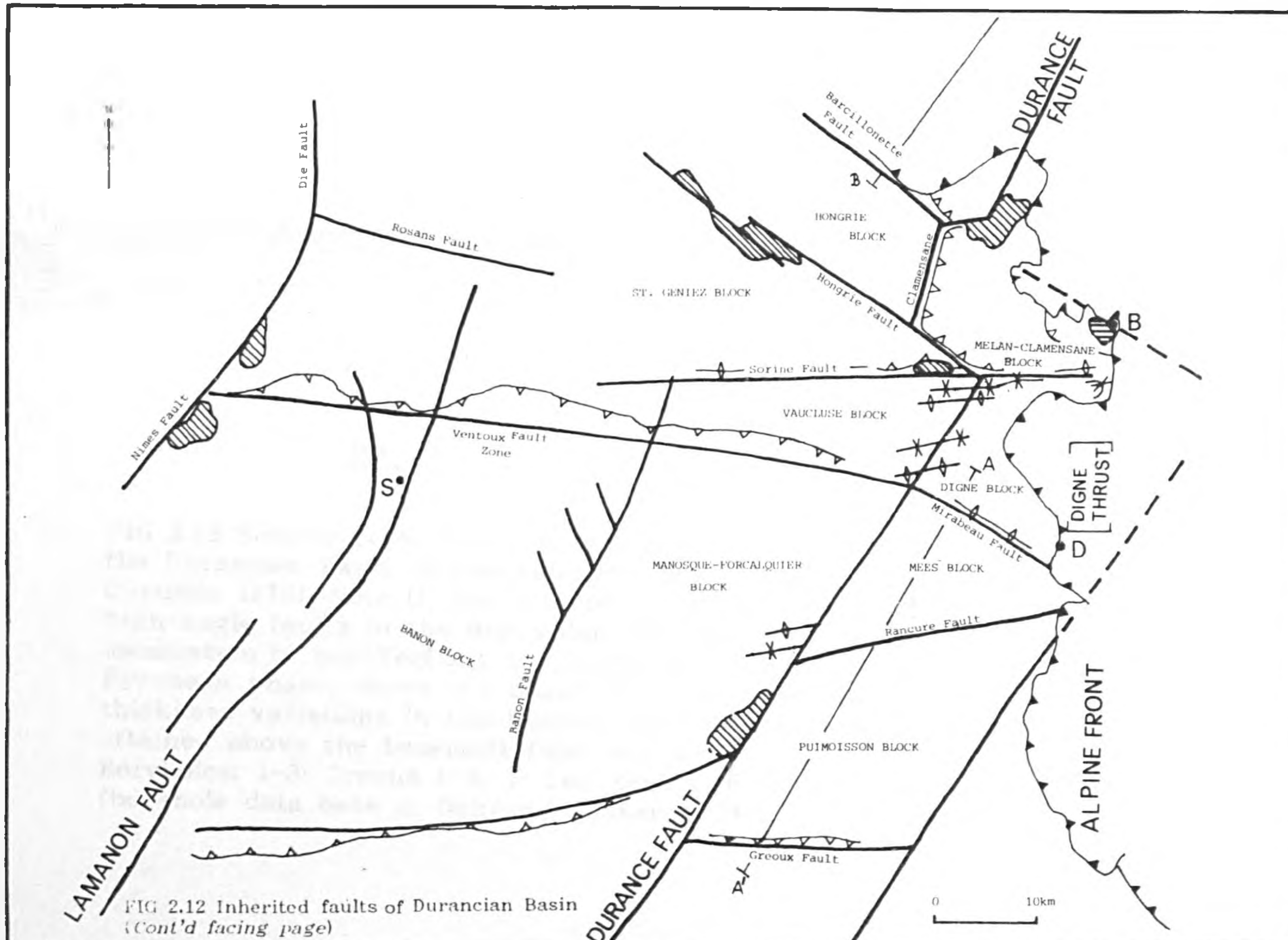


FIG 2.12 Inherited faults of Durancian Basin
(*Cont'd facing page*)

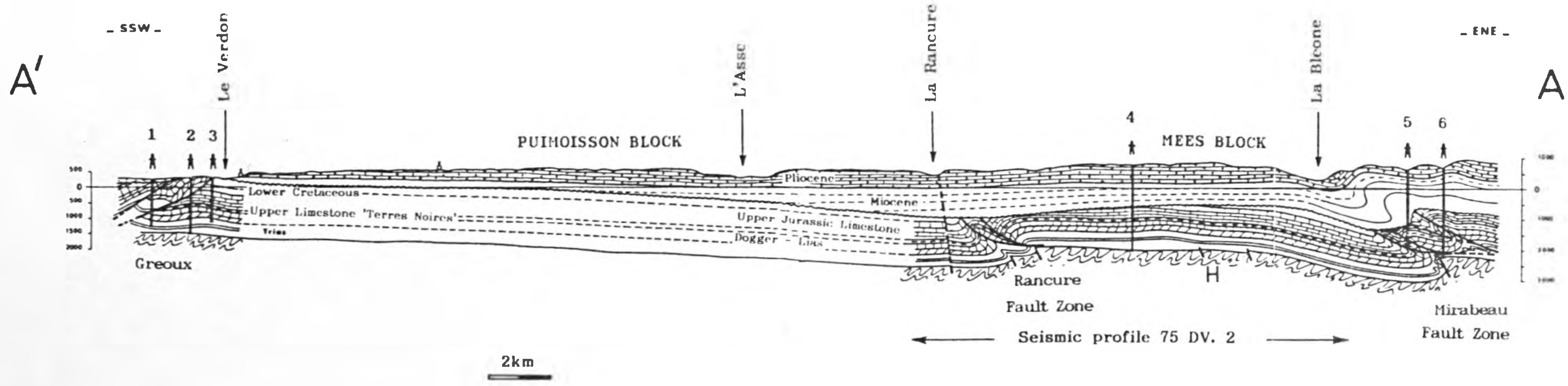


FIG 2.13 Seismic reflection profile across the eastern section of the Durancian Basin (Digne-Valensole basin) (from Dubois & Curnelle 1978). Note (i) the interpreted developement of high-angle faults in the Hercynian basement (H) (ii) the localisation of pre-Tertiary thrust-folds, interpreted as Pyrenean phase, above the basement fault zones (iii) the lateral thickness variations in the Miocene with maximum thicknesses attained above the basement fault zones.
Boreholes: 1-3: Greoux 1-3, 4: Les Mees 1, 5-6: Mirabeau 1-2 (borehole data base in Debrand-Passard 1984).

CHAPTER 3
THE MOLASSE ROUGE FORMATION

3.1 Introduction.

The Molasse Rouge Formation forms the basal fill of the northern Durancian basins and comprises late Eocene (Ludian) - lower Miocene (Aquitanian) continental deposits which unconformably overlie a folded Mesozoic substrate.

The formations' name reflects its predominantly reddish, and coarse grained nature. The formation forms ~~outcrops~~ immediately west of the alpine thrust front, forming an almost continuous NNE-SSW trending strip along the western margin of the study area, and a series of outliers in the extreme north of the area (Fig. 3.1)

Lateral thickness variations in the formation are interpreted to record its deposition within a system of differentially subsiding, extensional fault blocks and palaeovalleys.

3.2 Stratigraphy

The formation was initially attributed an Oligocene age by Goguel *et al.* (1964) with the base of the formation subsequently being more accurately dated (microfaunal evidence) as of Upper Eocene (Ludian) age by Gigot (1974), and Rousset (1983) ascribed an Oligocene age to the formation in the north of the area (Esparron), but on what evidence is not clear.

There is no established stratigraphy within the formation. The top of the formation is marked by Burdigalian age, marine sequences (Goguel 1964) at the base of the Marine Molasse Formation.

3.3 Previous Work.

Goguel *et al.*, (1964) defined the 'Molasse Rouge' in the area and described it as a sequence of locally and distally derived conglomerates, red sandstones and mudstones.

More recently, Gigot (1973) recognised the development of an association of sedimentary klippes and 'megabreccias' at the base of the formation in the extreme west of the area.

Rousset (1983) recognised a twofold lithological division to the succession in the north of the area (Esparron), with:

(1) a basal level of locally sourced conglomerates shed towards the north-east,
(2) an upper level of sandstones and mudstones sourced from a more internal part of the Alps.

3.4 Sedimentology of the Molasse Rouge Formation:- Introduction to Approach

In this thesis a facies analysis approach has been undertaken, with the sequences described in terms of **facies associations.**

After Allen (1965), a twofold division to the alluvial lithofacies of the formation is recognised. The first group, termed the **coarse members** comprises gravel and sandstone facies indicative of deposition within a river channel. The second group, called the **fine members** comprises mudstone, siltstone and sandstone facies deposited on alluvial plains. In areas of favourable exposure coarse members are related to their external channel body geometries as classified by Atkinson (1983) (Fig. 3.3). Discussion of channel processes and morphology as related to these characteristics is dealt with in the succeeding sections.

Alluvial associations have also been classified in terms of the characteristics of their alluvial channels and overbank sequences using the scheme of Galloway (1981) (Fig. 3.4).

3.5 Palaeogene Climate.

The study area has undergone a progressive northward drift since the late Eocene (Smith 1971, Smith *et al.*, 1981), from a latitude of ~ 37° N, through a late Miocene latitude of ~40° N, to its present position of 44° N, 6° E.

Climatic conditions during the Palaeogene of S.E France have been studied by Carbonnel *et al.*, (1972), Schuler & Sittler (1976), Truc (1978) and Cavalier *et al.*, (1984).

Pollen analyses in the Mormoiron basin (Rhone) by Truc (1978) show that the region had a warm probably sub-arid climate during the lower Ludian, which progressively became arid by the upper Ludian. The evaporitic sequences deposited throughout the Ludian of the region characteristically have the form of thin alternating mudstone and evaporite layers indicative of short time scale, probably seasonal, alternation of arid and relatively humid phases.

Pollen studies from the Stampian at Manosque, Provence (Schuler & Sittler 1976) indicate a hot, arid climate and distinguish an upland fauna (dominated by conifer pollen) from a more humid lowland fauna (with palms). An increase in aridity occurred towards the end of the Stampian.

3.6 Terminal Fan Systems - An Introduction

Detailed sedimentary analysis of the Molasse Rouge formation has shown that the principal depositional systems developed were continental, terminal fan systems. The classification schemes, and diagnostic features of alluvial fan systems, and in particular terminal fan systems will be reviewed before they are applied to the studied successions.

Alluvial fans are defined as "bodies of detrital sediment built up by a mountain stream at the base of a mountain front and commonly having the shape of a segment of a cone". (Heward 1978, see also Denny 1967, Bull 1972). Two principal types of alluvial fan are recognised (Collinson 1978), namely:

(1) **Stream dominated ('humid') fans** - those fans whose surface processes are dominated by streams flowing in channels. These fans are characterised by bedload streams though streams with a high suspended load do occur rarely (Collinson 1986).
(2) **Mass Flow ('semi-arid') fans** - those fans in which mass-flow processes are important.

Terminal fans are a particular type of fan in which discharge progressively decreases down fan by a combination of evaporation and infiltration such that no water exi ts from the basin by surface flow (Friend 1978, Collinson 1986). These fans occur in arid basins of inland drainage where stream flow is ephemeral (Friend 1978, Parkash, Awasthi & Gohain 1983, and Nichols 1987).

All alluvial fan deposits have a number of characteristic criteria (Bull 1972, Heward 1978), namely:
(i) A down fan and downflow decrease in grain size and bed thickness and increase in clast sorting.
(ii) The development of cyclic sequences on a variety of scales.

(iii) Most fan deposits are oxidised - Bull (1972) suggests that a lack of reduced deposits is a characteristic of fan deposits from arid and semi-arid regions. This statement is clearly an over generalisation, with the state of oxidation of the sediments being a reflection of the local drainage, so that the distal sections of fan systems in particular may lie in poorly drained, low lying areas.

Terminal fans have, in addition, the following characteristics:

(i) distributary channels decrease in depth downstream, terminating as shallow sheet flows.

(ii) the deposition of thick sequences of inter-channel deposits.

(iii) following on from (i and ii), the degree of interconnectiveness of channel bodies decreases downstream.

Based on the preceding discussion, two distinct types of terminal fan system are recognised within the Molasse Rouge Formation (Fig. 3.2), namely:

(1) **Mass Flow Terminal Fan System** - The Vancon fan system. Also included within this category are the fills of two palaeovalleys, the Peroure and La Tour palaeovalleys.

(2) **Stream Dominated Terminal Fan System** - The Esclangon fan system.

3.7 Mass Flow Terminal Fan Systems

The outcrop pattern of the mass flow alluvial fan deposits of the formation are detailed in Fig. 3.1 and shows that discrete fan systems were developed. The thickest (~700m), and most laterally continuous successions are those of the Vancon fan system which developed in the west of the study area, with the Peroure and La Tour fan systems restricted to shallow (~70m) palaeovalleys in the north-east of the area.

3.7.1 Petrography of the Mass Flow Systems

The sandstones of these fan systems typically have a reddish to yellowish brown colour (10R5/4 - 10YR5/4), and are poorly sorted within the channelised facies.

In thin section they comprise 90% calclithic grains of micritic and biomicritic limestones. The grains are subangular to subrounded. Unimodal strained and unstrained quartz form less than 5% of the grains being subordinate to chert. Glauconite is a common minor constituent forming 5% or less of the grains in the Vancon system.

The conglomerate grade clasts of these deposits are almost exclusively of Cretaceous and Upper Jurassic carbonate lithologies comprising pale grey and ochrous, micritic and biomicritic limestones. Chert which may form 5% of the clasts, is also present as nodules within grey micritic Tithonian limestone clasts.

Interpretation

The petrography of these deposits clearly indicates their local derivation from Mesozoic carbonate lithologies, and this is supported by the dispersal pattern of the systems. Limestone clasts have been identified as of Upper-Middle Cretaceous and Upper Jurassic lithologies. Chert was derived from the nodular horizons of the Tithonian limestones, whilst glauconite is abundant in a number of the Cretaceous carbonate formations.

3.8. The Vancon Terminal Fan System

Deposits of the Vancon Fan System form a continous outcrop along the western margin of the study area (Fig. 3.4), and are characterised by channelised bodies of conglomerate set within mudstone.

Lateral thickness variations in the successions show that the system was deposited within a subsiding 'trap-door' graben (Harding 1978) bounded to the east by the major Durance fault.

The deposits of the fan system form three distinct facies associations, namely:

1) Proximal fan facies association.
2) Medial fan facies association.
3) Distal fan facies asociation.

3.8.1 Facies Associations.

3.8.1.1 Proximal fan facies association.

This facies association is characterised by massive, horizontally stratified, conglomerate sheets (Figs. 3.6 - 3.8).

Two principal 'end-member' conglomerate deposits are recognised:
(ii) A poorly sorted, polymodal, clast-, or rarely matrix-supported, conglomerate with a mud rich pebbly siltstone matrix (Figs. 3.8 - 3.9). Clasts are sub-angular to sub-rounded, and of boulder - pebble grade, ranging up to 3m in size. The beds are sheetlike but may have cm to dm scale erosional relief, and have thicknesses of 0.3 - 1.5m. The beds are typically ungraded, and lack internal organisation or stratification sequences, but may show a weak normal, or inverse grading with outsize clasts at the top of strata.
(ii) Sheets of moderately well sorted, imbricate conglomerates which show a normal grading (Fig. 3.8). The conglomerates are polymodal, of boulder-pebble grade, with large boulders of the order of 1m restricted to the base of beds. The sheets have sharp erosive bases which are planar with an irregular cm scale basal relief, or may take the form of shallow channels of dm relief, and metre to tens of metres width. The shallow channel scours may contain low angle stratification which dips at 5-14°, transverse to flow.

Crude coarsening and fining upward sequences of 4 - 15m thickness occur within the conglomerates, but the conglomerates more typically show a random organisation.

Pale grey, algal laminated limestones form subordinate deposits (<1%) within the facies. The limestone beds are of 0.10 - 0.80m thickness and of metre to tens of metres (< 20m) lateral extent, with erosionally defined upper and lateral limits (Fig. 3.10).

As described by Gigot (1973) large olistoliths of Cretaceous limestone, and limestone and marl lithogies occur in this association (Figs. 3.11 - 3.13). The largest olistoliths occur at the Ravine de Taravon where blocks of Cenomanian limestone and marl 40m thick and 300m long, are enclosed within conglomerate facies. Bedding in the limestone and marl blocks is irregularly folded (Fig. 3.13). In the massive limestones, cleaved shear zones develop which commonly have a ramp and flat morphology, with flats paralleling bedding and ramps developing along jointing. Small duplex structures occur occasionally.

Boulder grade breccias comprising a polymodal assemblage of clast supported, angular clasts of the same lithology as the olistolith form metre thick zones at the top of limestone olistoliths (Fig. 3.12). The contact with the massive limestone is gradational with the larger breccia clasts seen to be collapsed joint blocks in which the jointing planes have been expanded and filled with breccia.

No organisation or trend to the occurrence of olistoliths within the association was apparent, though this may reflect the poorly exposed nature of the association.

Interpretation

Horizontally stratified 'sheets' of imbricate, and disorganised conglomerate, lacking a channelised organisation are typical of the proximal deposits of arid and semi-arid alluvial fans (Denny 1967, Hooke 1967, Bull 1972).

The coarser, disorganised conglomerate beds with a matrix supported fabric are characteristic of cohesive debris flow deposits.

The clast supported, disorganised conglomerates, with their erosive bases, are interpreted as non-cohesive sediment gravity flows, in which the clast support mechanisms included both turbulence and dispersive grain pressure (Blair 1987).
Debris flow deposits are important deposits of semi-arid alluvial fans being most prevalent at the fan heads (Hooke 1967).

The better sorted, clast supported conglomerates are interpreted to be the deposits of sheetfloods, with the scoured base, grading, and imbrication of the conglomerates, reflecting deposition from bedload by a turbulent, fluidal flow. As discussed by Bull (1972) sheets of well sorted conglomerate are characteristically deposited on arid fans, by flows emerging as sheet floods from the downstream end of channels below the fans intersection point. The development of small scour fill channels within these deposits may be attributed to waning flow stage modification of sheetfloods when the flow splits up into a series of shallow, braided channels (Hooke 1967)

The development of limestone indicates that at times, prolonged periods of non-sedimentation (clastic) occurred between conglomerate deposition. Similar limestones having been described from a palustrine environmemt by Freytet (1973), and from abandoned, alluvial fan, conglomerate channel fills by Nickel

(1982).

Olistoliths are defined as "masses of homogeneous rock which have slid under gravitational forces" (Abbate *et al.*, 1970). The development of internal shear zones within these olistoliths indicate that they underwent compressional deformation during emplacement. Marl beds of the blocks may have acted as 'lubricants' at the base of the olistolith reducing the frictional resistance to sliding (see Hubbard & Rubbey 1959).

The overlying breccia units are interpreted to be due to the in-situ reworking of the tops of the olistoliths, by the process of solution widening of joints and subsequent solution cavity collapse.

Olistolith deposition was interspersed with lower energy periods of breccia and conglomerate sheet facies deposition. This pulsed nature of facies development may suggest a spasmodic, tectonic creation of relief from which the olistoliths were shed, perhaps triggered by seismic shocking.

Analogous sheet conglomerate facies have been described from the proximal regions of modern day, arid and semi-arid, alluvial fans by Hooke (1967), Bull (1972), and Blair (1987). These authors noted that coarsening and fining upward sequences on the scale noted in this facies association were generated by the process of fanhead channel avulsion, or incision, causing shifting of the fans active lobe.

An analogous conglomerate dominated facies association containing olistoliths has been described from an Eocene age, fault-scarp fringing alluvial fan system (Cabrera *et al.*, 1985). The association was attributed to deposition within the proximal zone of the fan with olistolith development interpreted to record syn-sedimentary tectonic activity. Alluvial conglomerate and olistolith deposits have also been described from the proximal region of alluvial fans fringing the strike-slip, fault bound, basins of California (Crowell 1974, Link & Osborne 1978)

3.8.1.2 **Medial fan facies association**

This facies association comprises coarse member channelised conglomerate bodies and mudstone dominated fine members, which form fining upward sequences, of 20m - 50m thickness (Fig. 3.14 -

3.16).

The base of fining upward sequences is marked by a conglomerate channel body having a multistorey, ribbon (type 2b) (Fig. 3.15) or tabular (type 1b) geometry (Fig. 3.16). Channel storeys are 2-4m thick with channel bodies being 4-10m thick.

The channels have a fining upward fill with the basal section comprising a polymodal framework-supported, boulder - cobble conglomerate (Fig. 3.17), with clasts of 0.25 - 1.2m (typically 0.25 - 0.35m). This forms m-scale units of massive or horizontally stratified conglomerate. Shallow, trough shaped sheets of conglomerate may also be developed. Individual beds attain thicknesses of up to some 90cm, and commonly have an imbricate fabric, with the clasts' ab plane dipping upstream, and the a-axis being transverse to flow. Erosional lenses of cross-stratified sandstone and well sorted pebbles may be preserved between the coarse conglomerate beds.

Sigmoidal conglomerate stratification which dips perpendicular to palaeoflow, is occasionally developed. This stratification downlaps onto the erosive channel base passing up dip into planar stratified finer conglomerates and sandstones (Fig. 3.19). Sets of planar cross-stratified conglomerate are also rarely developed in which the sets are of 0.3 - 1m height. The top of these cross-strata may be convex upwards, passing up-dip into horizontally stratified conglomerate (Gm facies).

The channel fill fines upward into thinly interbedded, pebble conglomerates, and planar laminated, or trough stratified sandstone. These may form low angle ($< 15^{o}$) stratification dipping upstream, or oblique to the palaeoflow (Fig. 3.20). The very upper part of the channel fill comprise thinly interbedded siltstones and sandstones with desiccation cracks.

The channelised conglomerate bodies pass up into a mudstone dominated fine member comprising an association of small ribbon conglomerate channels, sheet sandstones, and heterogeneous mudstone (Fig. 3.14).

The small scale, single storey ribbon channels have a well sorted, cobble-pebble imbricate fill, and trend parallel to the larger channel bodies. They are isolated within the mudstone, or develop within units of sheet sandstone.

The sheet sandstones are graded with planar erosive bases and fine upward from a pebble conglomerate or coarse sandstone into mudstone. The sandstone may be structureless or show a weak planar- or ripple- lamination. The thicker beds show trough cross stratification of up to some 10cm. Small, ribbon scours of dm height, and dm - m scale (<3m) filled with ripple laminated sandstone are also developed. These beds form sandstone dominated, dm - m scale tabular units interbedded within the mudstone, and showing fining-, and coarsening-upward trends. These may commonly be traced laterally into the larger scale conglomerate-filled channel bodies. In these cases they may be seen to thin and fine away from the channel over a distance of tens to hundreds of metres (Fig. 3.18).

In complete fining-upward sequences, the top of the fine member interval is marked by a unit of heterolithic mudstone.

The fine member mudstones may show one of two styles of pedogenesis. In the first they have a pale grey colour (N5) and a yellow brown mottle (10YR6/6) (Fig. 3.16), whilst in the second, mudstones have a reddish to yellowish brown colour (10R4/6 - 10R5/4) and a pale to purplish grey (N7 - 5RP6/2) mottle (Fig. 3.15). Both are characterised by the development of calcium carbonate nodules, and show a vertical increase through the sequence, in their level of pedogenesis.

In the red mudstones, pedogenic modification takes the form of mottling with or without diffuse glaebules (nodules) (Stage 1) (Fig. 3.24) at the base of the member, and tubular nodule horizons (stage 2-3) (Fig. 3.24) at the top of the member. In the grey mudstones, gley mottling and diffuse glaebules, or thin tubules form at the base of the member and are succeeded by thin, micritic limestones with a nodular overprint at the top.

At a well exposed level in the Jonchier section, these large scale coarse-fine member sequences may be related to the lateral migration of the large scale channel conglomerate bodies (Fig. 3.15 (iv)). Vertically and laterally offset fining-upward sequences (10m -40thick) may be seen to be generated by the lateral migration of domains of multistorey ribbon channels across the line of the section.

Interpretation

Fining upward cycles from a channelised coarse member into a mudstone dominated fine member may be interpreted in terms of the vertical passage from a channel into an interchannel area (Allen 1964).

The imbricated and horizontally stratified conglomerate (Gm facies) of the channel fills are interpreted to be the deposits of longitudinal bars which migrated during peak flow stages (Rust 1972, Smith 1974, Hein & Walker 1977). The interbedded lenses of sandstone represent low flow stage dune deposits, with the well sorted pebbles interpreted to be the fills of ephemeral channels which drained the longitudinal bars (Rust 1972)

The large scale planar cross-bedded conglomerates are probably the result of the downstream and, or, the lateral accretion of longitudinal bars during waning flow stages (Hein & Walker 1977). The transverse to flow, sigmoidal conglomerate stratification is analogous to that detailed by Ramos & Sopena (1983), which forms in channels marginal to longitudinal bar facies. The upstream dipping conglomerate and sandstone stratification at the top of some channels is considered to reflect accretion on the upstream margins of longitudinal bars during lower and waning flow stages. The fining upward of the channel fill reflects a progressive waning of the flow with the vertical filling and abandonment of channels. Fluctuating, low energy flow conditions in the upper sections of channels are attested to by the association of small 3-D dune deposits and thinly interbedded mudstone and siltstone, with mud cracks indicating that channels periodically dried out.

In summary this facies is considered to represent the vertical fills of braided, conglomerate dominated, ribbon channels. The closest modern analogue to this facies are probably that of the braided Donjeck river (Rust 1971) whose deposits are dominantly of horizontal bedded conglomerate.

The overlying fine member records that deposition in the interchannel areas varied between two extremes. The grey mudstones with their yellow-brown mottling are typical of hydromorphic, gley soils developed in poorly drained flat lying areas, with high but fluctuating groundwater levels (Buurmann 1980) (see Fig. 3.23)

The brownish red-yellow sediments with grey mottling are characteristic of pseudogley soil processes which develop on flat areas with low groundwater level, but impeded drainage (Buurman 1980). The red colouration reflects a dominance of dehydrated ferric iron compounds over hydrated ferric iron and manganese compounds, and a low incidence of organic carbon, with an increase in the abundance of the hydrated iron compounds giving the more yellowish colours (Buurman 1980, Bown & Kraus 1981).

The calcareous nodules within the mudstones are typical of those formed in calcrete, or caliche soils (Collinson 1986). By comparison with recent soils the isolated glaebules (type 1) are analagous to youthful profiles, and the tubular to massive nodular horizons similar to late youth, to early mature calcrete profiles (Fig. 3.24) (Reeves 1970, Steel 1974, and Leeder 1975). The isolated glaebules (type 1) are analagous to youthful calcrete profiles, with the tubular to massive horizons (type 2-3) similar to mature calcrete profiles. In the pseudo-gley soils the carbonate was precipitated from downward percolating groundwaters (*per descencum* model of Goudie (1973), but in the gley soils the evaporation of local carbonate - rich groundwaters (*per ascencum* model of Goudie (1973) is a more likely source of carbonate (Atkinson 1983).

The graded sandstone beds of the fine member are interpreted as the deposits of waning, sheet like turbulent flows, with their lateral connection to channel bodies suggesting that they are overbank, sheetflood or crevasse splay deposits. A possible alternative explanation for those units not directly associated with channel bodies is that they are the deposits of sheetfloods issued from the downstream end of terminal distributary channels (Cabrera *et al.*, 1985).

In summary, the fine member shows that drainage in the interchannel areas varied from being moderately well drained with periodic impeded drainage (pseudogley), to poorly drained with ephemeral ponds. Overbank sedimentation from large channels generated units of graded sandstone and conglomerate. The vertical increase in the level of maturity of the calcrete profiles reflects the decreasing rates of sediment supply with distance from the main distributary channel.

The repeated coarsening and fining sequences of this association can be related to the progressive lateral migration of the main distributary channels (as exposed at the Jonchier section), or the rapid lateral shifting of the channels. Channel migration may have been in response to inherited conglomerate body topography (Allen 1978), fanhead entrenchment (Heward 1978), or the superimposition of a tectonic slope on the fan (Bridges & Leeder 1979, Alexander & Leeder 1987).

Allen *et al.*, (1983) described a similar association of large multistorey and small, single storey conglomerate bodies isolated within fine member deposits, with the larger channels interpreted to represent the principal distributary channels of an alluvial fan, between which existed small, ephemeral streams.

Cabrera *et al.* (1985) described analagous, multistorey channels isolated within fine member deposits. These were interpreted as the fills of the main braided distributary channels of a proximal fan fringe and medial fan areas, with the small ribbon channels and fine member deposits developing in the inter-channel areas.

3.8.1.3 **Distal fan facies association**

This association consists mainly of heterolithic mudstones and sheet sandstones, and is characterised by the development of stacked horizons of carbonate nodules (Figs. 3.21 - 3.22).

Silty mudstone dominates the heterolithic mudstone with thinly bedded siltstones also developed. Mudstones exhibit desiccation cracks.

Decimetre to metre (<2m) scale units of sheet sandstones are developed within the mudstones (Fig. 3.21). These commonly form small coarsening-upward sequences passing from mudstone to sheet sandstones, with thin conglomerate beds and small (dm scale) ribbon conglomerate bodies at the top. The unit may show an abrupt or gradational passage back into the mudstones. Small (dm scale) single storey, ribbon bodies of imbricate pebble conglomerate may also occur isolated within the mudstones. A few channel fills have tabular beds of algal limestones intraclast gravels in their upper part.

The mudstones have a reddish brown (10R4/6 - 10R5/4) colour and a pale to purplish grey (N7 - 5RP6/2) mottle. The mottling typically shows a trellis network dominated by a vertical fabric of shafts commonly showing a downward diverging branching. Mottle channels show a colour zonation, with a dark to purplish grey (N4

- 5RP4/2) central zone passing out through a pale grey zone (N5) into an outer yellow-brown (10YR 6/6 - 10YR5/4) zone.

The mudstones are characterised by the development of beds of calcium carbonate nodules (Figs. 3.25 - 3.28). Massive nodular beds develop with a dm-m spacing which show a well defined vertical organistion (Figs. 3.24 - 3.26). At the base, isolated 'glaebules' (nodules) develop (Stage 1) within the pale grey zone of the mottle channel (Fig. 3.27), vertically merging to form cylindrical 'pedotubules' (up to 2m length) (Stage 2). These commonly have branch downwards. Vertical sheets of coalesced glaebules, up to 1-2m in length, also develop which may be seen in plan view to comprise a series of polygons (Fig. 3.28). With progressive mergence of nodules, the relict host sediment is restricted to isolated lenses within a carbonate framework (Stage 3 horizon). In the best developed profiles a massive nodular carbonate horizon is developed having a weak horizontal fabric (Stage 4 horizon). In thin section this is seen to be defined by thin horizontal calcite filled fracture veins containing angular clasts of carbonate spalled-off the margins of the fracture.

The thickness of these carbonate profiles is extremely variable, and they are commonly vertically superimposed, but complete Stage 1 to Stage 3/4 profiles range in thickness from 0.5 to 4.0m.

The association may show a fining upward sequence, with the passage into grey mudstones with a yellowish-brown mottle. Massive, micritic limestone beds, often with a nodular fabric, and containing *Planorbis* gastropods develop within the mudstone. These may show a laminate fabric, which in thin section comprises laminate algal bands. Thin (cm thick) layers of granular conglomerate, comprising sub-rounded micrite intraclasts with coatings of micrite are also developed within the limestone.

Interpretation

The development of mudstone dominated sequences with pedogenic carbonate horizons reflects deposition in an area with low sediment supply rates.

Sheet sandstone and mudstone associations similar to those described here have been interpreted to represent the progradation and abandonement of depositional sheetflood lobes in a distal fan setting (Heward 1978). Cabrera *et al.*, (1985) also described analogous sheet sandstone coarsening-upward sequences which passed up into small conglomerate channels, and interpreted them as small depositional lobes on distal, alluvial fan fringes. By comparison, the small-scale sandstone sequences of this association are interpreted to record the periodic migration of small braided channels and their associated terminal sandstone sheet lobes into the distal regions of an alluvial fan. Alternatively they may have developed as the overbank deposits of the conglomerate ribbon channels. These small imbricate gravel channels are similar to those described in the fine member intervals of the medial fan facies association and are similarly interpreted as the deposits of small, ephemeral braided channels. The brownish-red colour and grey mottling of the mudstones is characteristic of pseudogley soil processes which are operative in low lying areas with a low groundwater level (Buurman 1980). The grey mottling is produced by the leaching of the sediment along rootholes, burrows *etc.* by downward percolating stagnant water.

The carbonate nodules are characteristic of caliche or calcrete soils (Reeves 1970), which form in warm to hot, semi-arid or arid climates, with a seasonal rainfall. The form of the carbonate deposits is relatable to the stages of maturity in recent and ancient paleosols (Reeves 1970, Steel 1974, Leeder 1975). The stage 3 profiles are comparable with mature calcrete soils with the vertical and branching form of the pedotubules in these profiles suggesting that they formed about vertical tap roots. Their length records the deep ground water level, and the well drained nature of the soils (Sarjeant 1975). The polygonal framework of the carbonate sheets is characteristic of non-orthogonal shrinkage crack systems typically developed on playas (Allen 1980). Alternatively they may be growth pressure release fracture fills ('expansion polygons') (Collinson 1986). The stage 4 profiles are similar to late mature calcrete soils (Reeves 1970), and are considered by Leeder (1975) to have a minimum formation time of 10,000 years. The horizontal calcite sheets with brecciated clasts at the top of these beds are similar to those detailed by Steel (1974) and Reeves (1970) and interpreted to be infilled horizontal dessication cracks in plugged caliches.

In summary, the dominance of psuedo-gley paleosols, and the development of pedo-tubular calcretes suggests that this region was a low lying, vegetated area of impeded drainage, characteristic of a playa environment. Active sediment supply in the form of the development of sheet sandstone lobes and shallow, ephemeral channel fills suggests that the area lay in the distal part of an alluvial fan.

The vertical transition into grey, mottled mudstones reflects a change to gley soil processes which operate in poorly drained areas with a high, but fluctuating groundwater level (Buurman 1980). The limestones with their freshwater fauna are interpreted as lacustrine deposits, with their nodular overprinting indicating that the lakes were ephemeral and subjected to soil reworking (palustrine limestones).

This trend from pseudo-gley soils, to gley soils with limestones, is similar to that described by Nickel (1982) and interpreted to record the passage from a distal alluvial fan to a marginal lacustrine environment.

3.8.2 The Vancon Alluvial Fan System - A First Order Sequence

A first order sequence (mega-sequence) is considered to record major progradations and retrogradations which affected the whole alluvial fan system, and on this criterion the Vancon fan system as a whole may be considered a mega-sequence (Fig. 3.30). The mega-sequence shows the following trends:

a) an upward decrease in the scale and degree of complexity of conglomerate channel bodies.

b) an upward decrease in the maximum particle size (MPS) from boulder to small cobble and pebble grade.

c) an upward decrease in the coarse to fine member ratio, from values of <90% in the proximal fan association at the base of the sequence, to some 20-30% in the medial fan section, to negligible in the distal fan system.

d) an upward increase in the level of pedogenesis of the fine member intervals.

The base of the mega-sequence comprises a unit (up to 150m thick) of conglomerate dominated deposits of the proximal fan facies association, including olistostromal blocks. This extends from Volonne in the south of the study area as far north as St. Symphorien (Fig 3.1), beyond which it passes into medial fan facies association deposits in the Jonchier section.

This lower unit passes gradationally into a thick middle unit (up to 600m thick) characterised by stream dominated facies of the medial fan facies association. This middle unit comprises repeated fining-upward sequences of multi-storey conglomerate bodies and mudstone dominated fine members. Toward the top of this unit in the north of the basin, channel bodies decrease in dimension and mark the transitional passage into a fine grained upper unit.

The top of the mega-sequence in the north of the basin comprises a thick unit (up to 100m) of mudstones with stacked, mature calcrete profiles of the distal fan facies association. In the south, medial fan facies association deposits persisted.

Palaeocurrent analysis of the conglomerate facies of the proximal and medial fan facies associations (Fig. 3.31) clearly show that the palaeoflow of the system was toward the W-SW and that it was sourced from Mesozoic sequences to the east.

Interpretation.

The large olistolithic blocks and sheet flood conglomerates at the base of the mega-sequence are interpreted to represent the sedimentary response to a phase of rapid tectonic uplift along the Durance fault zone at the eastern margin of the basin. The creation of topographic relief on the Mesozoic basement was accompanied by the development of an arid type alluvial fan. Pulsed input of olistoliths was probably triggered by fault movement or seismic shocking and interspersed with periods of deposition in the proximal part of the alluvial fan system.

The vertical passage into cyclic sequences of stream and interchannel deposits, is attributed to deposition in a proximal alluvial fan fringe to mid fan environment.

The development of stacked calcrete profiles and transition into palustrine conditions at the top of the succession reflects the final phase of fan retrogradation. The absence of this level in the south of the basin (Volonne and Sourribes sections) may reflect either that increased rates of sediment supply compensated for the transgressive event in these areas, or that the upper part of the mega-sequences were reworked during the transgressive event at the base of the overlying Marine Molasse formation. The accumulation and preservation of such a thick mega-sequence of alluvial fan deposits indicates active basinal subsidence throughout their depositional history.

By comparison with other ancient fan sequences (see Heward 1978, and Collinson 1986 for reviews) the general fining upward of the fan system may be interpreted to reflect:

(i) the erosional lowering of relief and, or, associated erosional scarp retreat.

(ii) a progressive decrease in the rate of the basinal subsidence.

(iii) the source area retreat through backfaulting (see Heward 1978).

The fining upwards of the upper part of the mega-sequence may also reflect the influence of a rising base level associated with the Burdigalian marine transgression. A similar interpretation was suggested by Steel (1974) for fining upward mega-sequences in the New Red Sandstone fan systems which are succeeded by transgressive Jurassic deposits.

3.8.3 Lateral Variation in Sequences of the Vancon Fan System

Graphic logs through the Vancon fan system (Fig. 3.31) reveal:

(i) a marked lateral variation in the thickness of the sequences in a perpendicular to palaeoflow sense, ranging from a maximum of some 700m at Jonchier, to only some 100m at Volonne.

(ii) a lateral and vertical variation in the development of gley and pseudo-gley characteristics to the medial fan facies association.

(i) Lateral thickness variations

This thickness variation takes the form of a half-graben fill defined in the north by the Sorine fault, and to the south by the Ventoux fault, with a sub-basin developed in the hanging wall block, defined by the Citadelle fault (Fig. 3.31).

Gravel channel body thickness shows a lateral variation

(Fig. 3.32). Maximum channel body thicknesses are present in the Jonchier and St Symphorien sections in the axis of the half-graben where channels are up to 4m thick and conglomerate bodies attain some 10m. There is a secondary occurrence of thick channel bodies in the Sourribes section, in the sub-basin of the half graben, where conglomerate bodies are up to 6m thickness and channels attain thicknesses of some 2.5m.

The channels at these localities are contrasted with those at La Citadelle and Volonne where coarse member bodies are typically of only 1 - 2m, with channels being only 0.5 - 1.5m. This lateral variation in palaeochannel magnitude indicates that the St.Symphorien - Jonchier, and Sourribes areas were preferred sites of the major distributary channels of the alluvial fan system. Palaeocurrent data clearly shows that the coarse members were derived from the east with channel orientations varying between the NNW and SSW (Fig. 3.31). Radial palaeocurrent patterns were not detected.

Coarse member percentages attain values of 21% - 24% in the 'axial' areas of St. Symphorien - Sourribes respectively, whilst in the intervening areas (interaxial zone) of la Citadelle and Volonne they are of only 10% - 14%.

A pattern thus emerges of structurally defined, topographic lows which were the sites of the westward flowing axial drainage of the fans. These were seperated by structurally defined highs (footwalls) characterised by smaller, ephemeral channels. This pattern allows the definition of a larger St. Syphorien fan and a smaller Sourribes fan within the Vancon alluvial fan system.

(ii) **Lateral and vertical variation in palaeo-drainage**

The basal section of the mega-sequences at Jonchier, and St. Symphorien and to a lesser degree Sourribes, show the development of gley soils whilst their upper sections and the complete mega-sequences at Volonne and la Citadelle, have pseudogley characteristics (Fig. 4.31).

These gley soil zones are coincident with the structurally defined lows of the basin, suggesting that during the early phase of the alluvial systems deposition, relatively high ground water levels persisted in the structural lows and conversely low groundwater levels over the structural highs.

Two models are proposed to account for this. The first model (after Bowen 1980) considers the ground water table to be horizontal, with lateral variations in the topographic relief causing changes in the depth above or below the ground water level. Such a model is rather simplistic because it presumes the lack of operation of recharge zones, and a horizontal homogeneity to the substrate. Therefore, a second model is proposed. It is suggested that the Mesozoic 'basement' of limestone and shale lithologies acted as an aquitard or aquiclude beneath the basin (half-graben) causing ground water flow lines to parallel its structurally created topography (see Todd 1980). Flow would have been in response to the energy head, flowing from the highland areas which bounded the basin (areas of high fluid potential) to the structural lows in the adjacent basin (areas of the lowest fluid potential), thus raising the ground water levels and reducing the drainage in those areas.

The development of better drained, pseudogley soils in the upper part of the succession may be attributed to 1) decreased rates of recharge of the ground water system as a consequence of a) reduced runoff resulting from the errosional lowering of relief on the adjacent highland - this interpretation is supported by the general fining upward characteristics of the fan succession. b) climatic change - decreased precipitation as a consequence of climatic change. This is supported by palaeoclimatic data which suggests an increase in aridity at the end of the Stampian (Schuler & Sittler 1979).

3.8.4 An Alluvial Fan Model

A schematic model of the Vancon fan system is presented in Fig. 3.29 in which it is interpreted as a terminal fan system fringing the western margin of the Durance fault. Discharge was westward directed parallel to the trend of the Sorine and Ventoux intra-basinal faults, with the principal distributary channels developed above structurally defined lows. A braided and sheetflood, conglomerate dominated proximal zone, passed down fan into a medial zone with major distributary channels separated by interchannel areas with small ephemeral streams. Channel magnitude decreased down fan with the distal region comprising ephemeral braided streams and their terminal lobes, issuing across a low relief, vegetated area with impeded drainage. The playa area fringing the fan was characterised by high groundwater levels, with the periodic development of ephemeral lakes.

Comparable, fault fringing fan systems have been described by Steel and Gloppen (1980), and Cabrera *et al.*, (1985).

3.9 La Pategue and La Tour Palaeovalleys

3.9.1 Introduction

In the north of the study area, the base of the Molasse Rouge formation comprises a succession of conglomerates and reddish mudstones which unconformably overlie folded, Cretaceous and upper Jurassic 'basement' (Fig. 3.1). The succession is laterally discontinous, attaining a maximum thickness of 70m and is interpreted as the fill of a series of palaeovalleys.

Taking the base of the overlying Esclangon fluvial system as a horizontal datum (Fig 3.33), suggests that three shallow palaeovalleys, scoured into the Jurassic - Cretaceous substrate, are present in the north of the area. Palaeovalley relief was of the order of some 70m and had apparent widths of the order of 3-4km.

The Peroure and La Tour palaeo-valley fills were studied in detail at the localities of la Tour, le Pategue, la Pare, and Peroure (Fig. 3.1 & 3.33) where similar fining-upward mega-sequences from a conglomerate dominated basal section, to an upper, mudstone dominated section are developed. The mega-sequences comprise three facies associations, a proximal, medial, and distal fan association. The first two are similar to those detailed for the Vancon fan system, but, with a number of differences which will be discussed in describing the sections. The latter is unique to these successions and is therefore described below.

3.9.2 Distal fan facies association

This association is dominated by red (10R4/6) silty mudstones (60-70%) within which are interbedded thin cm-dm scale granule - pebble grade sheets (Fig. 3.35 - 3.37). The mudstones typically have a homogenous red (10R4/6) colour, though a weak grey (N5) mottle may be developed. Des cation cracks are found in the mudstone.

The conglomerate beds (averaging 1-8cm, and up to 30cm), typically have non- to weakly erosional bases, and comprise a poorly sorted assemblage of angular - subrounded clasts of granule-pebble conglomerate, supported by a silt rich mudstone matrix. Beds are commonly inverse graded (Fig. 3.37). Small, steep sided ribbons of clast to matrix supported conglomerate, of maximum 20cm depth and 60cm width, and with thin wings of matrix supported conglomerate occur (Fig. 3.35).

The facies association may show a fining-upwards with a decrease in the proportion of thin conglomerates and the interbedding of thin (cm scale) beds of micritic limestone having a nodular overprint. Isolate calcrete nodules (Stage 1) may also develop in the mudstone.

Interpretation

The thin conglomerate sheets display all of the characteristics of cohesive debris flow deposits (Nemec & Steel 1984). Flows apparently had variable degrees of cohesiveness with local erosion by the more turbulent flows generating the small ribbon shaped scours, whose steep sides and low W/H ratio reflect the cohesiveness of the interbedded mudstone.

The predominance of mudstone deposits in the association indicates that extremely low flow conditions prevailed, with the reddened nature of the mudstone indicating deposition in a generally well drained area. The weak grey mottling is characteristic of gley soils, developed by the temporary formation of stagnant surface water (Buurmann 1979), on flat lying areas or depressions with impeded drainage.

The relatively fine grained nature of the conglomerate reflects the low competence of the depositing flow and the relatively great distance from the source area. The high proportion of mudstones indicates that there were considerable periods between conglomerate depositing events.

This association is analagous to the interbedded marl and mudflow deposits described by Bluck (1967) and considered to represent the interfingering of the distal portions of the mudflow sheets into playa marl sediments. The developement of thin limestone beds at the top of the association is interpreted to record the periodic developement of shallow, ephemeral lakes in the more distal playa environment.

3.9.3 Palaeovalley fills

3.9.3.1 La Tour Palaeovalley.

At the la Tour and le Pategue sections in the La Tour palaeo-valley (Fig. 3.1 & 3.33), the basal section of the valley fill mega-sequence is dominated by conglomerate sheet facies of the proximal fan association (Figs. 3.34 & 3.38). Boulder grade material of up to 2m diameter is present in the cohesive, and non-cohesive, debris flow beds (Gms facies) of this asociation.

As the mega-sequence fines upward, cobble grade, sheet-flood deposits (Gm facies) predominate. Within the association metre scale (3 - 10m) intervals dominated by either debris flow, (Gms) or sheetflood (Gm) conglomerates alternate. Thin (dm scale) limestone beds are developed within the conglomerate facies (as detailed from the Vancon Fan System) but form less than 1% of the association. Olistoliths are not developed in this association.

The upper section of the mega-sequence (Fig. 3.33 & 3.38) involves an increase in the proportion of mudstone, to form 40-70% of the succession. At la Tour thin (dm thick), sheetflood, and debris flow conglomerate sheets and ribbon channels, become interbedded with reddish, gley mottled, mudstones to give a medial fan facies association. The single and multistorey ribbon channels (type 2a-b) of imbricate conglomerate (Gravel channel facies), are of the order of 1 - 2.2m thickness, and up to some 15m width. The channelised ribbon conglomerates, and clast supported sheetflood conglomerates (Gm facies) show little difference in their texture. Channel facies are not developed at le Pategue where the upper section shows the rapid passage from proximal to distal facies association deposits.

The top of the sequence at both localities, comprises mudstones and cohesive debris-flows of the distal fan facies association, with thin (cm scale) palustrine limestones at la Tour passing northward into thicker (dm-m scale) palustrine limestone beds and immature calcrete profiles at Peroure.

The conglomerate clasts throughout this mega-sequenceare of two types of (Neocomian - Cenomanian ?) micritic limestone, having a grey (80%), and ochrous colour.

3.9.3.2 **Peroure Palaeovalley**

The Peroure palaeovalley (Fig.3.1 & 3.33) achieves a maximum thickness of 70m at Peroure where it comprises a fining-upward mega-sequence passing from a conglomerate dominated basal 40m, into a mudstone dominated upper section. A similar but reduced thickness (20m) mega-sequence is developed at la Pare (Fig 3.1).

At Peroure the basal section of the mega-sequences comprises an association of type 2a and, 2b, conglomerate ribbon bodies (range from 1 - 2.4m in thickness), sheetflood, and debris flow conglomerates, with metre thick intervals of reddish mudstone (15-20% of sequence). This sequence is attributed to a transitional proximal to medial fan association. The conglomerate sheets occur as isolated sheets or form composite units of metre scale. The thicker sheetflood conglomerate strata are commonly channelised particularly where they overlie a mudstone dominated interval, reflecting a transition stage between a sheet and channelised flow.

At la Pare only small (1-1.5m) ribbon channels are developed at the base of the succession within a limited thicknes of medial fan facies deposits.

The upper 25m of the mega-sequences shows a fining upward into distal fan association deposits. Reddish mudstones forms 60-70% of deposits with thinly interbedded pebble-conglomerate, cohesive debris flow beds.

The clasts of the conglomerate facies of the mega-sequences are of boulder-cobble grade, with 60-70% being Tithonian-Portlandian and 30-40% Kimmeridgian limestones.

The top of the mega-sequences are marked by a thick level of mature calcrete profiles and to lesser degree palustrine limestones attributed to the overlying Esclangon fluvial system.

3.9.3.3 **Interpretation of palaeovalley fills.**

In the La Pategue and La Tour palaeovalley fills the association of sheetflood and debris flow conglomerates in the basal section are considered characteristic of the deposits developed in the more proximal regions of arid alluvial fans. As discussed by Hooke (1967), debris flow deposition predominates immediately below the intersection point (typically near mid-fan). The fining-upward in the basal section may be interpreted in terms of fan retrogradation with the passage from debris flow proximal zone conglomerates, into sheetflood and channel fill conglomerates of the mid-fan region.

The upper section of the mega-sequences with their association of thin debris flows and mudstones are interpreted to represent the interfingering of distal fan deposits and playa deposits. The thin limestone strata reflect that the playa area fringing the fan periodically had a relatively high groundwater level with the development of ephemeral lakes.

In the la Tour palaeovalley fill, the thick, conglomerate that dominates the La Tour section is interpreted as reflecting an axial region of the palaeo-valley, with the thinner and finer grained sequences at La Pategue, being the lower energy deposits toward the margins of the valley. A similar interpretation is given to the thicker and coarser gravel facies developement at Peroure relative to those at la Pare in the Peroure palaeovalley.

Analogous (though larger scale), fining upward alluvial sequences have been described from alluvial fan successions by Bluck (1967) and Steel & Wilson (1975). Bluck (*op cit.*) suggested that vertical decreases in grain size indicated a recession of the source area as a consequence of the progressive reduction of relief in the source area with time. Similarly, Steel (1975) suggested that this type of alluvial fan stratigraphy can be most easily explained by a sourceward migration of the focus of sedimentation, and a consequent recession of the lithofacies belts, causing proximal fan deposits to be overlain by younger fan deposits of a progressively more distal aspect.

As suggested by Steel (1975) this implies a general lowering of relief within the fans drainage basin, and the sequence may represent the sediment yield from a drainage basin during a cycle of erosion (Schumm 1977) following the initiation of relief.

The formation of palaeovalleys with fining-upward fills records that a phase of alluvial erosion and incision preceded an aggradational phase. The local sourcing of clasts and paleocurrents (Fig. 3.33) suggests in the case of the Peroure fan that they were derived from Upper Jurassic sequences of the underlying, Peroure anticline (see Fig 1.2 (iii)), indicating that it was a positive structure during this period. At Esparron (la Tour palaeovalley) the fans were sourced from Lower Cretaceous lithologies to the south-west (Fig. 3.33), suggesting activity on the Clamensane (Durance) fault which lies some 3km in that direction (Fig.1.2 & 2.12).

3.9.4 **Comparison of the Vancon, Peroure and La Tour Fan Systems.**

A number of principle differences are recognised between the 'mass-flow dominated' alluvial fan systems of Vancon and those filling the La Tour and Peroure paleovalleys.

1) a scale difference- the northern, Peroure and La Tour 'fans' are smaller than those of the Vancon fan system, with the La Tour 'fan' having a width of less than 3 kms as a consequence of being confined within a small palaeovalley.

2) the northern palaeovalley deposits are of limited thickness whilst those of the Vancon system achieve thicknesses of up to 700m. This thickness difference reflects prolonged tectonic uplift with consequent renewal of source area relief and clastic supply in the case of the Vancon fans, whilst to the north, a single phase of tectonic uplift is inferred.

3) the northern 'fans' show a greater predominance of mass flow and sheet flood facies, and the absence of channels in the distal section of the system. Cohesive debris flow deposits extended into the playa environment, beyond the limitations of the small and ephemeral streams of the fan.

Gravity flows in the Vancon fan however are on a larger scale with huge olistolithic blocks and associated olistostromal sheets developed. In contrast to the northern 'fans', small braid channels and sheet sandstone lobes, as opposed to debris flows, fed into a palustrine to lacustrine environment.

Small, mass flow dominated, alluvial fans similar to the Peroure and La Tour 'fans', and which interfinger distally with a playa environment have been described by Denny (1967), Hooke (1967) and Bull (1972) from the arid regions of California.

In contrast 'Vancon type' alluvial fan systems which show a proximal to distal trend from massflow/sheetflood processes to braided channel and overbank processes have been described by Steel (1974) and Cabrera *et al.*, (1985). Steel (*op cit.*) envisaged the more distal braided stream deposits to be sourced from local drainage basins on the fans surface which reworked the more proximal sheetflood and mudflow deposits. In contrast Cabrera *et al.*, (1985) suggested that the braided streams were sourced beyond the fan, and issued from the fan apex into a proximal channel system of low sinuosity braided streams depositing sheet conglomerates, and passing distally into ribbon braided channels isolated within overbank fines.

3.10 Stream Dominated Terminal Alluvial Fan System - The Esclangon System

Deposits of the Esclangon system outcrop as a series of overthrust outliers in the north of the study area (Fig. 3.1). They unconformably overlie an upper Jurassic - lower Cretaceous substrate, except where locally underlain by earlier Oligocene, palaeovalley fills.

3.10.1 Petrography.

The sandstones of the Esclangon system are predominantly pale grey in colour, with channel sandstones being poorly to moderately well sorted, granular sandstones. In thin section grains are sub-rounded, with 70-80% being quartz (of which 80% is strained unimodal quartz and the remainder polycrystalline strained quartz). The remainder of the grains are predominantly lithic fragments of limestone (20%) and quartz arenite. Chert and glauconite grains are ubiquitous and may form some 5%. Muscovite, plagioclase and microcline feldspar form a minor constituent (<5%) of the sandstone.

Heavy mineral analysis of the sandstones by Planchon (1959) showed a mature heavy mineral assemblage in which zircon and garnet were abundant, with apatite, epidote, chloritoid, anatase, rutile, tourmaline and hornblende also present.

Conglomerates of the system have well rounded and sorted clasts which are predominantly of Mesozoic limestones (micritic and biomicritic), vein quartz, and chert (sub-equal abundance). Clasts of pale grey, mature quartz arenites (Triassic) and reddish brown, quartz arenites (Permian) are also common, with occasional Eocene limestone clasts (Calcaire Nummulitique).

Interpretation - Provenance.

Palaeocurrent analysis of the Esclangon system (see succeeding sections) show that it was derived from the alpine thrust belt to the east, the most proximal zones of which are presently occupied by Mesozoic carbonates and quartz-rich sandstones of the Tertiary, Gres d'Annot basin (see Fig. 3.60). The petrography of the Gres d'Annot basin fill (northern basins) is very similar to that of the Esclangon system, having a mature heavy mineral assemblage (characterised by zircon, tourmaline and rutile) and a texturally immature light component (dominated by quartz with subordinate feldspar (10-50%), and abundant Permo-Triassic lithic clasts (Stanley 1965).

The maturity of the heavy mineral assemblage of the Esclangon sandstones indicates that they were not being directly supplied from the igneous and metamorphic basement of the internal and external zones but rather that they were derived from reworking of the Gres d'Annot sandstones. The principal differences between the Esclangon and Gres d'Annot sandstones are: i) the far lower proportion of feldspar in the Esclangon sandstones (1-3% as opposed to 10-50%) - interpreted to reflect a further erosional cycle ii) the lower proportion of Permo-Triassic clasts, but significantly higher proportion of Mesozoic limestone clasts in the Esclangon sandstones.

3.10.2 Facies Associations

Four distinct facies associations are recognised within the system, namely:

1) Alluvial plain facies association.
2) Distal alluvial plain facies association.
3) Distal terminal fan facies associat ion.
4) Lacustrine facies association.

3.10.2.1 Alluvial plain facies association.

This association comprises 10-30m thick, erosive based, fining upward sequences which are vertically stacked, and involve the vertical passage from a coarse member channelised sandstone body into an association of reddish, fine member deposits (Fig. 3.39).

The coarse members are single and multistorey channel bodies which range in thickness from some 4-10m, with storeys being 3 - 6m thick. The bodies external geometry have a ribbon (type 2a, 2b and 2c) (Fig. 3.40), sheet (multistorey type 1b) (Fig. 3.42), or rarely a tabular (type 3a) (Fig. 3.41) form.

The channel fills show a fining upward trend dominated by trough cross-stratification. Channel bases are often lagged by well rounded, quartz, chert, and limestone, cobble-pebble clasts, and rarely by sub-angular oncolitic limestone clasts. The basal section of the channel fill comprises trough cross-stratified sets of granular to pebbly sandstone ranging from 0.20 - 0.70m set height and, forming cosets of 0.8 - 4.5m thickness. Vertically, sets decrease in height and grain size passing into small scale trough cross-stratified and current ripple laminated, medium grained sandstones.

A few of the thicker channel storeys have a gravel dominated basal section up to 2.0m thick (Fig. 3.43) The gravel typically forms 10-30cm thick, tabular sheets or shallow troughs, of well sorted imbricated clasts (Gm facies). More rarely scour filled, cross- stratified sets of gravel of 1.6 - 0.7m thickness develop. The foresets are of 10-30 cm thickness and tangential, dipping at 14-20 degrees, and show a normal grading from cobble - pebble grade to coarse sandstone. These sets are traceable up to 2m parallel to flow before passing into sheet gravels. Thin, erosional lenses of trough cross stratified and planar laminated granular sandstone are interbedded within the gravel facies.

The upper part of the channel fill comprises thinly interbedded sandstone and red mudstone-siltstone bundles, which are heavily bioturbated. The siltstone-mudstone intervals are red (10R4/6) with a grey mottle and may contain small calcrete glaebules. The glaebules may also be reworked into thin basal lags to the sandstone strata.

These fining-upward channel fills often show the development of large scale sets of inclined bedding, dipping at 12-18° into the channel, perpendicular to palaeoflow (Fig. 3.40-3.41). The inclined bedding shows variations in its form between two end members. The first has a massive, basal section of a few metres thickness of gravel and or, large scale cross-stratified sandstone. The inclined bedding is absent from this basal section, being confined to the finer grained, upper section of the channel fill. The second type develops in finer grained channel sequences of 4-5.5m thickness, with inclined bedding filling almost the entire thickness of the channel, extending to within some 1m of the base of the channel where trough cross -stratified sandstone is developed.

The inclined bedding is laterally interrupted by internal erosion surfaces, having a low angle convex upward form and which erosionally truncate the upper part of the bedding.

The channelised sandstones pass either abruptly or, gradually into a fine member interval comprising small ribbon channels (<2m), sheet sandstones and hetrogeneous mudstones. The fine member may show a general fining upward with a progressive decrease in grainsize and an increase in the proportion of mudstone and the degree of pedogenesis (Fig. 3.39).

The small ribbon channels have a single storey form, and are 0.9 - 2.0m thick, with widths of 5-15m. They show a simple fining upward fill, with a basal section of trough cross-stratified coarse to medium sandstone, and an upper section of thin sandstone beds showing ripple lamination. The upper section is heavily bioturbated, and pedogenically mottled such that sedimentary structures are commonly unrecognisable. These channels are typically isolated within mudstone beds, or develop within a unit of sheet sandstones, and rarely develop adjacent to large scale channel bodies (as detailed above).

The sheet sandstones are graded with sharp, planar bases and commonly show current ripple lamination (including ripple drift lamination). The thicker sandstones commonly show a trough cross stratified (5-25cm) basal section. Typically however the beds are structureless as a consequence of heavy bioturbation and pedogenic reworking. Bioturbation takes the form of two main burrow types (Fig. 3.47). The largest comprises cylindrical burrows 0.4 - 1.8cm diameter and 5 - 15cm long which may show vertical to horizontal orientation, and a simple Y-shaped branching. The burrow walls are smooth or irregular, and internally burrows show U-shaped meniscate laminae. The second burrow type is small, of 1-2mm diameter, and primarily horizontally inclined and also have a Y-shaped branching. These sandstones form dm-m scale (~1-3m) tabular units of heterogeneous, sandstone dominated units of thinly interbedded sandstone, siltsone and mudstone which have a fining upward or symmetrical coarsening-fining upward trend (3.46). They may often be traced back into the top of the large channelised sandstone bodies from which they extend as lateral wings for several hundreds of metres.

The fine member mudstones are a reddish brown with a weak grey mottling and with small, isolate carbonate nodules (Stage 1 nodules) occasionally developed.

These coarse to fine member fining-upward sequences are well exposed in the Ravine du Rousset, Esclangon (Fig. 3.45) where they are seen to be generated by the lateral migration of channel 'domains' (Allen 1978). The succession at this point comprise some 25 sandstone channel bodies isolated within fine member deposits. The channels are predominantly multistorey ribbon bodies, (tabular sheet geometry in flow parallel sections) which form laterally offset 'domains'. Three 'domains' are recognised within the exposure which show a progressive, north-westward directed, vertical and lateral, offset. The domains, which contain between 5 - 9 channel bodies, have apparent widths of the order of 200 - 300m and a vertical height of some 15 - 30m. Large scale inclined bedding within the channels, indicates that channel migration was predominantly in the same sense as the domain offset, but opposed sets of inclined bedding are developed (channel body 4). The termination of a domain typically involves an abrupt, south-eastward translation of the channels to a new position from which another north-westward domain is generated. (see channel bodies 5-7 of domains 1-2). A more progressive offset stacking may be suggested to terminate one domain (Channels 1-3).

Interpretation.

These coarse member - fine member sequences are interpreted to record deposition in mixed-load rivers and overbank sedimentation.

The channel fill deposits may be interpreted in terms of the processes of lateral deposition in streams with sinuous thalwegs

(Allen 1970). Support for this is found in the abundant large scale inclined bedding which shows all of the characteristics of lateral accretion bedding (epsilon cross stratification Allen (1963). The variations in grain size in this accretion bedding are interpreted to reflect fluctuating discharge conditions with the coarser sandstone deposited during normal flood events and fine mudstone and siltstone during low river stages (Elliott 1976). The development of erosive surfaces within this accretion bedding provides further support for fluctuating discharge conditions and probably developed by erosion of the point bar during exceptional flood events (see Elliott 1976).

Puigdefabregas & Van Vliet (1978) recognised similar bedding styles and interpreted those point bar sequences with a 'massive' lower section to represent channels in which accretion was discontinuous only in the upper part of the point bar, in the zone between bankfull and seasonal low water levels. In contrast, point bar sequences in which accretion bedding, marked by mud and silt drapes, could be traced almost to the base of the bodies, represented the deposits of channels which alternated (seasonally?) wet and dry. This latter type of point bar sequence is finer grained than most found in the study area, as silt drapes typically extend down only to within ~1.5m of the base of the channels and indicating that the discharge in the river channels of the Esclangon System was primarily perennial.

The development of imbricated stratified gravels in the basal section of channel bodies may be interpreted to result from bedload deposition at the upstream end of channel riffles (riffle head) or on the adjacent point bar heads (Bluck 1971). The downstream migration of riffles with steep lee faces into scour pools may generate the planar-trough stratified gravel sets (Bluck 1971).

The single storey ribbon bodies reflect minimal channel migration, low channel sinuosity, and a vertical accretion style of channel fill (Nami & Leeder 1978), with multistorey (2b) fill development indicating that these channels were a more permanent feature and built up by successive channel scour episodes (Atkinson 1983). The similar nature of the fill of the multistorey and multilateral ribbon bodies, and the co-existence of these channel bodies, indicates that discharge conditions within the channels were essentially the same, and that their variable external geometry was controlled by factors other than discharge regime. As discussed by Friend *et al.* (1979), the development of ribbon as opposed to tabular channels, and multistorey as opposed to single storey ribbon channels is promoted by vertical movements in the relative base level of the system and this may explain the variation in channel forms. The finer grained nature of the type 3a channel fill sequences however, suggests a primary discharge control.

The reddish colour of the fine member interval indicates that interchannel areas of the alluvial plain were well drained, with the poor development of calcrete carbonate nodules suggesting that sedimentation rates were relativly high.

The small ribbon channels within the fine member interval represent small channelised flows which laterally co-existed with the main river channel. They may be crevasse channels (Collinson 1978, Stear 1983), but their coarse grained fill is more typical of chute cut-off channels (Collinson 1986). Their general lack of development adjacent to the larger channels suggests that they may be small distributary channels which co-existed with the larger channels.

The graded sandstone beds of the fine member reflect deposition from a waning unconfined sheet flow, which uniformly scoured the substrate. The lateral passage of the sheet sandstones into major channel bodies clearly indicates that they are the deposits of flood events which overtopped channels and flowed out into interchannel areas, and they are interpreted as overbank, levee or crevasse splay deposits.

The intense bioturbation exhibited by the sandstone beds indicates that relatively long periods passed between depositional events. The larger meniscate, and smaller, smooth walled burrows are similar in appearance to the burrows described by Stanley & Fagestrom (1974), and Atkinson (1983) and interpreted to have been produced by heterocerid (Staphylinoidea), and rove beetles, and by tiger-beetle (Cicindelidae) larvae.

The generation of offset channel domains at Esclangon is considered to be the response of the river channels to a tectonic slope on the floodplain perpendicular to the channel (Bridges &

Leeder 1979, Leeder & Gawthorpe 1987, Alexander & Leeder 1987). Progressive channel shifting through meander bend migration and minor (short distance) avulsions, was interrupted by major avulsion events. These are interpreted to be the result of down-faulting on the basin margin causing the river to shift down the tectonic tilt superimposed on the floodplain toward the fault. Between active fault events, periods of quiesence were marked by progressive 'offlap' away from the downfaulted zone and the generation of offset domains. The channels preferential migration direction in these domains is interpreted to reflect that the sedimentation rate progressively compensated for the tectonic tilt and filled the tectonically generated low.

3.10.2.2 **Distal alluvial plain facies association**

This association is similar in its sequential organisation to the 'alluvial plain facies association', but importantly differs in the scale of the channel bodies and in the pedogenic characteristics of the fine members.

The facies association shows fining upward sequences passing from a coarse member, channel body into a mudstone dominated fine member (Fig. 3.48). The channel bodies are single and multistorey ribbons (type 2a, b, and c), having a medium to coarse grained sandstone fill which is devoid of gravel grade material. Channel bodies attain thicknesses of 4m, with storeys being up to 2.3m thick. Channels show a simple fining upward fill from a trough cross-stratified basal section into a heterolithic upper section. Large scale inclined bedding, dipping perpendicular to palaeoflow is developed in one of the channel fills.

The overlying mudstone dominated fine member shows a vertical organisation. Immediately above the channel, it comprises mudstones and dm-m thick sheet sandstones, which fine upward into an association of limestones and mudstones. In more detail, the fine member mudstone is a pale to greenish grey (N5 - 5G4/1), with a yellow brown (10YR6/6-10YR7/4) mottle. Thin layers of black, oragnic rich mudstone may develop within this mudstone. At the base of the fine member, carbonate nodules are present in the mudstone taking the form of isolate glaebules, or more rarely, tubular horizons (Stage 1-2 nodules).

In the upper part of the sequence, massive micritic carbonate horizons, often with a nodular fabric and with *Planorbis* gastropods, develop transitionally above these isolate nodular and tubular horizons (Stage 3-4 calcretes). The upper level of these beds may have a horizontally laminated fabric. Massive beds of micritic limestone up to 2m thick, with no nodular overprint may also be developed within the mudststone. In thin sections these massive limestones have a micritic fabric with patches of peloidal texture. Thin layers of granular limestone, with micrite coated intraclasts are developed within the micrite. Nodules in the limestones are defined by curved and circular fractures filled with microspar (Fig. 3.49). The horizontal lamination at the top of the limestones is made up of alternating layers of micrite, and microspar filled crenulate filaments.

Thin cm scale beds of gypsum are rarely developed within the greenish grey mudstones of the association. The gypsum takes the form of horizontal layers of vertical to sub-vertical, blade shaped prismatic crystals (selenite) of up to 3cm length enclosed within a mudstone matrix.

Interpretation

This association is interpreted to reflect the avulsive behaviour of channels in a distal alluvial plain setting.

The channel bodies with their ribbon form, fining upward fill, and inclined accretion bedding are interpreted to be the fills of sinuous, mixed-load rivers.

The predominantly grey colour and mottling of the fine-member mudstones is characteristic of hydromorphic gley soils formed under high, but fluctuating groundwater conditions in poorly drained, low lying areas. The yellow-reddish brown mottling is given by oxidised iron compounds introduced along cracks or roots into the reduced mudstone (Buurman 1980). Reduced conditions are also attested to by the preservation of organic rich mudstones.

The nodular carbonate horizons are characteristic of calcrete deposits, with the isolated-tubular glaebules analagous to type 1-2 youthful calcrete profiles, and the massive nodular horizons to type 3-4 mature calcrete profiles as classified by Reeves (1970), Steel (1974), and Leeder (1975). These calcretes were probably precipitated through the evaporation of carbonate-rich groundwaters (*per ascenscum* model of Goudie (1973), see also

Atkinson 1983).

The beds of massive micritic limestones are interpreted to be the deposits of shallow, and ephemeral lakes. The gravelly limestone layers present are similar to the palustrine (paludal) 'crumbly limestone' of Freytet (1973), with the intraclasts interpreted to have been mechanically reworked from the semi-lithified lacustrine limestone. Agitation and rolling around of the clasts prior to deposition produced their micritic coatings. The finely laminated bands within the limestones are algal crusts with algal filaments being replaced by microspar. The algal filaments are envisaged to trap micritic material and build up the laminate layering. These low energy deposits represent low energy periods of deposition interspersed with periods of vigorous current activity when the granular intraclast layers were generated. Similar algal stromatolite layers have been described by Nickel (1983) from ephemeral lakes in an alluvial fan (interchannel) setting.

Gypsum precipitation occurs early during evaporitic brine evolution (Allen & Collinson 1986), and as detailed by Truc (1984) the blade-like crystals grow just beneath the sediment water interface in muds with saturated interstitial waters. Similar gypsum and mudstone deposits commonly develop in saline playa mudflats but with a greater abundance than preserved in this association (Handford 1982).

In summary the facies association comprises the deposits of mixed-load channels and poorly drained, interchannel, mudflat or alluvial plain environments. Periodically fresh water, ephemeral lacustrine and palustrine environments were developed, and rarely evaporitic mudflats developed.

3.10.3 Distal terminal fan facies association

This association comprises an alternation of sandstone and heterogeneous mudstones dominate units on a dm-m scale, typically of the order of 1-4m (although they be of up to some 10m) (Fig.s.. 3.51 - 3.53).

The sandstone dominated units comprise thinly interbedded sandstones, siltstones and mudstones, and are tabular and traceable for hundreds of metres across exposures with no associated channel bodies. The sandstone beds have planar, erosive bases with groove casts and intraformational siltstone pebble lags, and range in thickness from cm scale to 50cm. They comprise well sorted sandstones which show a normal grading into reddish siltstone - silty mudstone. The thicker sandstones commonly have a trough cross-stratified (5-25cm) basal section overlain by ripple lamination, with the thinner sandstones showing a planar to ripple lamination sequence or simply ripple lamination (including ripple drift lamination). The beds are commonly structureless as a consequence of heavy bioturbation and pedogenic reworking. Bioturbation takes the form of two main burrows as detailed in the fine member deposits of the alluvial plain association.

The sheet sandstone units may show a coarsening upward, fining upward, or a symmetrical coarsening-fining upward trend. Small single storey, sandstone ribbon channels of dm scale thickness and m scale width (<10m) may be isolated within the sandstone units, typically in their coarser parts.

Occasionally larger (<2.5m thick), single storey sandstone ribbon channels are interbedded within the association (Fig. 3.52). These show a simple fining-upward fill passing from trough cross-stratified sandstone into ripple laminated siltstone and mudstone. Inclined bedding developed within the fills may extend down onto the base of the channel.

The mudstones are reddish brown, often with a weak grey mottle and may contain small calcrete glaebules. Rarely, the tops of sandstones units are the sites of the development of a horizon of tubular carbonate nodule beds, or a thin micritic limestone bed with a nodular overprint and a freshwater gastropod fauna.

Interpretation

This association with its small cyclic sequences of sheet sandstone and mudstone facies is similar to that interpreted as developing in the distal sections of terminal alluvial fan systems (Hubert & Hyde 1982). By comparison, the sheet sandstone facies are interpreted to have been deposited from flows issu ing from the mouths of active channels in the distal sections of a terminal alluvial fan.

The predominantly reddish-brown colour of the mudstones and their grey mottling are characteristic of pseudo-gley processes, which operate in flat, low lying areas with impeded drainage (Buurman 1980). The discrete carbonate nodules are similar to those developed in immature calcrete profiles (Reeves 1970, Steel 1974, Leeder 1975) and reflect that sedimentation rates which remained relatively high even in the lower energy periods of deposition.

In contrast the heavy bioturbation by beetles and beetle larvae (see alluvial plain facies interpretation) show that sufficient periods of time elapsed between sedimentation events for the disruption of bedding. The thin micritic limestones are interpreted as the deposits of ephemeral bodies of fresh water, which were subjected to pedogenic overprinting.

The cyclic coarsening and fining-upward sequences of the association are attributed to channel avulsion on the fan. Channel avulsion to a new site initiated a coarsening upward, mudstone to sheet sandstone sequence with the fining-upward section reflecting a progressive filling and abandonment of the channel. Periodic progradation and incursion of the distal sections of feeder channels are considered to be recorded by the presence of small ribbons in the sequences. The development of inclined bedding which fills the whole of these channels is similar to that detailed by Puigdefabregas and Van Vliet (1978) and interpreted to develop in the meandering channels of ephemeral rivers.

3.1 0.2.4 **Lacustrine facies association.**

This association consists of interbedded tabular limestones and calcareous mudstones (Fig. 3.55 - 3.56).

The limestones form laterally persistent strata 0.2-1.6m thick, and are a pale grey (N5-N6) colour, whilst the mudstones are fissile and dark grey (N6). The limestones have a homogeneous micritic texture with a faint horizontal lamination. Thin (0.3-1.0 cm) sharp based, and graded calcareous siltstone-micrite beds, with upper laminae rich in plant-debris, are randomly interbedded within the limestone.

Petrographically the limestones are micritic and devoid of clastic grains, with the micrite commonly showing a vague peloidal and clotted texture. Admixed within the micrite are recrystallised remnants of *Planorbid* gastropods, ostracods and Characea (Fig. 3.57). The graded silt beds are ~10- 20% quartz grains admixed within micrite intraclasts and a micrite matrix.

Also randomly interbedded within the limestones are thin, monospecific coquinas of whole *Planorbid* gastropods. These coquinas are the sites of development of transparent brownish (5YR3/2) chert nodules. The nodules form discontinous bands parallel to bedding, having a thickness of 0.5–6.0 cm, lengths of cm - m scale (~3m), and forming horizons of nodules which extend for several hundreds of metres. In thin section the gastropod shells within the chert nodules may be seen to be replaced by silica, and to be enclosed within a mass of microcrystalline chert which is replacing the micritic limestone (Fig. 3.58).

Interpretation.

The features of these limestones and marls and the fauna are characteristic of formation in a shallow, freshwater lacustrine environment, with no evidence of any subaerial exposure.

The thin graded silt layers record the periodic incursion of externally derived quartz clastic material deposited out of suspension from waning turbulent flows.

The precipitation of chert in a predominatly carbonate environment has been studied by Peters & von de Birch (1965), and Zijlstra (1987). The silica is believed to have been sourced from the tests of diatoms, detrital quartz and clay minerals and to have been dissolved under relatively high pH conditions developed during periods (seasonal?) of extremely active bio-synthesis. The silica dissolves in the upper oxidising zone of the sediment column producing monomeric silica complexes. The silica is precipitated in the immediately underlying reduction zone where the pH is lowered to some 6.5 in layers of decaying organic matter. The silica preferentially polymerises in pore spaces where the hydrogen ion concentration is lowest, these sites being created by hydrogen-removing bacteria which are involved in the breakdown of the organic material.

The development of chert nodules along the gastropod coquinas appears to be a function of the locally reduced pH conditions produced by the anaerobic decay of the gastropods organic matter. The gastropod coquinas may themselves have been produced by periods of oxygen deficiency and high pH conditions in the hypolimnion zone as a consequence of periodic eutrophication of the lake.

The inferred fluctuations in pH conditions within the lacustrine body are characteristic of those which occur in hydrologically closed lakes (Allen & Collinson 1986)

Nickel (1982) described analogous deposits from alluvial sequences in the Pyrenees. They comprised fresh water micritic limestones and marls with chert, and were interpreted as 'offshore' lacustrine deposits having developed in a distal alluvial plain and/or alluvial fan-edge position.

3.11 Studied Sections.

Deposits of the Esclangon fluvial system are well exposed in a number of outliers along the northern margin of the Durancian basin, namely at Esparron, Esclangon, and St.Geniez (see Fig. 3.1). Representative logs of the exposed successions at these localities are detailed in Fig. 3.59) and described and interpreted in this section. The lateral variability in facies associations between these localities is subsequently discussed in section 3.12.

3.11.1 The Esparron Section

Sequences of the Esclangon fluvial system are spectacularly exposed in the cliffs beneath the Crete de Charene (Fig 3.54), 9 km to the west of the hamlet of Esparron-la-Battie (903,3330, Seyne 5-6). In the NE of the exposure the deposits conformably overlie a thin sequence of coarse alluvial deposits of the La Tour palaeovalley system which form the base of the Molasse Range Formation (See Fig. 3.1). In the SE of the section they unconformably overlie lower Cretaceous and Upper Jurassic lithologies. The top of the succession is truncated by the Digne thrust sheet.

The succession at Esparron has a relatively simple fining upward mega-sequence over some 180m, which is represented in Fig. (3.59).

The basal 50m of the mega-sequence comprises the alluvial plain facies association. Channel bodies, of 3-6m height form between 25-40% of this basal level and are arranged in repeated fining-upward sequences of 8-20m thickness. Channel bodies have a tabular external geometry, with well developed lateral accretion bedding and are classified as type 3b tabular bodies.

The basal part of the mega- sequence passes transitionally upward, into a 100m thick succession of sheet sandstone and facies of the distal fan facies association which form the upper part of the exposure. Sheet sandstones form about 60% of the deposits. They are traceable across outcrop for several hundreds of metres forming persistent parallel sided tabular strata (Fig. 3.53). Channels are rarely developed (ca 5%) within this upper section, but occasionaly are present as isolated ribbon bodies.

Palaeocurrent analysis of the succession indicates that the mean palaeoflow direction of the palaeochannels was to the WSW (254^{0}N), with sheet sandstones having an essentially parallel mean palaeoflow direction of WSW (251^{0}N).

Interpretation

The large-scale and laterally persistent nature of the fining upward mega-sequence is considered to reflect the establishment of mixed load alluvial plain system and its subsequent replacement by a terminal 'sand flat' system. The overall vertical decrease in grain size and channel body scale, coupled with an increase in the proportion of fine member 'sheetflood' deposits are analogous to the proximal to distal trends attributed to 'terminal fans' by Friend (1978).

3.11.2 Esclangon Section

At Esclangon the fluvial system attains a maximum exposed thickness of some 300m (Fig. 3.59). The studied section lies on the southern limb of an E-W trending anticline in a structurally complex zone (Fig. 3.1; also Fig 1.2 (iii)). The base of the succession conformably and gradationally overlies a gravel dominated fill of the Peroure palaeovalley.

The top of the succession is absent in the extreme east and west of the section (where it is tectonically truncated by the Digne and Clamensane thrust sheets respectively). However, a relatively complete sequence is developed along the the D900 road where it passes conformably into the overlying Marine Molasse Formation. Graphic logs from the section are presented in Fig. 3.59.

The succession shows a large scale, assymetric coarsening-, followed by -fining upward mega-sequence.

The coarsening-upward part comprises a relatively thin unit of some 30-40m unit of stacked mature calcrete profiles enclosed within reddish brown pseodogley mottled mudstone facies with a few, thin sheet sandstone beds. At the base of this unit the occasional thin debris flow from the underlying Peroure palaeovalley system is interbedded within the mudstone. The calcretes are of type 2-3, immature-mature profiles with well developed rhizotubular horizons in which rhizotubules attain lengths up to 2.0m.

This basal unit passes transitionally up into a central unit 160 m thick which comprises repeated fining-upward sequences of the order of tens of metres thick, of the alluvial plain facies association.

The channel bodies of these sequences typically have multistorey and multi-lateral ribbon form (type 2b,&c ribbon channels, more rarely multistorey tabular bodies are developed. Channel storeys attain thickness of up to 6.5m, typically ranging from 2-6.5m, with multistorey bodies being of up to 10m thickness. The larger channels have a gravel basal section.

This central unit passes transitionally into a poorly exposed upper unit of 50m, comprising sequences of the distal alluvial plain facies association. The top of the unit is marked by the development of a 20m thick unit of palustrine, lacustrine limestones and mudstones.

Palaeocurrent analysis of channel sandstones indicates that the fluvial system had a WSW (254^{0}N) mean palaeoflow direction.

Interpretation.

The establishment of the fluvial system at Esclangon was gradational one marked by a period of slow flood plain accretion and the development of stacked mature calcrete profiles. The oxidised nature of the overbank facies reflects the well drained nature of the flood plain during this and the overlying unit. This interpretation is supported in the basal section by the large length of the rhizotubular calcrete profiles which reflect the deep penetration required of the tap root systems to reach the ground water level.

The range of channel dimensions at any level in the channelised mudstone and red mudstone facies association is considered to reflect that a multiple channel system was present.

The coarsening-upwards of the mega-sequence with an increase in channel thickness and the development of gravel channels in the middle unit reflects progradation of the fluvial system. The smaller scale fining-upward, channel and overbank facies sequences within the mega-sequence may be attributed to channel avulsion, which are interpreted to have been in response to vertical tectonic movements on the margin of the alluvial plain (see facies association interpretation).

The vertical passage into a distal floodplain association is interpreted to reflect retrogradation of the system, or possibly, a lateral shifting of its principal area of deposition, with the vertical decrease in channel body thickness considered to reflect a downstream decrease in channel dimensions, a characteristic feature of a terminal fluvial distributary system (Friend 1978).

The transition into the overlying, shallow marine succession of the Marine Molasse formation is considered to support the former interpretation, with the fining-upward of the mega-sequence reflecting a progressive increase in the base level of the system in response to a sea level rise (the Burdigalian eustatic sea level rise). Similar large scale fining-upward trends in fluvial successions of the New Red Sandstone have been suggested by Steel (1974) to predict a marine transgression which is evidenced by conformably overlying marine Jurassic rocks.

Nichols (1984 & 1987) and Hirst & Nichols (1987) detailed a possibly analogous 'terminal distributary system', the Luna system from the Ebro Basin, N. Spain. Medial zone deposits of the system comprise sandstone channel bodies within overbank fines analogous to the central unit of the section whereas distal zone lacustrine limestone and sheet sandstone adeposits are analogous to the upper unit of the section.

3.11.3 St. Geniez Section.

Deposits of the Esclangon fluvial system attain a thickness of some 200m as exposed in the slopes of the Gorge du Vancon

(900,5/3222, Sisteron 3340) some 4km ESE of the village of St. Geniez (Fig 3.1). The succession is detailed in Fig. 3.59. The top of the section is poorly exposed and also influenced by thrust tip folding associated with the overlying Clamensane thrust sheet. Nevertheless a clear picture of a coarsening upward mega-sequence can be reconstructed from the preserved section.

The basal section of the mega-sequence comprises deposits of the **lacustrine facies association** which are laterally continous over 0.6 km. This lacustrine succession has a maximum thickness of 100m, and may be seen to progressively thin northward as the deposits onlap onto the Cretaceous 'basement' at the margins of the St. Geniez graben (see Fig. 6.7).

This basal section shows a transitional passage through interbedded sheet sandstones and palustrine limestones into deposits of the distal fan facies association, which form the upper 140m of the mega-sequence. Palaeocurrents from the coarse member channel fills from this association section have a mean 274^{o}W palaeoflow direction with individual channel fill means ranging between 216-330^{o} (SW-NW).

Interpretation.

This mega-sequence is interpreted to reflect the westward progradation and establishment of the Esclangon fluvial system, displacing a hydrologically closed, lacustrine system. The vertical passage from chert limestone and marl, to pedogenic mudstone and limestone and channelised facies is interpreted as the passage from an offshore, open lacustrine environment into an ephemeral, lake margin and distal distributary channel system. The intimate interbedding of fluvial and lacustrine facies is interpreted to record that the fluvial systems terminated in, and supplied, the lake system.

3.12 Spatial Variations of the Esclangon Fluvial System - Comparison with other fluvial systems.

Correlation of the studied sections of the Esclangon fluvial system is not possible, but given their similar facies associations, close geographical spacing, and position directly beneath the Marine Molasse formation (Fig. 3.1) they are assumed to be coeval. Channel fill and overbank facies from all the studied locations have the characteristics of a mixed load fluvial system. Vertical facies association trends in the Esclangon mega-sequence reflect that the more proximal sections of the fluvial system comprised sinuous, mixed load, distributary channels traversing a well drained alluvial plain, and that distally the channel dimensions decreased and traversed a poorly drained alluvial plain with ephemeral interchannel lakes.

The proximal-distal relationship interpreted from this mega-sequence is supported by the coeval existance of 'proximal' and 'distal' facies associations in the studied sections. Thus, whilst the more proximal fluvial channel system was established at the base of the eastern sections of Esparron and Esclangon, the more distal, lacustrine equivalents of the system were being deposited some 8 - 15 km towards the west, at St. Geniez. From the limited exposure it appears that the depocentre of the system was established in the Esclangon where the largest channels with the coarsest bedload developed.

All three sections show a relatively simple vertical trend which may be interpreted in terms of the progradational establishment, and subsequent retrogradationl retreat of the fluvial system. In Esparron and Esclangon the progradation of the system was relatively rapid, with a lack of, or only a thin, coarsening upward progradational sequence. In the more southwesterly section of St. Geniez, the development of a thick coarsening-upward mega-sequence suggests a time lag in the progradation of the system relative to Esclangon and Esparron.

The fluvial dispersal pattern of the Esclangon system, as indicated by palaeocurrents (Fig. 3.59) and supported by proximal-distal lithofacies trends, shows that the provenance area of the system lay to the west-northwest (Fig. 3.60). The probable source of the system was the Tertiary sandstones of the Gres d'Annot (see section 3.10.1). At present these sequences lie some 20km to the west, above the Digne thrust sheet and Haute-Provence duplex. Siddans (1979) considers there to have been some 13km of southwestward pre-Pleistocene displacement on the Digne thrust sheet, and Graham's (1985) structural section of the structurally higher Haute-Provence Duplex (see Fig. 1.3 (iii)) suggests that no more than 5km southwestward displacement on the frontal thrusts of this section of the thrust system. The probable source area for the Esclangon fluvial system therefore potentially lay only some 40km to the west of its present most easterly outcrops (Fig. 3.60)

By comparison with the Luna system (Nicholls 1984, Hirst & Nicholls 1987) the association of ribbon and sheet channels are envisaged to have become progressively smaller, probably through bifurcation and, as a consequence of the reduced discharge which they accommodated. Flows become poorly channelised or unconfined with sheet flow deposits becoming predominant (as at the upper section at Esparron). A shallow lacustrine system developed at the distal fringes of the fluvial system as in the Luna System. The lacustrine - fluvial boundary was transitional and probably fluctuated with variable discharge in the system as a whole as indicated by the intimate interbedding of deposits of the two systems. Whilst channels form approximately 20% of the fluvial sequence in the distal succesions of St Geniez, they form only approximately 5% of the upper terminal sand-flat association of the sheet sandstones and mudstones at Esparron. This suggests that two types of "end-member" depositional environments may fringe the terminal zone of the Esclangon fluvial system, namely a shallow lacustrine system, or a playa sand-flat system. The lacustrine system at St. Geniez lies within a structurally defined low, the St. Geniez Graben, and it may be suggested that intra-basinal faulting trapped rivers and acted as a floodwater dam to create a lake (see Alexander & Leeder (1987)).

3.13 Tectonic Controls on Terminal Systems of the Molasse Rouge Formation.

At its most fundamental, tectonic control on the continental sedimentation may be considered in terms of slopes. Tectonic activity, be it compressional, extensional or strike-slip will result in areas of active uplift and subsidence relative to base level. Tectonic slopes thus arise, which will influence all gravity driven sedimentary processes, including alluvial systems. As discussed by Alexander and Leeder (1987) rivers in an active tectonic setting will (i) tend toward the position of topographic minimum which is commonly the site of maximum subsidence (ii) often be deflected from this course toward the basinal low by an intrabasinal structure ('structural alients' of Eisbacher *et al.* 1974).

Facies distribution patterns, clast types, sandstone petrographies and palaeocurrent directions of the 'Molasse Rouge' deposits show that two spatially distinct alluvial systems coevally existed, namely the Esclangon, and Vancon, Terminal Fan Systems. The relationship between these systems and the structural framework of the Haute-Provence region is detailed in Fig. 3.60. The frontal alpine thrust sheet in this region, the Digne thrust sheet, has been restored by its maximum displacement distance of 20km (after Siddans 1979). This restored position is coincident with the position of the Carboniferous basement inlier at Barles (B) and the high angle basement fault of Sorine, suggesting that the Digne thrust footwall ramp developed at an extensional fault basement scarp (see also Fig. 2.8). The Digne thrust is known to have had a pre-Priabonian phase of movement from the thrust tip fold relationships with the fill of the Barreme Basin (Graham 1985). Therefore, any fluvial systems draining south west and into the Provence Basin would have to breach the frontal structure of the Digne thrust (emergent thrust scarp (?) or, frontal thrust tip fold (?) as well as any compressional structures associated with the out of sequence thrust system west of the Digne thrust (see Graham 1985). The Gres d'Annot and Barreme basins have been tentatively restored using Graham's section (1985) and adopting a best fit approach with respect to Siddans (1979) data.

The foreland to the west of the thrust belt is transected by a series of inherited extensional faults. The principal fault set trends NE-SW, defining a regional graben strike, and is transected at a high angle by transfer fault sets. The multidirectional nature of the fault sets creates a system of, four sided, grabens and horsts ('trapdoors' of Harding 1983) in the foreland. Reference to figure 3.60 shows that in the the north and east of the study area an 'internally sourced', mixed bed-load, fluvial system was developed. In the study area, this system has been termed the Esclangon system, whilst its temporal equivalents have been described from the Barreme basin (Evans 1985) and Faucon region (Arnaud *et al.* 1977). Palaeocurrent directions from the Esclangon system show that it flowed parallel to the strike of the extensional fault sets within the foreland. The rivers are perceived as trending towards the sites of maximum subsidence on the hanging wall blocks of the fault bound basin. This is supported by the development of offlap stratigraphy in the Esclangon terminal fan system succession at Esclangon. The Esclangon system is noted to have flowed parallel to both the principal and transfer fault sets. It is suggested that the fluvial system issu ing out into the foreland basin initially followed the regional graben strike until the point where it met a 'structural salient' in the form of a transfer fault footwall scarp, or hanging wall slope. Such features are envisaged to have trapped or diverted the axial system. For example, the fluvial system changes from a regional fault trend in the Esclangon section, to a transfer fault trend in the more distal St. Geniez section. Whether the fluvial system is 'structurally terminated', or whether it forms a true 'terminal distributary' system (Nichols 1987) dificult to ascertain.

The relationship between the fluvial system and the compressional structures to the west of the Digne thrust front is not apparent because most of the comparable age deposits have since been stripped off.

Leeder & Gawthorpe (1987) considered continental extensional basins with axial through drainage as a tectono-sedimentary facies model. The fundamental characteristics of such a system were considered to be a preferential stacking of channels towards the footwall.

Continental Eocene-Oligocene sequences of the Barreme basin record an alternation of local and internal zone sourced fluvial systems (Evans 1985), suggesting that periodic thrusting to the east of the basin diverted the internally sourced systems until such a stage that headward capture re-routed them through the basin.

In Fig. 3.60 it has been speculatively suggested that the terminal fan nature of the fluvial system may reflect deposition as a consequence of the change in gradient in transecting the Digne thrust front and issu ing into the foreland. Such a style of fluvial dispersal has been described by Hirst & Nichols (1986) and Nichols (1987).

The south western area of the basin has a markedly different fluvial system to that in the north east, comprising a series of locally sourced, alluvial fans which developed in the hanging block of the regional Forcalquier graben (Fig. 3.61). The footwall sourced fans had a limited radius of the order of 5-8kms passing westward into a lacustrine system (Cavelier et al 1986). Differential subsidence in the hanging wall of the regional graben occurred along oblique to transverse, transfer faults. As a consequence, a series of structurally differentiated fan systems developed, namely the Aix, Manosque-Forcalquier, and Vancon fan systems, whose principal dispersal axes were parallel to the E-W trending transfer faults. In the study area the Vancon fan system developed in the Vaucluse graben. The limited radius of these footwall fringing fans was apparently a consequence of their limited drainage area. Reference to Fig. 3.60 shows that deposits of the Esclagon fluvial system lay only some 10km to the east of the Durance fault on its footwall block indicating that the fulcrum of the block lay further to the west.

Summary

1) The dispersal and depositional characteristics of the deposits of the Molasse Rouge formation of the study area reflect deposition in a structurally active extensional setting in the foreland to an active thrust belt.

2) The principal role played by compressional tectonics was in generating an erosional domain from which siliciclastic sourced alpine fluvial systems flowed south westward into the subsiding foreland basin. The role of thrust load induced flexural subsidence is difficult to assess as the response of the fluvial systems was dominated by the extensional structures.

3) Within the foreland basin, the alpine fluvial systems did not flow longitudinally with respect to the alpine thrust belt. Rather their flow direction was parallel to the strike of reactivated extensional basement faults.

4) Sediment supplied to the foreland basin from the thrust belt was apparently restricted to the immediate vicinity of the thrust front by extensional graben structural sediment traps.

5) Intra-basinal terminal fan systems, locally sourced from Mesozoic carbonates, developed within grabens to the west of the immediate alpine thrust front. They preferentially developed in

the hanging-wall block to the regional NE-SW trending, Durance fault, where they accumulated as thick successions within sub-basins defined by E-W trending transfer faults.

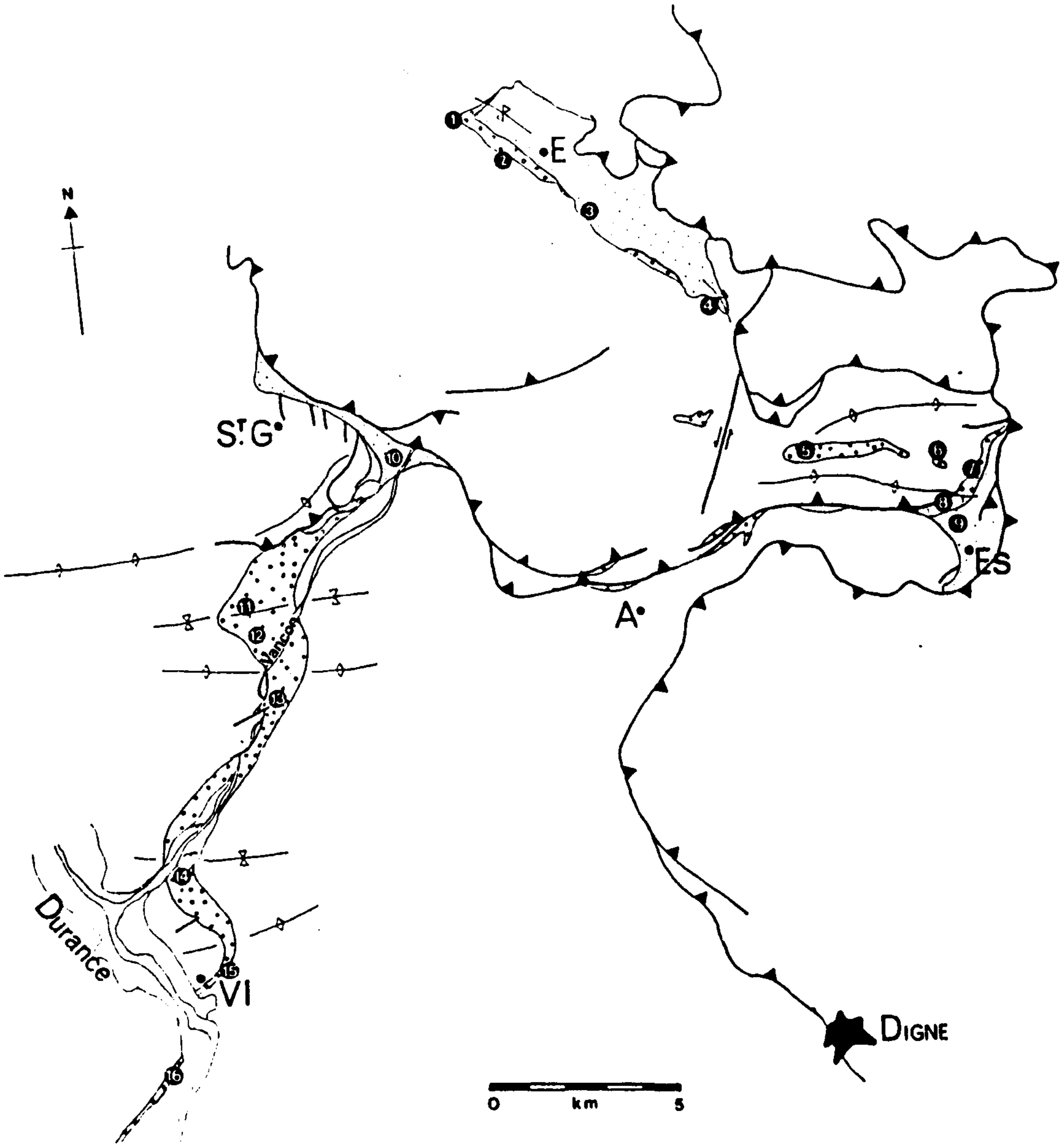


FIG 3.1 Outcrop pattern of the Molasse Rouge Formation along the northern margin of the Durancian basin showing the distribution of terminal fan systems and studied localities. E-Esclangon terminal fan system V-Vancon terminal fan system P-Peroure palaeovalley LT-La Tour palaeovalley.

Localities studied:1-le Pategue 2-la Tour 3-Maregineste 4-Maladrech 5-Aubeze 6-le Fiere 7-la Pare 8-Peroure 9-Esclangon 10-St.Geniez 11-Jonchier 12-St.Symphorien 13 Citadelle/Beadument 14-Sourribes 15-Volonne 16-Chateau Arnoux

St.G: St. Geniez Es: Esclangon E: Esparron A: Auribeau

Vl: Volonne

INTRABASINAL MESOZOIC CARBONATE SOURCED

ALPINE SILICLASTIC SOURCED

FLUVIAL SYSTEM	FAN TYPE	SOURCE	MIN-MAX THICKNESS (m).	FACIES ASSOCIATIONS	CHARACTERISTICS	CHANNEL DIMENSIONS W (m)	PALAEOFLOW
ESCLANGON TERMINAL FAN	Stream Dominated	Alpine Hinterland - Gres d'Annot basin {early foreland basin} depocentre.	180-290	Alluvial Plain	Mixed-load distributary channels, well drained floodplains.	40-60	-> W, SW
				Distal Alluvial Plain	Mixed-load distributary channels floodplains with ephemeral lakes.		
				Distal Terminal Fan	Sand sheet lobes, ephemeral channels playa.		
				Lacustrine	Hydrologicaly closed lakes.		
VANCON TERMINAL FAN	Mass Flow	Local footwall block of Durance fault	100 - 700	Proximal Fan	Massive conglomerate, with olistoliths.	10-30	-> W, SW
				Medial Fan	Braided bedload distributary channels. well-poorly drained interchannel ares.		
				Distal Fan	Ephemeral bedload distributary channels. well-poorly drained inter-/ fore-channel areas, vegetated. Palustrine fringe.		
PEROURE & LA TOUR PALAEOVALLEY FILL TERMINAL FANS.	Mass Flow	Local	0 - 70.	Proximal Fan	Massive conglomerate	5-15	-> NE & SE
				Medial Fan	Braided bedload distributary channels, debris flows, well drained interchannel area.		
				Distal Fan	Debris flows, and well drained fan fringe mudstone. Palustrine playa, valley margins.		

FIG 3.2 Terminal alluvial fan systems of the Molasse Rouge formation.

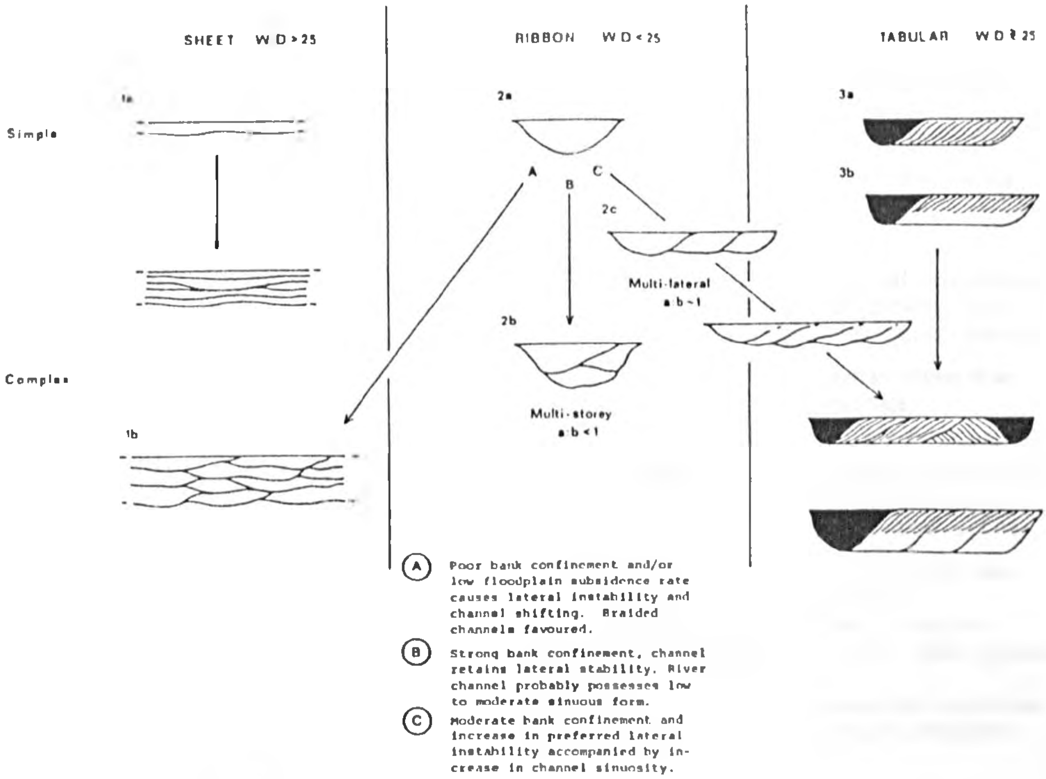


FIG 3.3 Geometric classification scheme for coarse member bodies of the Molasse Rouge Formation (from Atkinson 1983).

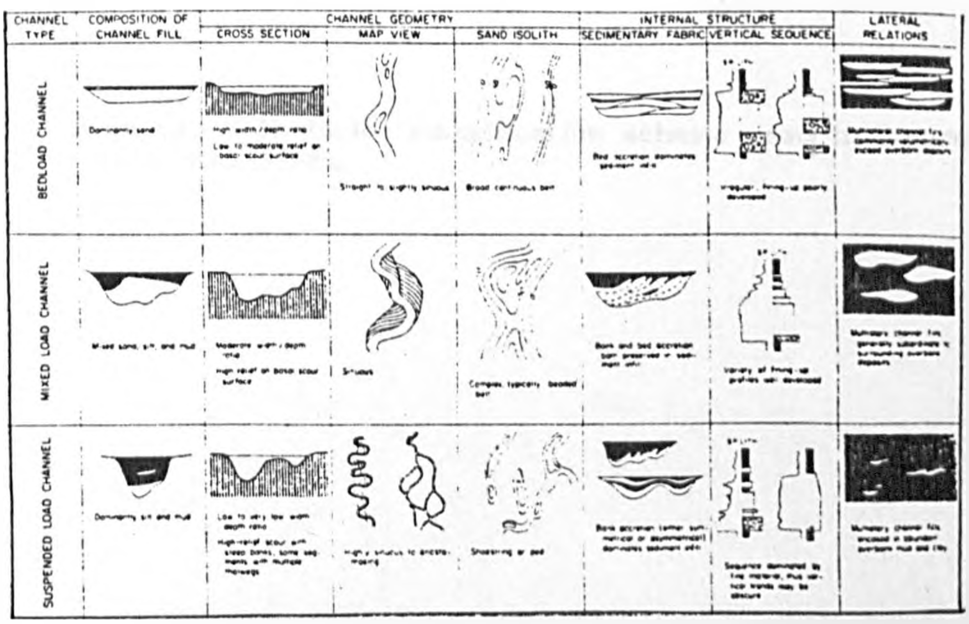
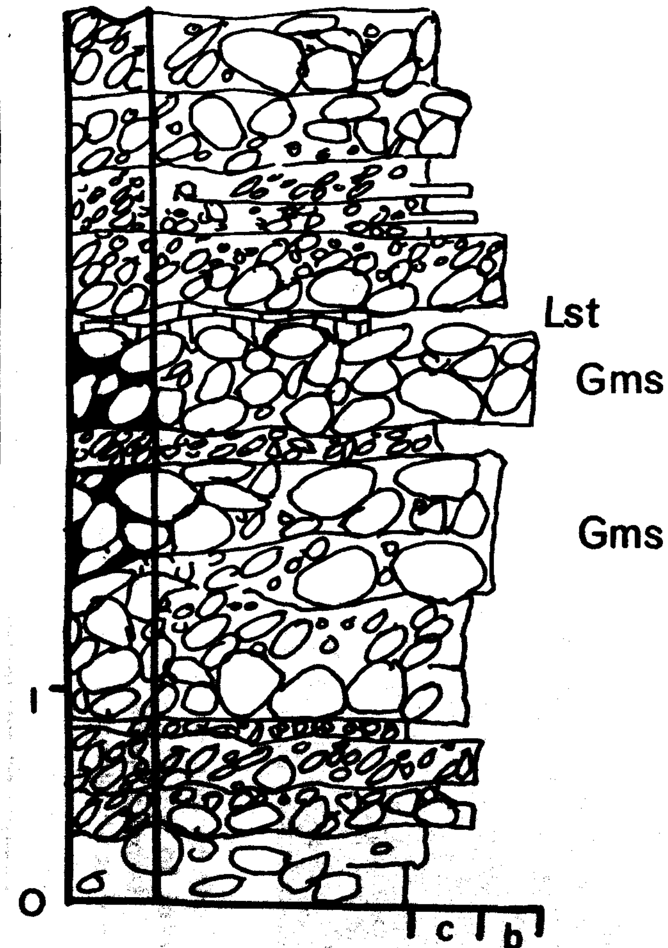


FIG 3.4 Geomorphic and sedimentary characteristics of bedload, mixed-load and suspended-load channel systems (from Galloway 1981).

Facies Code	Lithofacies	Sedimentary structures	Interpretation
Gms	massive, matrix supported gravel	none	debris flow deposits
Gm	massive or crudely bedded gravel	horizontal bedding, imbrication	longitudinal bars, lag deposits, sieve deposits
Gs	gravel, stratified	sigmoidal crossbeds	minor channel fills
Gp	gravel, stratified	planar crossbeds	linguoid bars or deltaic growths from older bar remnants
St	sand, medium to v. coarse, may be pebbly	solitary (theta) or grouped (pi) trough crossbeds	dunes (lower flow regime)
Sp	sand, medium to v. coarse, may be pebbly	solitary (alpha) or grouped (omikron) planar crossbeds	linguoid, transverse bars, sand waves (lower flow regime)
Sr	sand, very fine to coarse	ripple marks of all types	ripples (lower flow regime)
Sh	sand, very fine to very coarse, may be pebbly	horizontal lamination, parting or streaming lineation	planar bed flow (l. and u. flow regime)
Sl	sand, fine	low angle (<10°) crossbeds	scour fills, crevasse splays, antidunes
Fm	mud, silt	massive, desiccation cracks	overbank or drape deposits
P	carbonate	pedogenic features	soil

FIG 3.5 Lithofacies classification scheme used in figures (after Miall 1978).



DESCRIPTION	INTERPRETATION
Pale grey algal laminated limestone.	Period of non-clastic deposition on section of fan due to channel avulsion etc.
Poorly sorted, polymodal matrix/clast supported conglomerate, beds ungraded or with weak normal or reverse grading	Cohesive and non-cohesive sediment gravity flows Important deposits on heads of semi-arid alluvial fans.
Well sorted, clast supported imbricate sheet conglomerate, commonly normally graded.	Turbulent sheetflows, deposition from bedload. Sheetfloods issueing from downstream ends of channels below fan intersection point, shallow braided channels.

FIG 3.6 Reprentative graphic log of the Proximal Fan Facies Association.

FIG 3.7 General view of massive conglomerates of the proximal fan facies

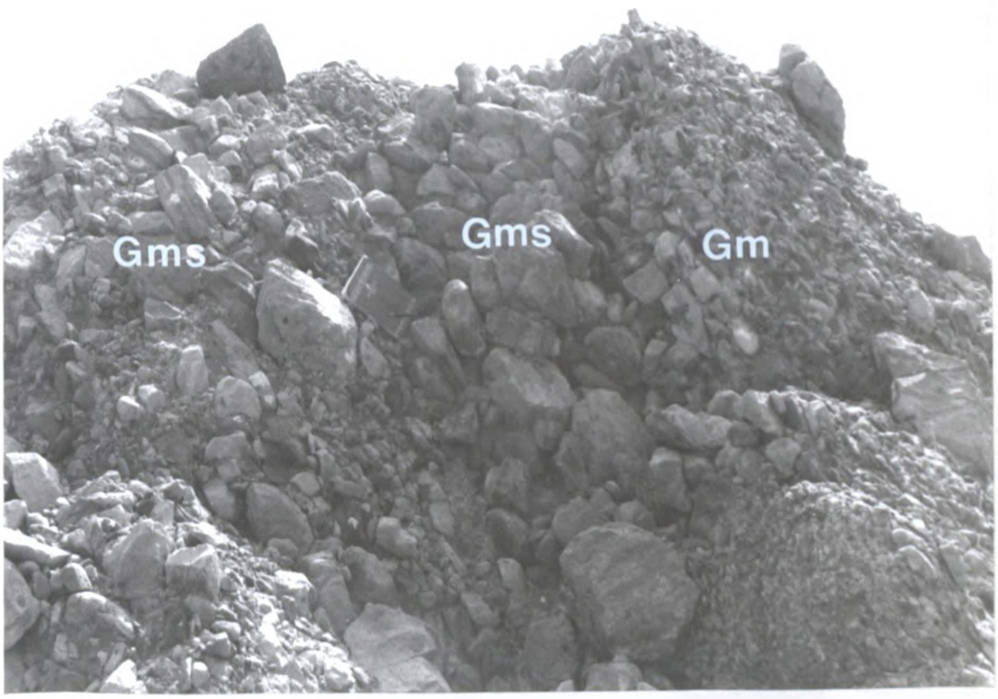


FIG 3.8 Horizontally stratified sheet conglomerates of the proximal fan facies association. Note the interbedding of inverse graded, cohesive debris flows (Gms) and normally graded, imbricate sheet flood conglomerates. (Gm) Notebook 30cm for scale.

FIG 3.9 Close up view of mud rich pebbly, siltstone matrix in a cohesive debris flow. Lens cap 5cm for scale.

FIG 3.10 Algal laminated micritic limestone within conglomerates of the proximal fan facies association. Note the irregular base to the limestone which passively infills boulder relief, and the erosionally defined upper limit of the limestone. Bedding dips shallowly to the right (~10 degrees). Lens cap 5cm for scale.

Fig 3.11 Olistolith (O) of massive Cenomanian limestone (~100m in length), erosively overlies Valanginian marls at the base of the Molasse Rouge Formation, Sourribes.

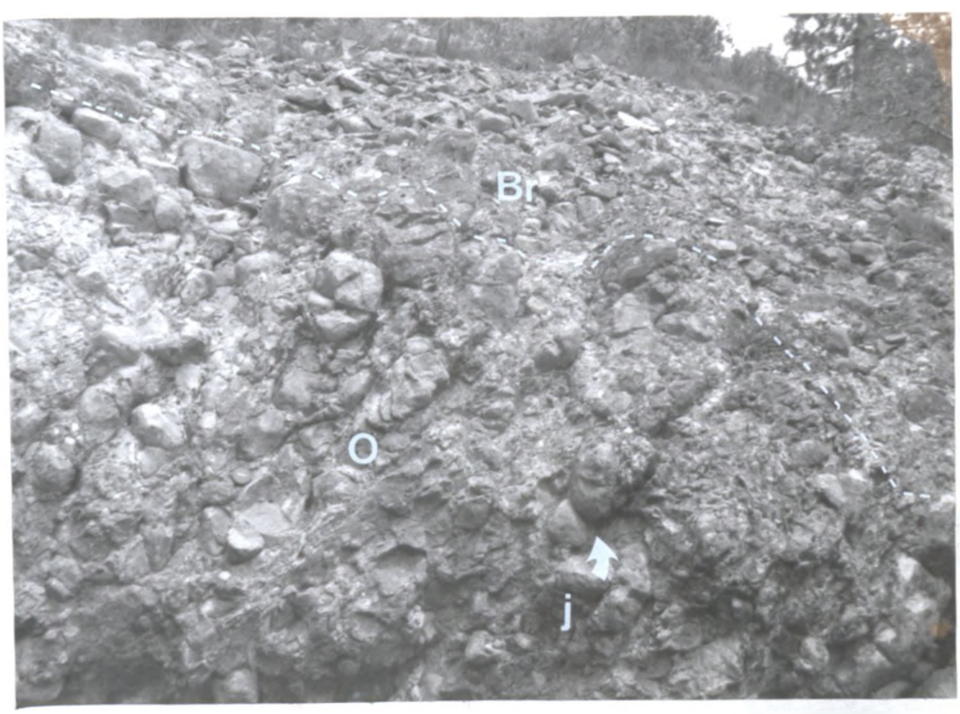
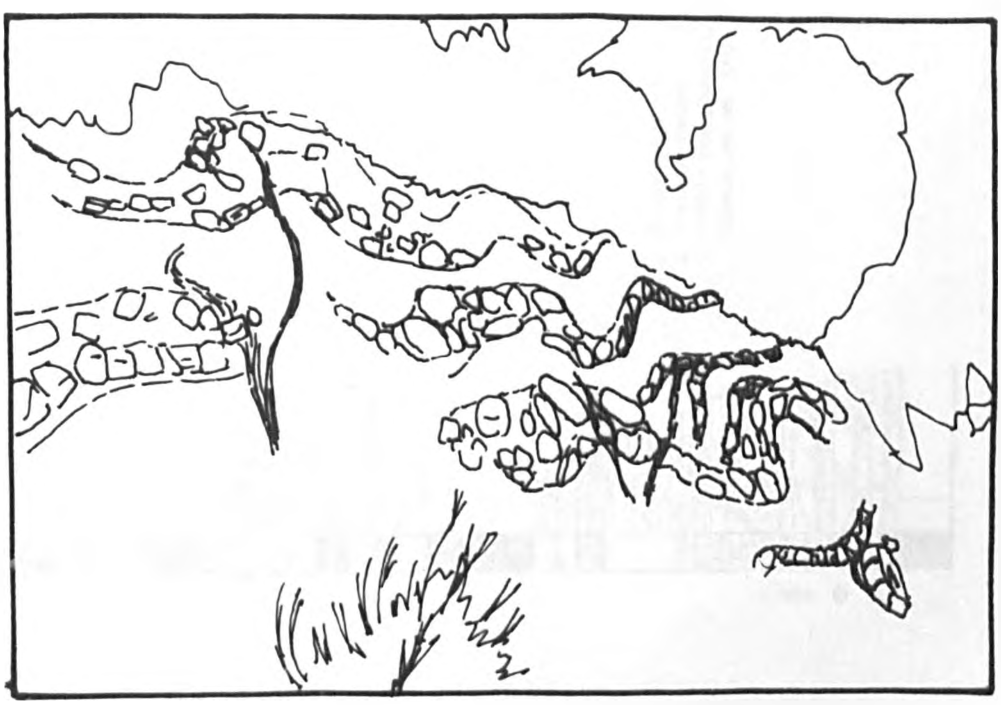
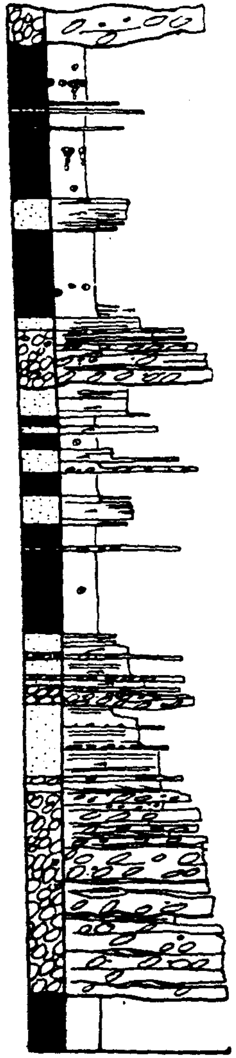


FIG 3.12 Olistolith (O) of heavily jointed (j) Cenomanian limestone overlain by a limestone breccia (Br) interpreted to have developed through the in-situ palaeo-weathering of the olistolith.

FIG 3.13 Folded bedding within Cenomanian limestone and marl olistolith, near Sourribes (Garampont). The limestone beds are divided into joint blocks which have rotated out of the bedding plane. Narrow shear zones (s) transect the bedding.





DESCRIPTION	PROCESS	INTERPRETATION
Heterolithic mudstone with thin units of graded sheet sandstones and conglomerate. Small ribbon gravel channels. Mudstones are a reddish-brown with grey mottling. Level of pedogenesis increases vertically from isolate glaebules (Stage 1) to pedotubular carbonate (Stage 2-3).	Sheet sandstones - overbank sheet flows Shallow channels with low relief bars. Pseudogley soil and calcrete formation.	Low lying interchannel area with impeded drainage, ephemeral channels.
Multistorey ribbon / sheet gravel channel. Fining-upward sequence of tabular imbricate cobble conglomerate . Planar and sigmoidal stratification. Erosional lenses sandstone.	Peak flow longitudinal bar migration. Low flow stage reworking of bars.	Major braided distributary channel.

FIG 3.14 Idealised coarsening upward sequence of the Medial Fan facies association.

FIG 3.15 Medial fan facies association, Vancon fan system. Jonchier section. (i) Multistorey ribbon channel conglomerate bodies and fine member gley mudstones form repeated fining upward sequences.

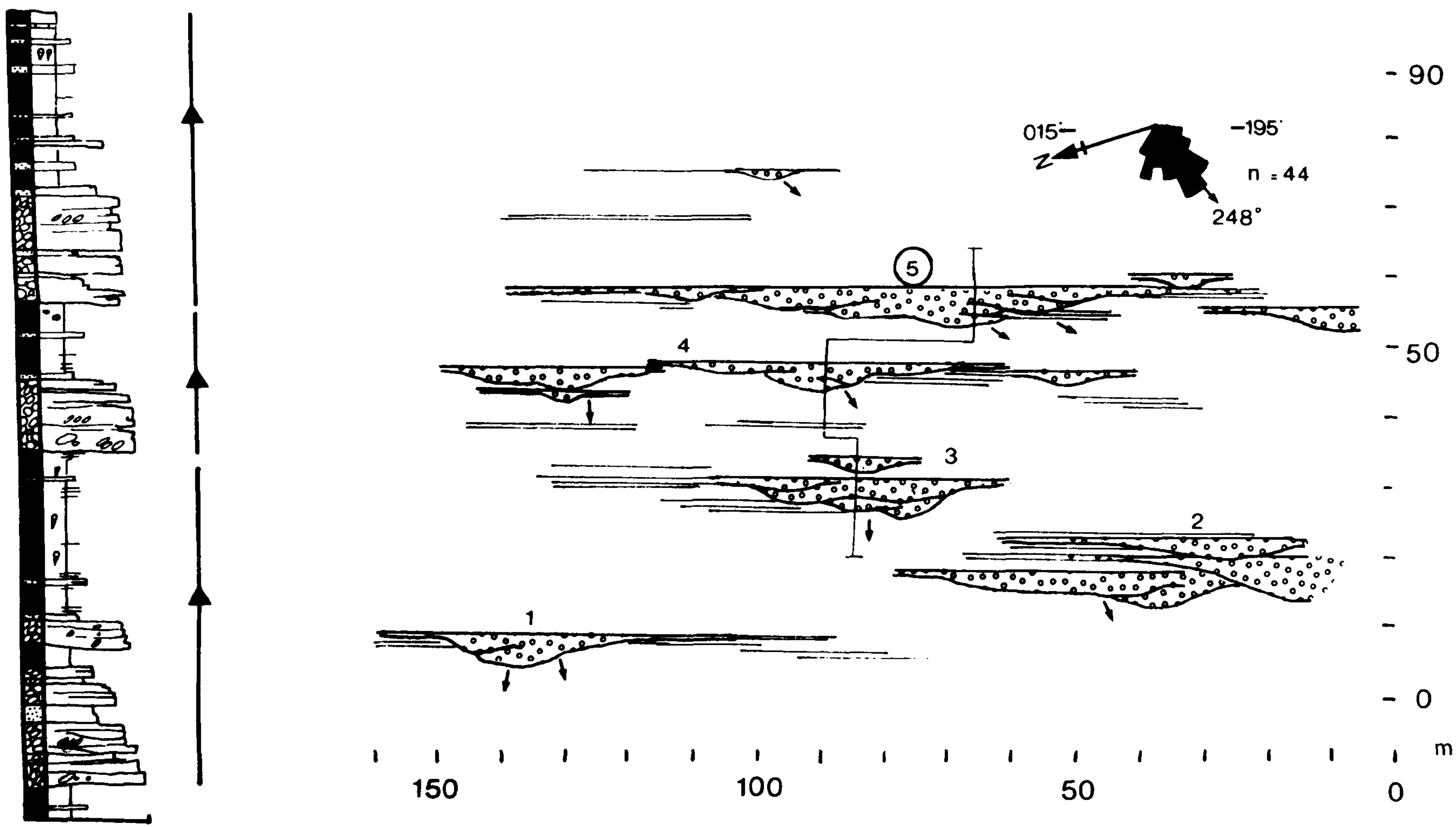


Fig 3.15 (ii) Sketch interpretation and graphic logs of coarse - fine member fining upward and coarsening upward sequences detailed in Fig 3.15 (i)

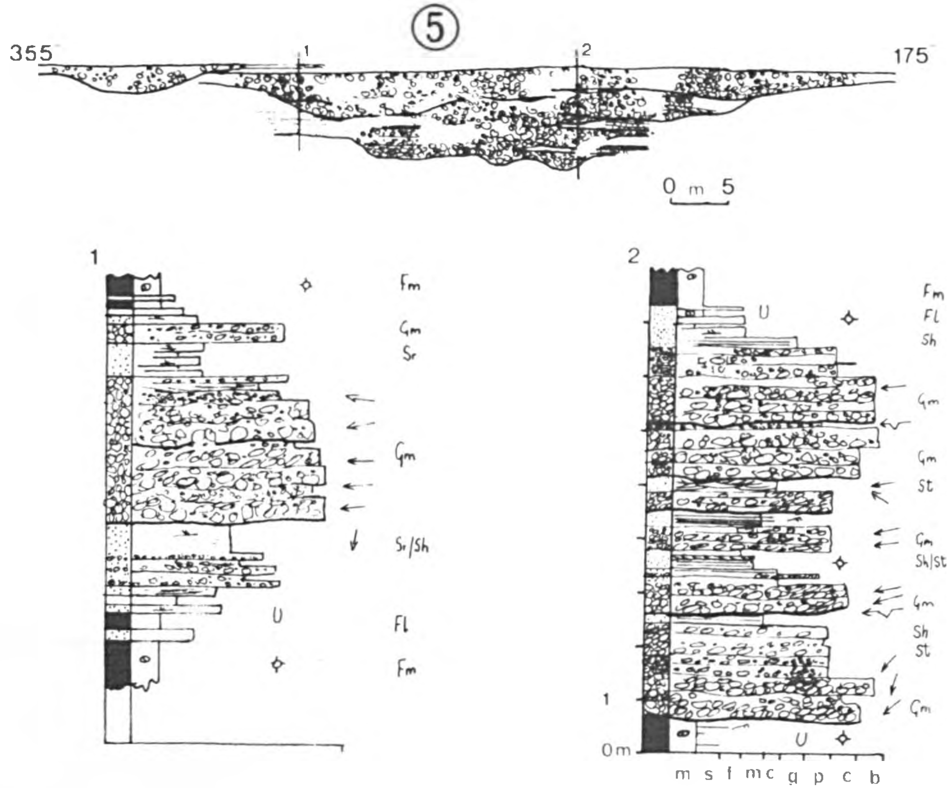


Fig 3.15 (iii) Detail of internal structure of multistorey ribbon channel (5) (D1-D2) in Fig 3.15 (i-ii).

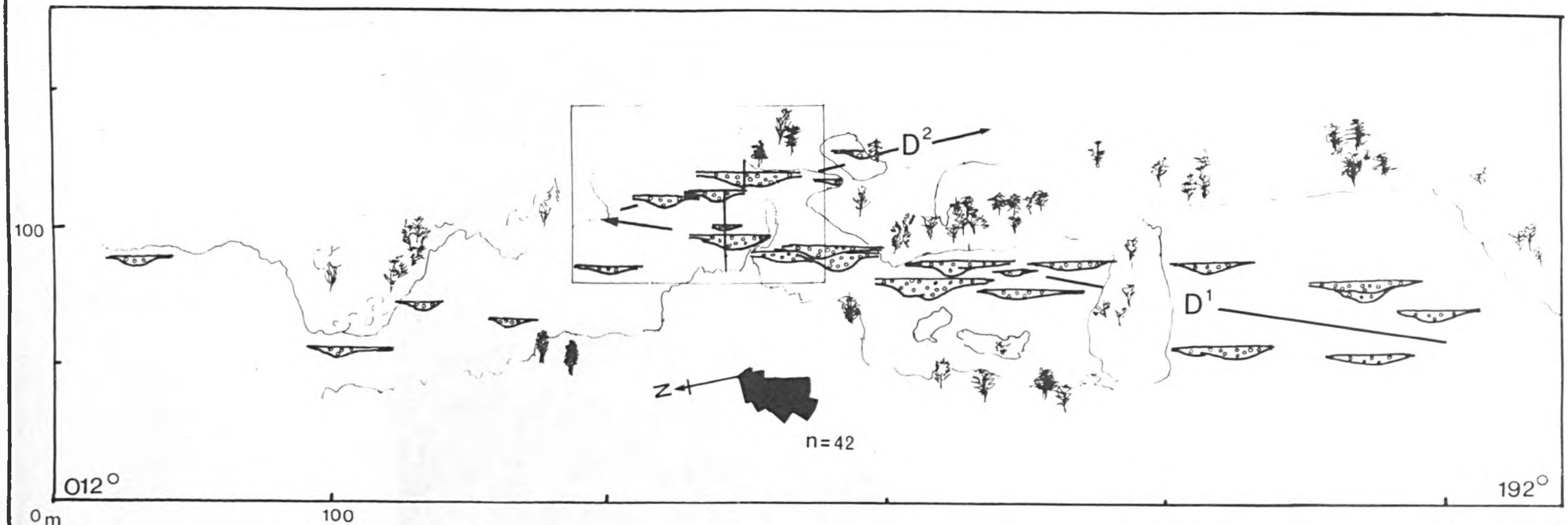


Fig 3.15 (iv) Channel domains of the medial fan facies association. The small box shows the section of the exposure detailed in Figs 3.15 (i) - 3.15 (ii). Finening upward sequence generated by the lateral migration of the channel domains across the line of section at Jonchier. Channel migration may have been in response to inherited channel body topography, or the superimposition of a tectonic slope on the fan (See text).

FIG 3.16 (i) Medial fan facies association, Vancon fan system. St. Symphorien section. Multistorey sheet channel conglomerate bodies (1-3) and gley mudstones form repeated fining upward sequences within the medial fan facies association.

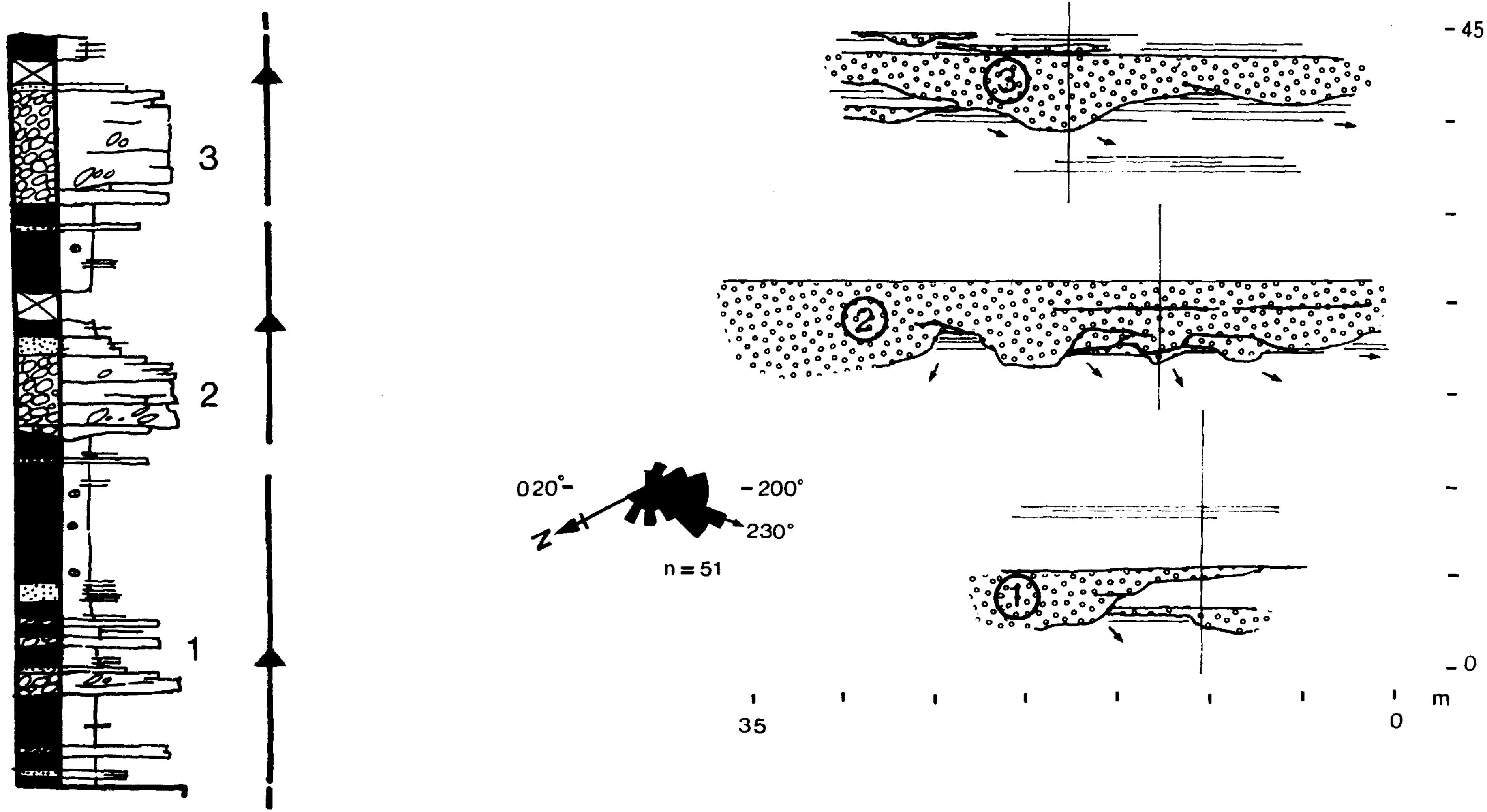


FIG 3.16 (ii) Sketch interpretation of (i) showing position of graphic logs of fining up sequences detailed in Fig 3.14 (i).

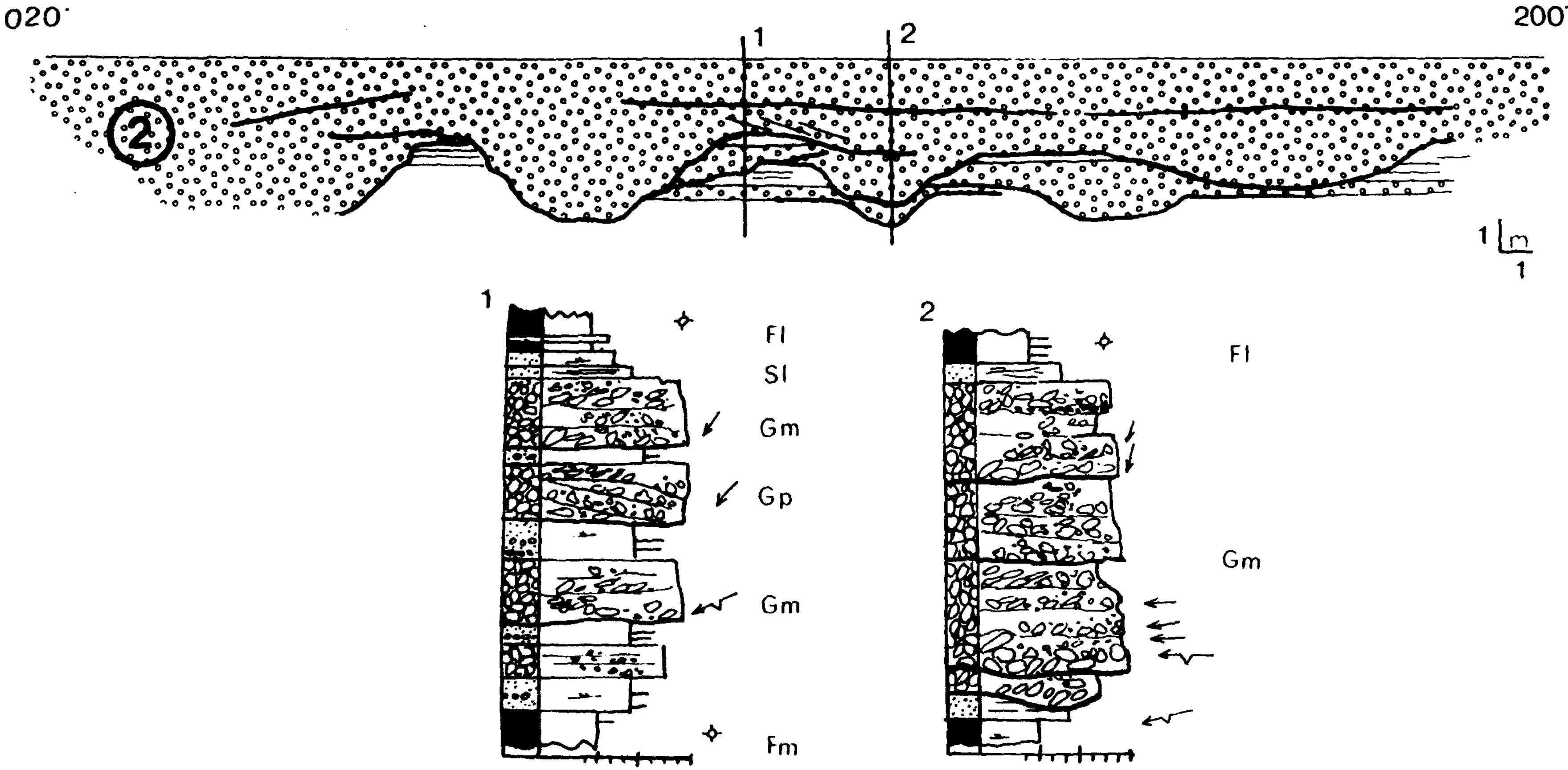


FIG 3.16 (iii) Detail of a multistorey sheet conglomerate body (2) of Fig 3.14a - b. Note development of imbricate (Gm) and planar cross stratified (Gp) conglomerate fill.

FIG 3.17 Horizontally bedded, imbricate, boulder-cobble grade, conglomerate fills a multistorey ribbon channel of the medial fan facies association 1-2 storeys. Note (i) the reddish-brown pseudogley mottling of the fine member mudstones, and (ii) the development of stacked calcrete profiles (C $^{1-4}$) at the top of the fine member immeadiately beneath the channel. Channel is 4.4m thick.

FIG 3.18 Multistorey, ribbon conglomerate, channel body passes out laterally over ~80m into a tabular unit of graded sheet sandstones and pebbly conglomerate.

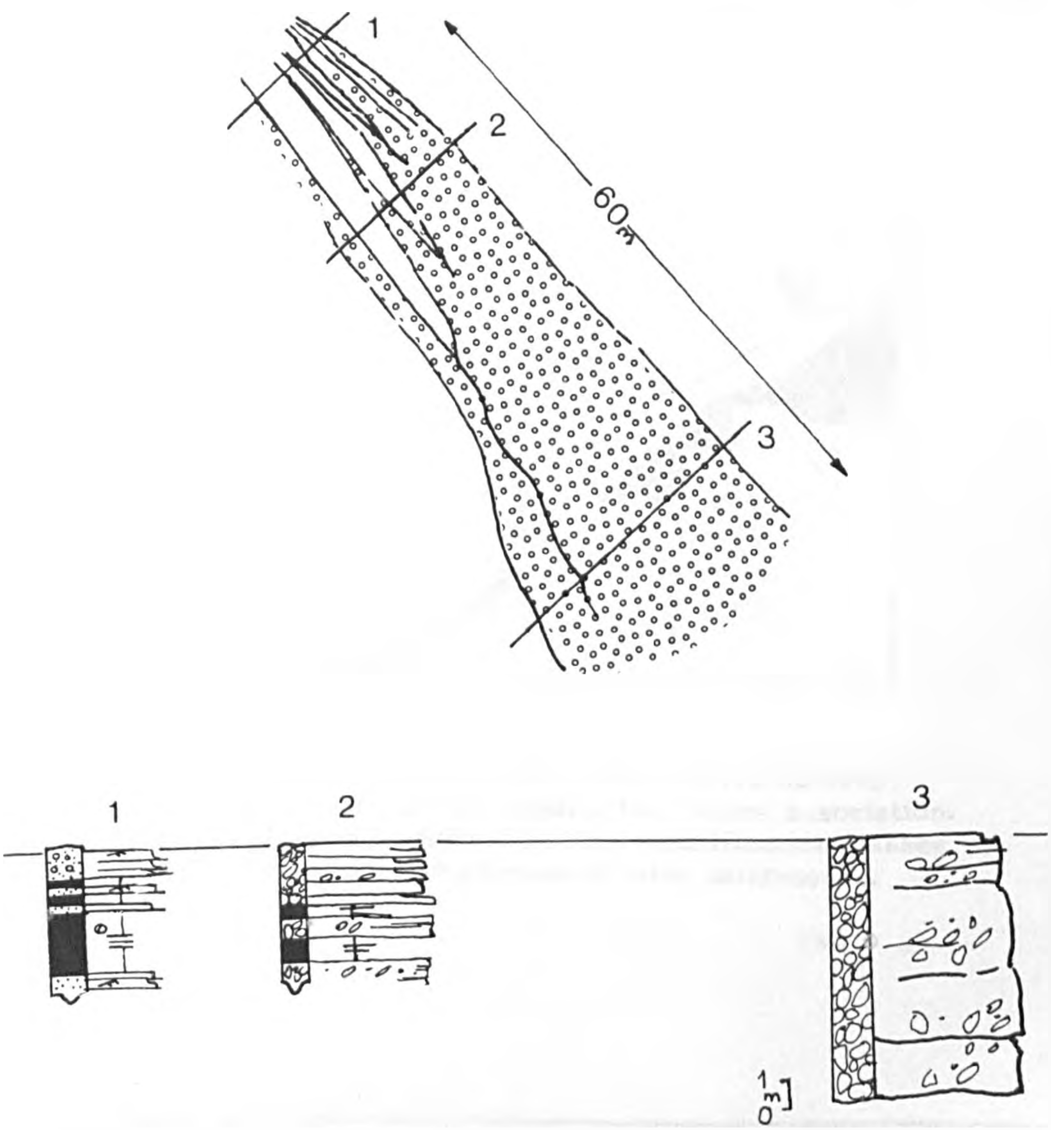


FIg 3.19 (i) Sigmoidal stratification within a multistorey, sheet conglomerate body of the medial fan facies association. (ii) Interpretative sketch. Note how the stratification passes up dip into planar stratified conglomerates and sandstones.

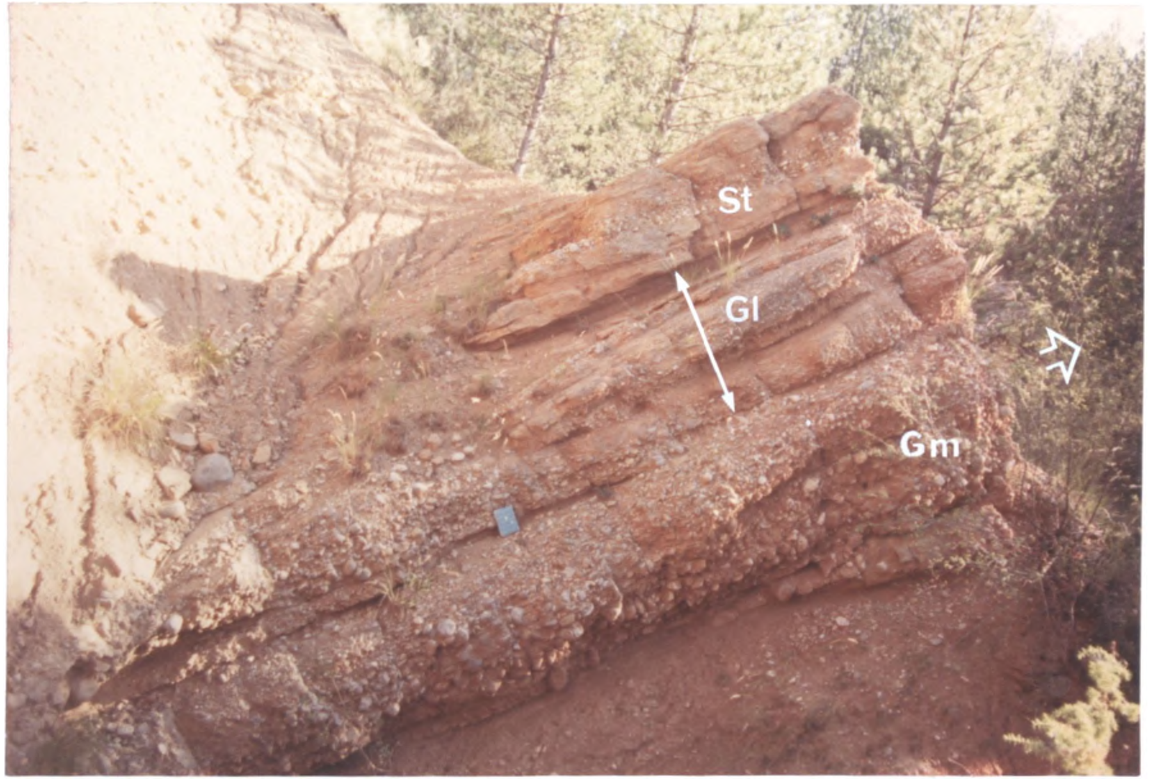
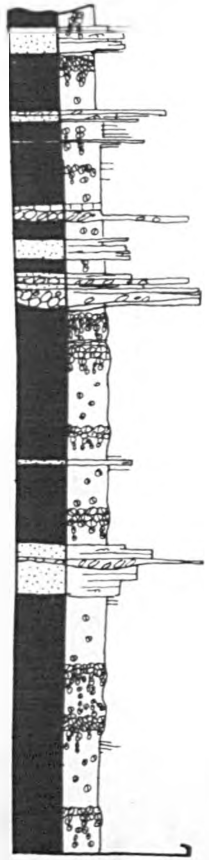


FIG 3.20 Low angle stratification dips upstream at ~15 degrees within the upper part of a medial fan channel body. Internally, it comprises thin, imbricate pebble conglomerates and planar and cross stratified sandstone. View is parallel to flow. Gm - Horizontally stratified conglomerate. Gl - Low angle pebbly stratification. St - Trough cross stratified sandstone. Notebook 25cm for scale.



DESCRIPTION	PROCESS	INTERPRETATION
Single storey ribbon gravel channel with imbricate gravel.	Shallow channel with low relief gravel bars.	
		Distal fan fringe, low lying with impeded drainage, vegetated. Ephemeral braided distributary channels. Overbank splays / terminal channel lobes.
Sandstone units often form small coarsening-upward sequences with small ribbon gravel channel. Sandstones, thin graded with ripple cross-lamination. Dessication cracks.	Deposits of waning sheet flows, - periodic dessication.	
Reddish-brown mudstones (10R4/6) with grey mottle. Repeated stage 3/4 carbonate profiles.	Pseudogley soils. Mature-late mature calcrete profiles reflect low rates of sedimentation.	

FIG 3.21 Distal fan facies association.

FIG 3.22 Distal fan facies association, Vancon fan system, St. Symphorien section. Note the small single storey ribbon channels isolated within the pseudogley mudstone dominated association. The Miocene age marine molasse formation (MM), may be seen conformably overlying the fan sequences in the background.

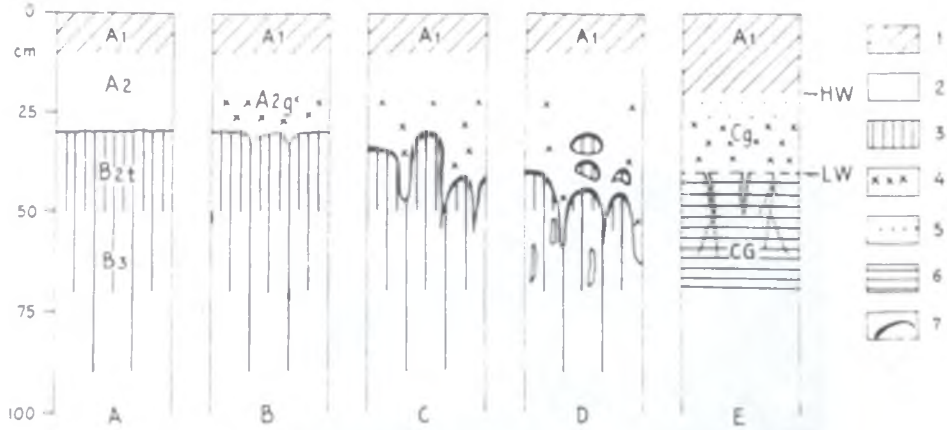
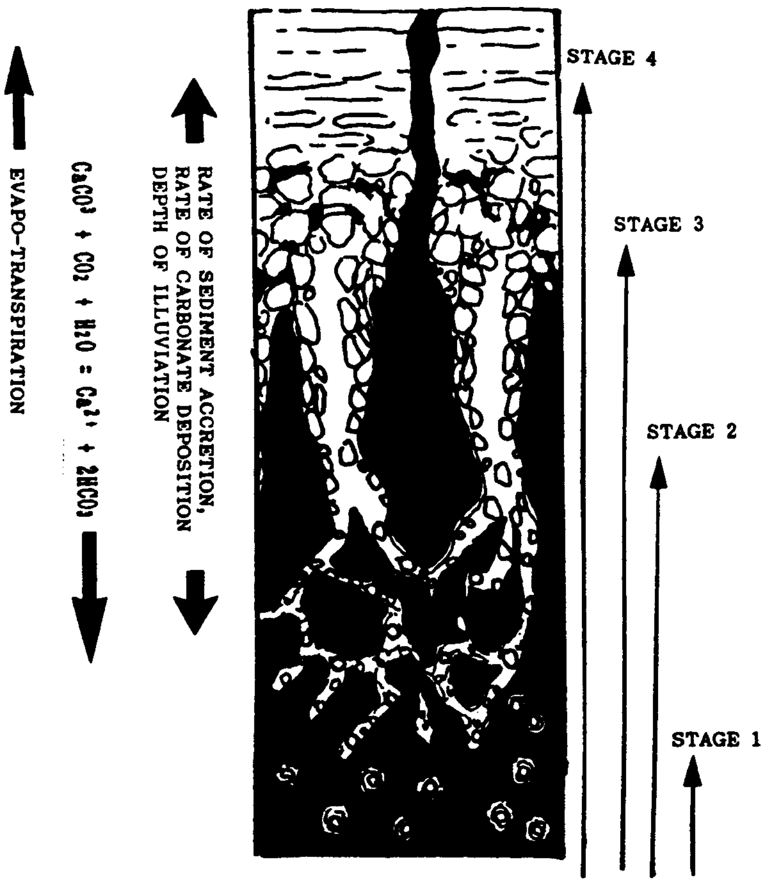


FIG 3.23

The evolution of pseudogley features (A–D) compared with a groundwater gley profile (E) 1. Surface horizon with accumulation of humus; 2. horizon from which clay has been removed; 3. accumulation of clay, decreasing with depth. 4. concentrations of iron compounds (mottles, concretions); 5. concentrations of manganese compounds; 6. reduced (grey) horizon; 7. accumulations of iron as caps and linings.



DESCRIPTION	CALCRETE MATURITY (after Reeves 1970)	CALCRETE DEVELOPEMENT TIME (after Leeder 1975)
Massive carbonate with horizontal fabric given by calcrete filled fracture veins. Mud filled channels replace plant roots.	LATE MATURE	10000+yrs
Glaebules merge to form carbonate framework with host sediment forming relict patches only.	MATURE	6000-10000yrs
Carbonate nodules form elongate ~ vertical pedotubules or sheets. Pedotubules have a downward divergent branching. Carbonate preferentially precipitated within mottle channels.	LATE YOUTH - EARLY MATURE	3500-7000yrs
Carbonate nodules	IMMATURE	1000yrs
Initiation of carbonate precipitation.		

FIG 3.24 Idealised calcrete profiles - stages of calcrete maturity recognised within the fine member deposits of the Vancon fan system.

FIG 3.25 Stacked calcrete profiles within reddish brown pseudogley mottled heterolithic mudstones of the distal fan facies association. Section is some 40m thick.

FIG 3.26 Close up view of stacked, mature (stage 3-4), calcrete profiles in pseudogley mottled mudstone. The mud filled channels passing through the massive calcrete in the foreground are probably plant root moulds. Lens cap 5cm.

FIG 3.27 Pedotubular calcrete - type 1/2. Mottle channel in pseudogley soil shows a downward diverging branching. Note that the glaebules preferentially develop along the margins of the mottle channel (see also Fig 3.23).

FIG 3.28 **Plan view of stage 2 calcrete profile revealing a polygonal framework interpreted as either (i) growth pressure relief fracture fills ('expansion polygons'), or, (ii) dessication crack fills.**

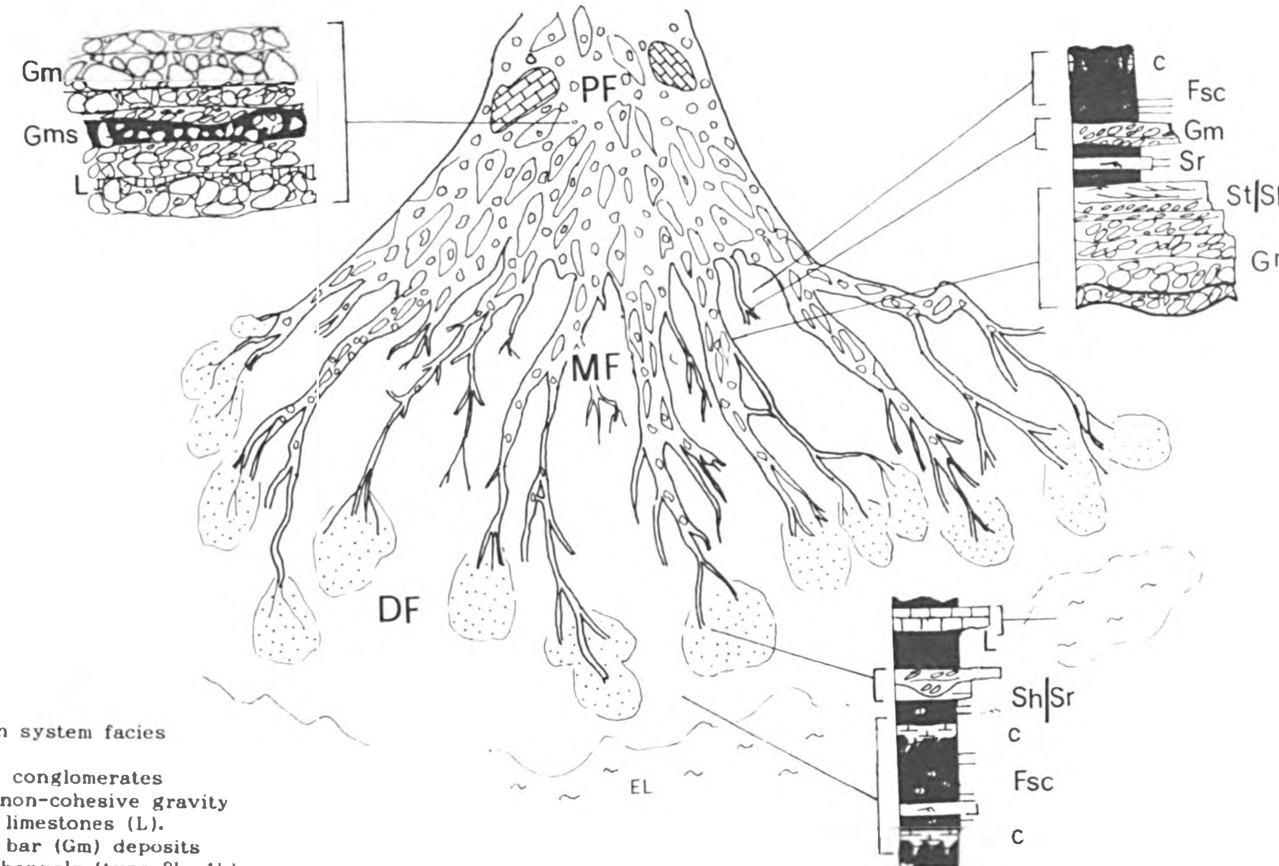


FIG 3.29 Reconstruction of the Vancon fan system facies association distributions.
PF - Proximal fan association:- sheet flood conglomerates (Gm) dominant with cohesive debris flows/non-cohesive gravity flows (Gms) and olistoliths, and occasional limestones (L).
MF - Medial fan association.:- longitudinal bar (Gm) deposits dominant in multistorey, ribbon or sheet channels (type 2b, 1b). Also, longitudinal bar margin deposits (Gp) & (Gs). Heterolithic, sandstone upper channel fill (St, Sh, Sr, Fl). Fine member: ephemeral ribbon channel (Gm), overbank sheet sandstone (St, Sh, Sr), heterolithic mudstone (Fsc), with stage 1-3 calcrete (c). Pseudogley/gley mottling.
DF - Distal fanassociation:- ephemeral ribbon channel with conglomerate fill (Gm), terminal channel lobes/overbank - sheet sandstones (Sh/Sr). Heterolithic mudstones (Fsc), with stage 1-4 calcretes. Palustrine limestones (L).

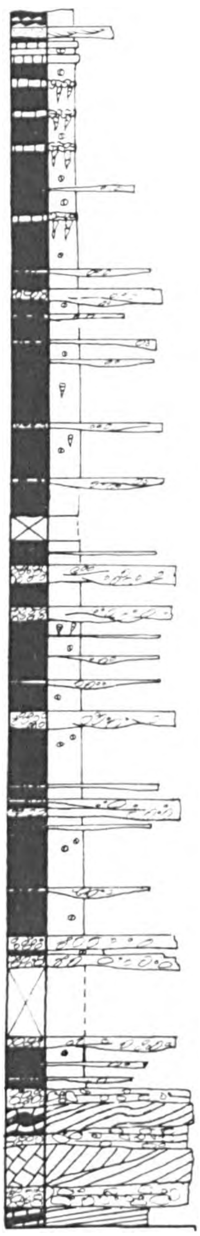


FIG 3.30 Fining up mega-sequence of the Vancon fan system (idealised after the St. Symphorien section).

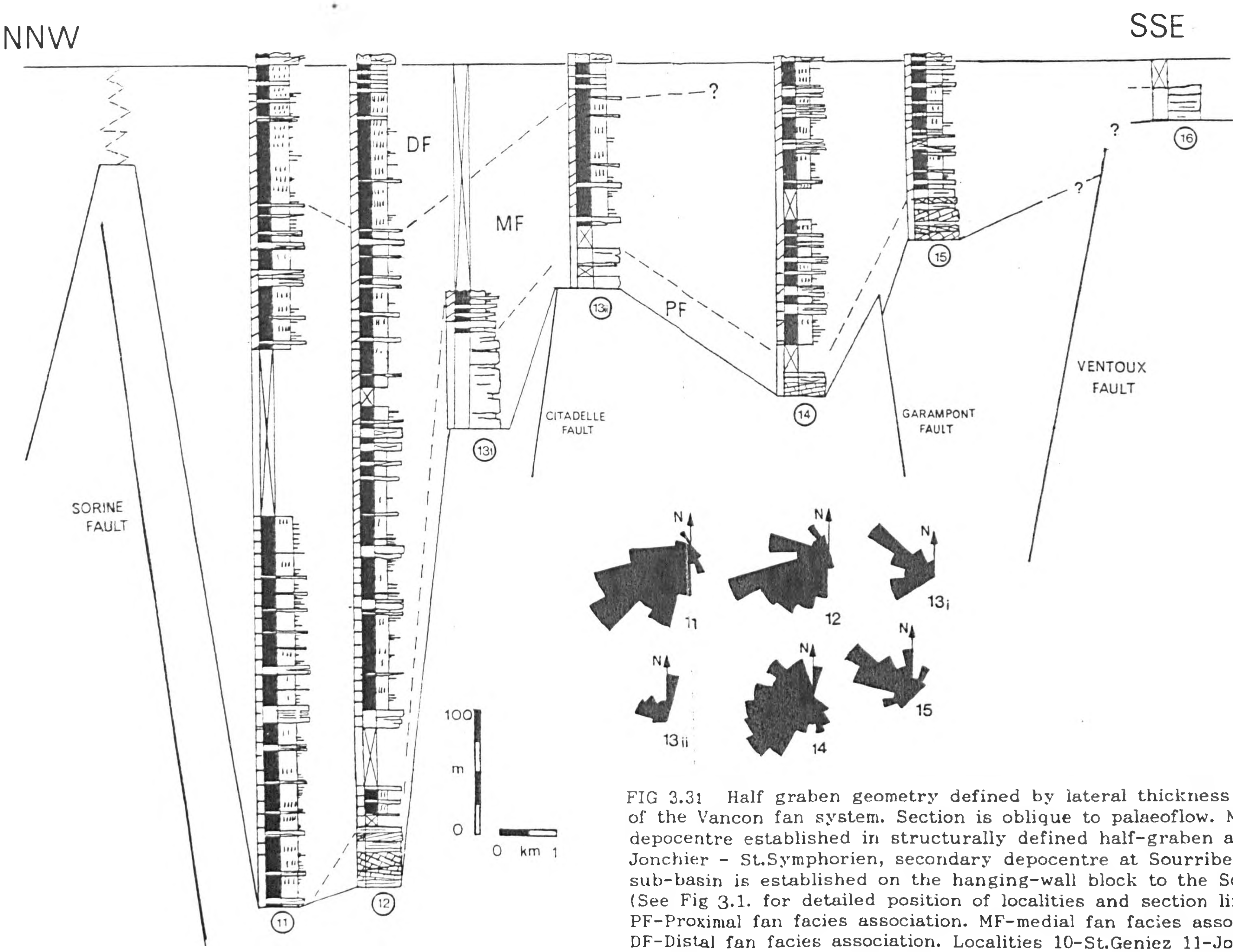


FIG 3.31 Half graben geometry defined by lateral thickness variations of the Vancon fan system. Section is oblique to palaeoflow. Major depocentre established in structurally defined half-graben axis at Jonchier - St.Symphorien, secondary depocentre at Sourribes where a sub-basin is established on the hanging-wall block to the Sorine fault (See Fig 3.1. for detailed position of localities and section line). PF-Proximal fan facies association. MF-medial fan facies association. DF-Distal fan facies association. Localities 10-St.Geniez 11-Jonchier 12-St.Symphorien 13(i) Citadelle 13(ii) Tour de Beaudument 14-Sourribes 15-Volonne 16-Chateau Arnoux.

Pseudo-gley paleosoils ▨ Gley paleosoils ▤

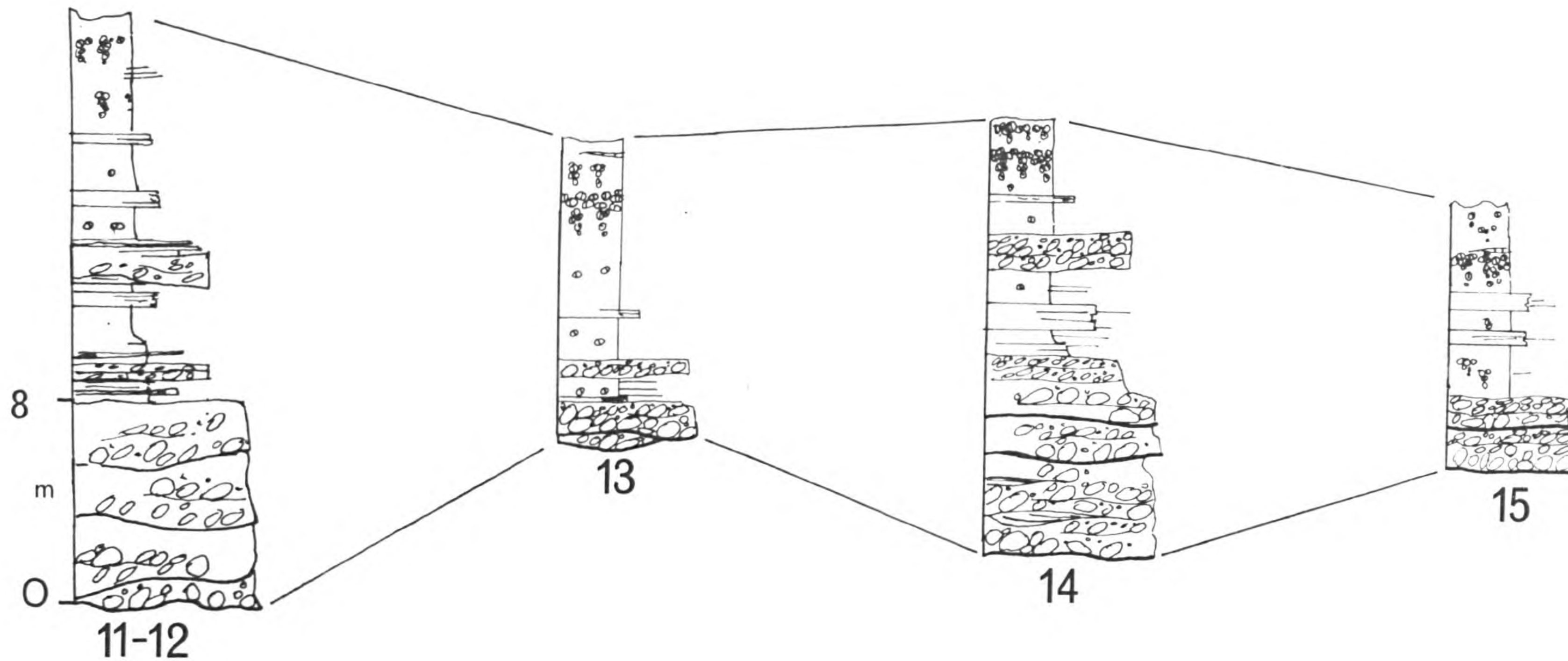


FIG 3.32 Lateral variation in scale of medial fan facies association sequences within the Vancon fan system. The maximum channel thicknesses are attained in the axis of the half-graben (Jonchier-St.Symphorien : localities 11-12), and in the axis of the sub-basin on the hanging-wall block (Sourribes : locality 14) The intervening structural highs are characterised by smaller and finer grained conglomerate channels.

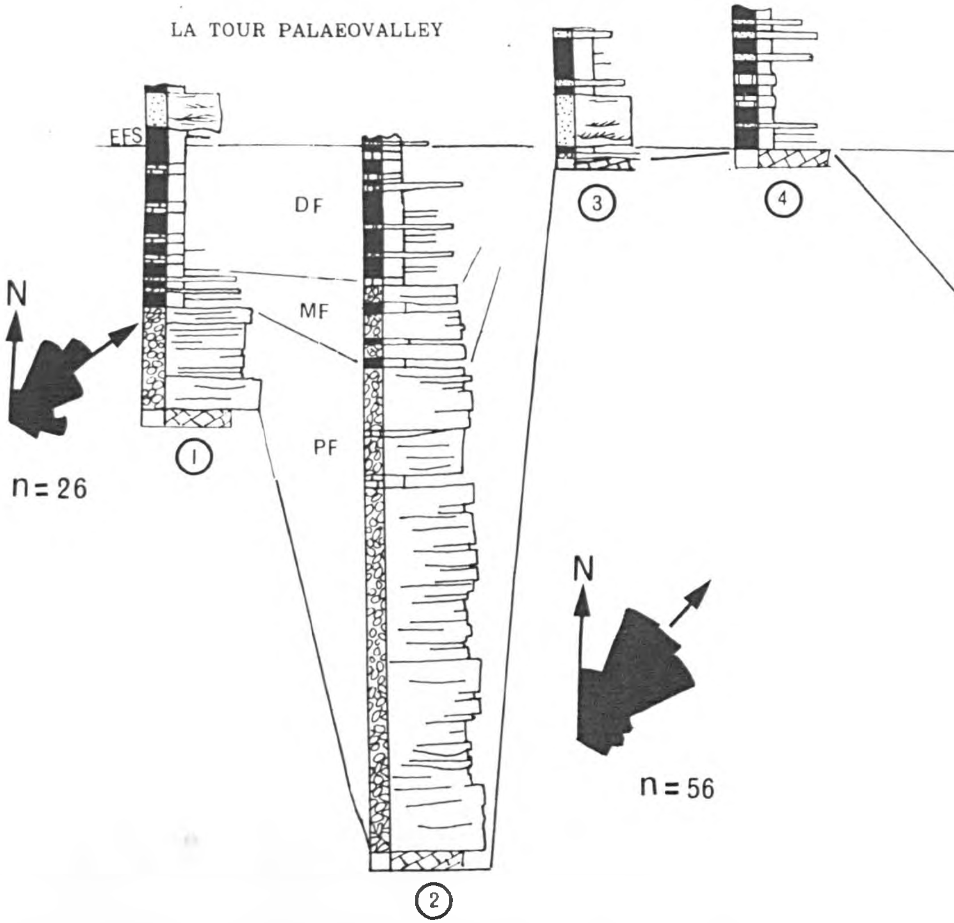
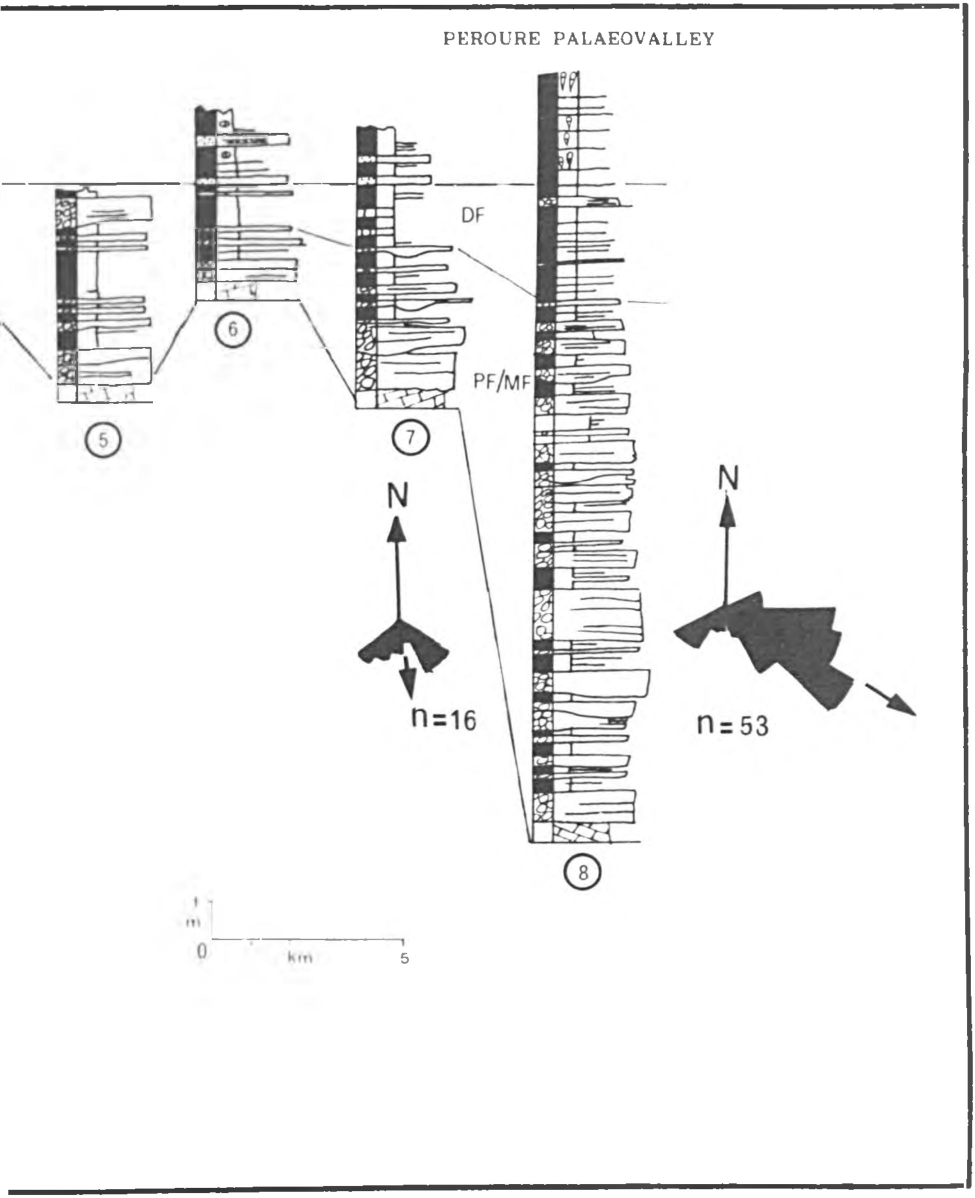


FIG. 3.33 Palaeovalleys at the base of the Molasse Rouge Formation lateral thickness variations of locally sourced, mass-flow fan deposits beneath the Esclangon fluvial system suggest the developement of a least three palaeovalleys, of which the La Tour and Peroure have been studied. Sections: 1-Le Pategue 2-la Tour 3-Maregineste 4-Maladrech 5-Aubeze 6-le Fiere 7-la Pare 8-Peroure.



PEROURE PALAEOVALLEY
DF
PF/MF
5
6
7
8
N
n=16
N
n=53
m
0
km
5

FIG 3.34 Horizontally stratified conglomerates of the Proximal fan facies association, La Tour section.

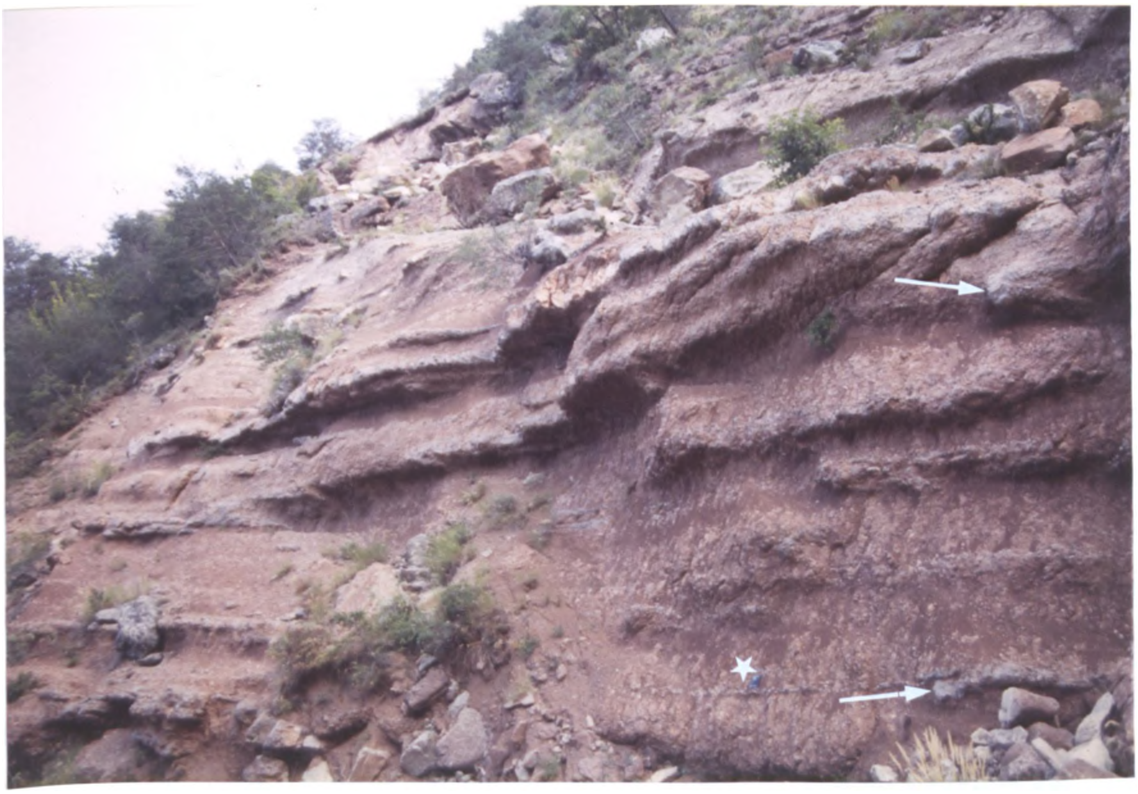


FIG 3.35 Thin cohesive debris flow conglomerates and mudstones of th Distal fan facies association. Note the small, steep sided ribbon scours developed along the length of a some flows. Arrows indicate palaeoflow of channels. Notebook 30cm for scale.

FIG 3.36 Matrix supported, cohesive debris flows, with inverse grading and angular clasts.

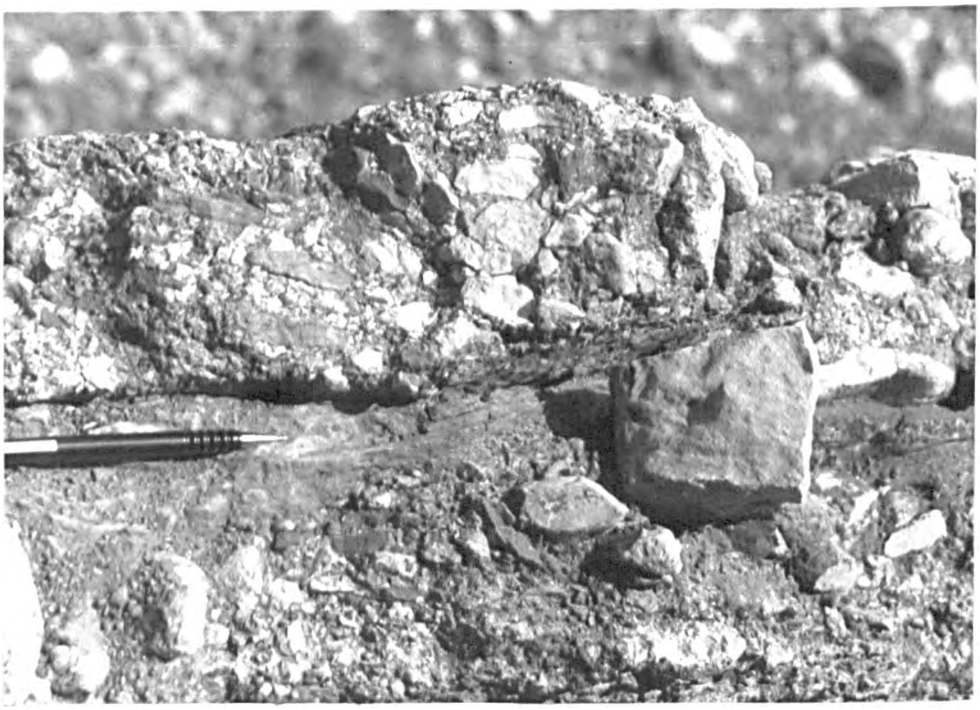


FIG 3.37 Close up of cohesive debris flows. Note (i) the inverse grading of the lower bed (ii) erosive base and imbrication of upper bed suggesting deposition from a more fluidal sheetflow.

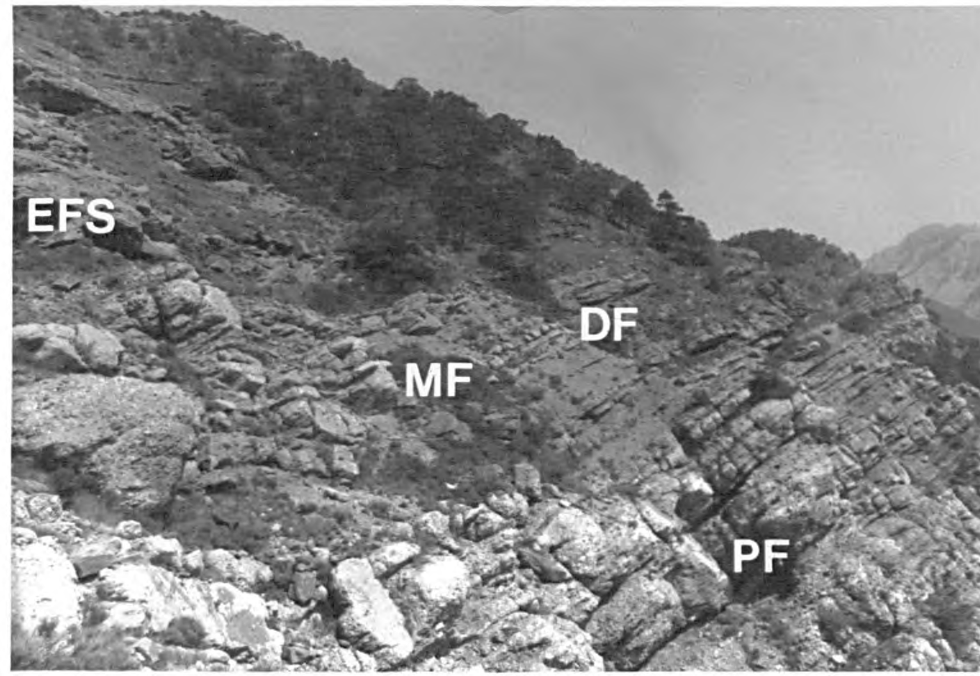
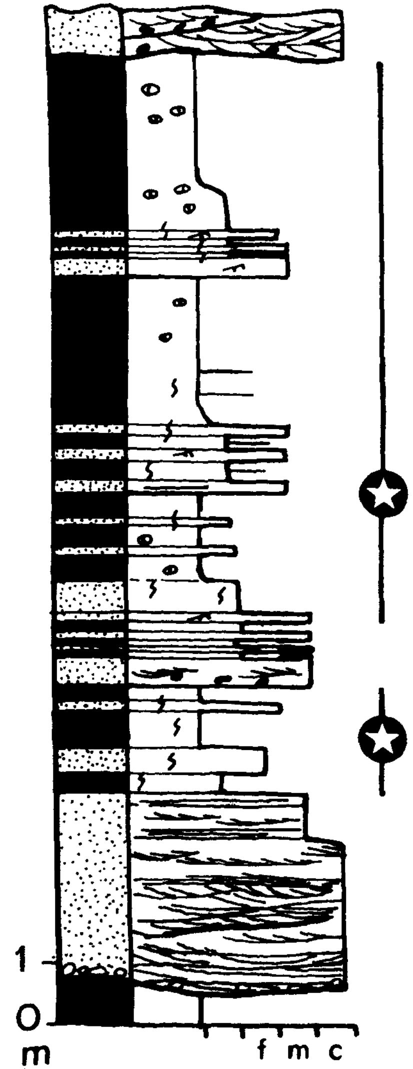


FIG 3.38 Fining-upward (mega-sequence) fill of the La Tour palaeovalley at the La Tour section. PF-Proximal fan MF-medial fan DF-Distal fan facies association. EFS-Esclangon fan system.



DESCRIPTION	INTERPRETATION	
Fine Member Progressive fining-upward, reddish mudstones with grey mottle and carbonate glaebules (stage 1). Small ribbon channel sandstones, tabular units of sheet sandstones.	Impeded drainage, pseudo-gley mottling Immature calcretes. Chute cut-off / crevasse splay / small distributary channel. Levee / crevasse splay sandstones.	Interchannel Floodplain.
Coarse Member Multistorey channel body (ribbon, sheet, tabular) Fining-upward fill dominated by trough x-stratified granular sandstone. Inclined bedding.	Sinouos, mixed-load fluvial channels Developement of lateral accretion bedding. 3-D dunes migrated peak discharge.	Main Distributary Channel

FIG 3.39 Coarse-fine member fining upward sequence of the Alluvial Plain facies association.

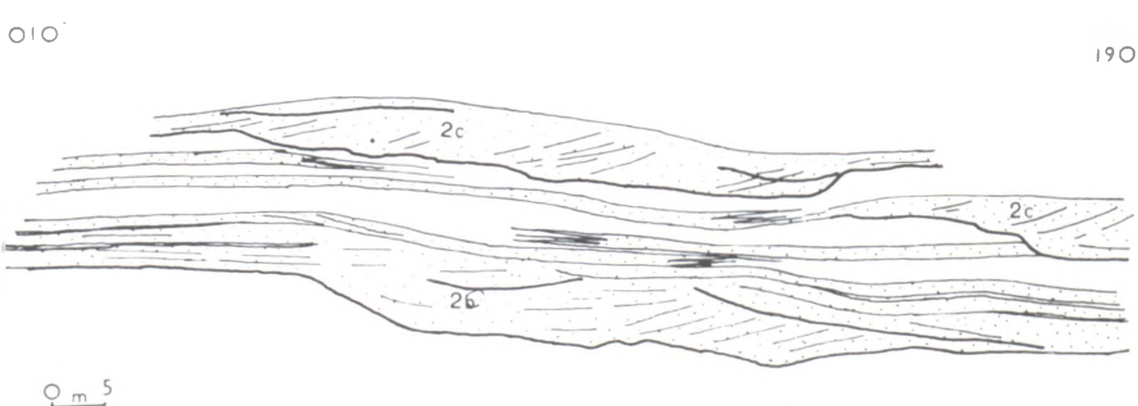


FIG 3.40 Photograph (i) and interpretative sketch (ii) of multistorey (2b) and multilateral (type 2c) ribbon channel bodies (coarse member) of the Alluvial Plain facies association, Esclangon fan system, Esclangon section.

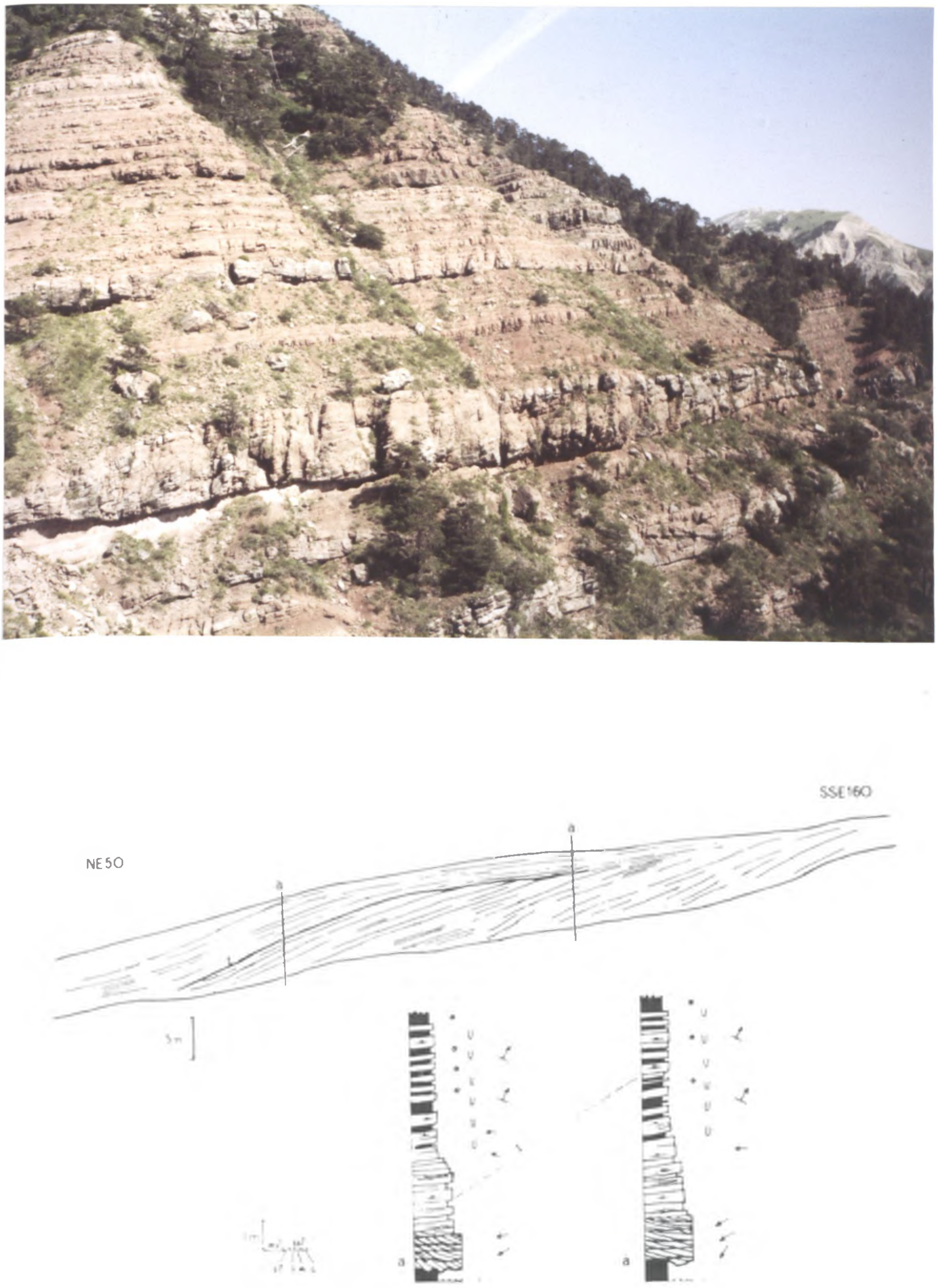


FIG 3.41 Photograph (i) and interpretative sketch (ii) of tabular (type 3a) channel body of the Alluvial Plain association, Esparron section. Note the well developed lateral accretion bedding interupted by an internal erosion surface [1].

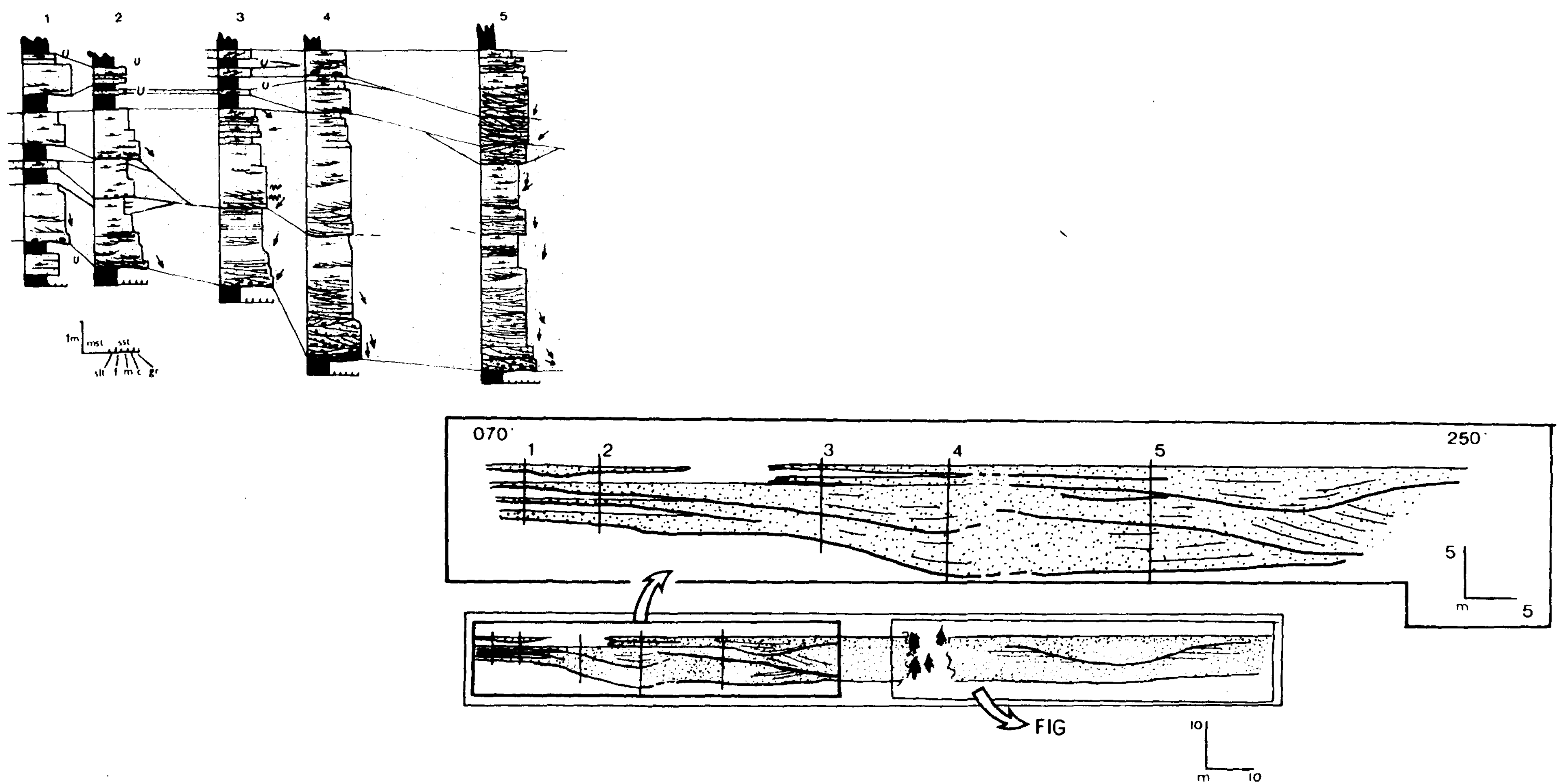


FIG 3.42 Photograph (i) and interpretative sketch (ii) of multistorey sheet (type 1b) channel body, Esclangon fan system, Esclangon section. Note the developement of small, single storey ribbon channels (type 2a-arrowed) in the fine member.

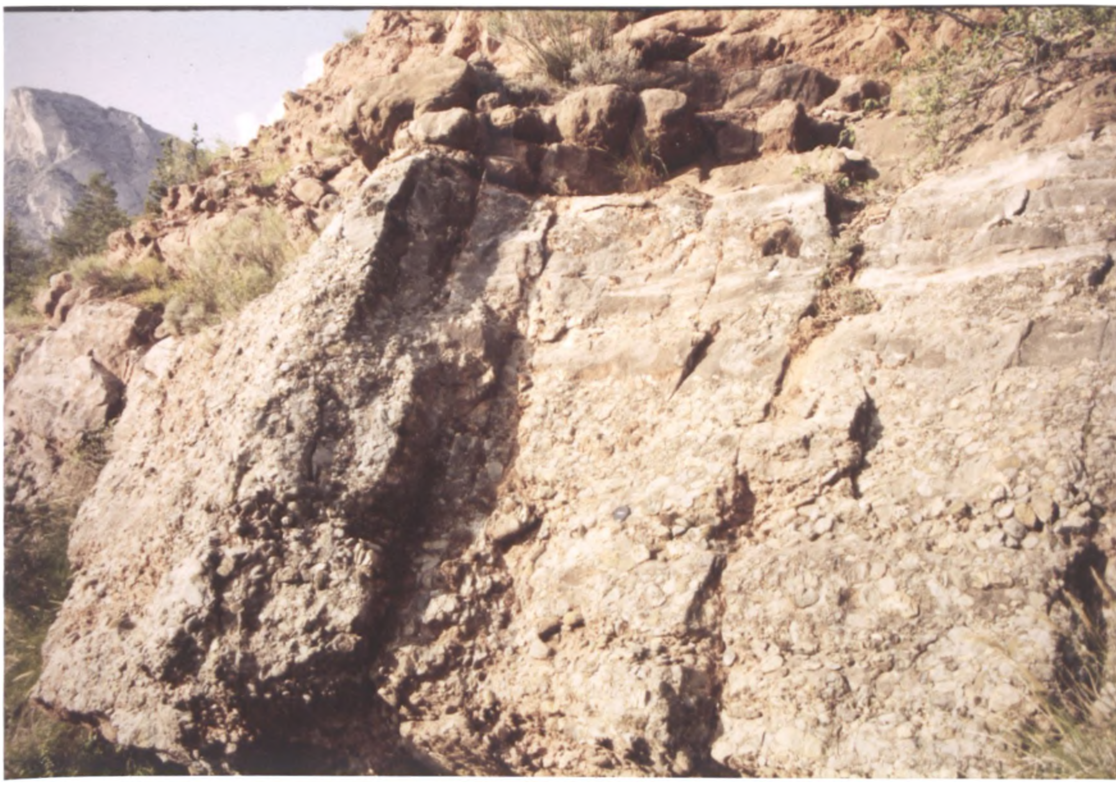


FIG 3.43 Horizontally stratified, imbricate conglomerate at the base of a multistorey channel body, Alluvial Plain facies association.

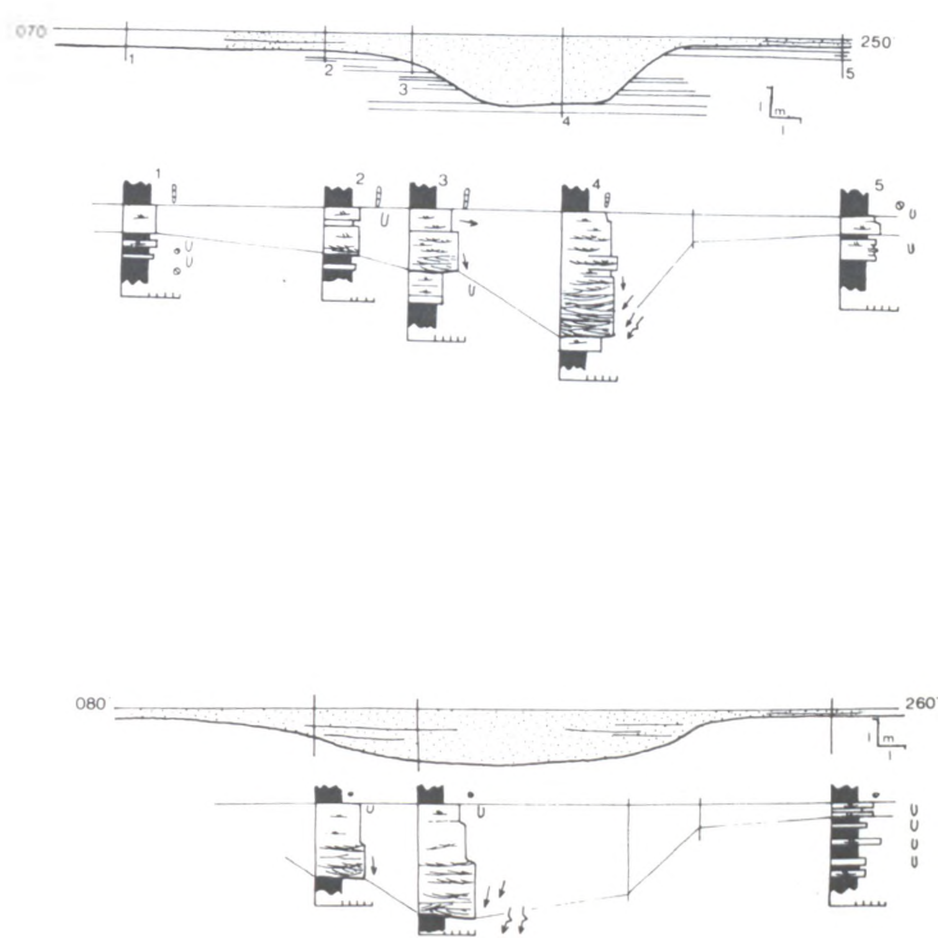


FIG 3.44 Single storey ribbon channels (type 2a) of the fine member, Alluvial Plain facies association.

FIG 3.45 Offset channel domains within the alluvial plain facies association at Esclangon fluvial system at Ravine du Rouuset, Esclangon, (416, 899. La Javie 1:25 000).
(i) Oblique view of multistorey channel bodies within fine member mudstone and sheet sandstones of the alluvial plain association.

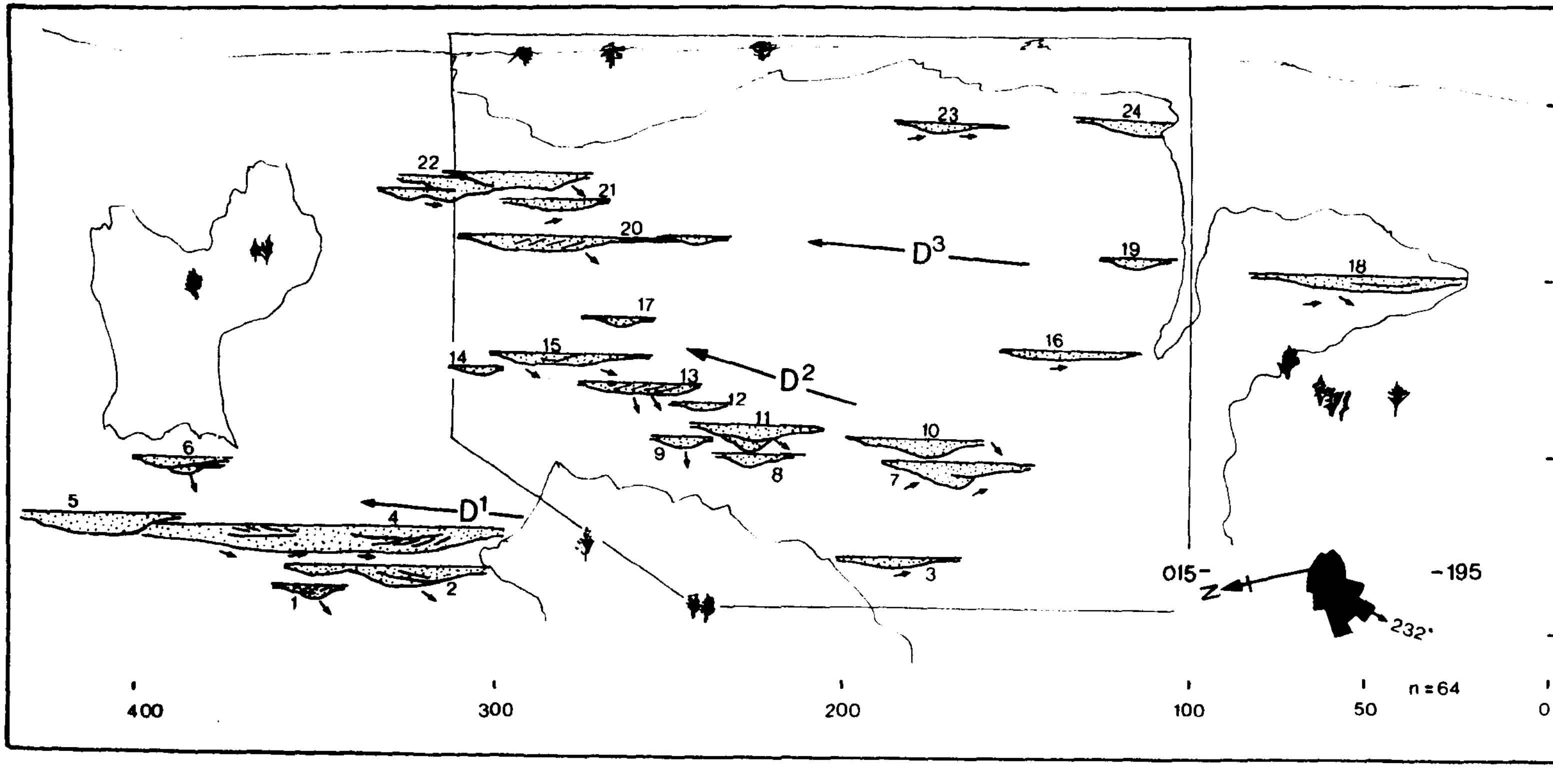


FIG 3.45 (ii) Interperative sketch of Ravine du Rousset exposure, with position of plate in Fig 3.44 (i) outlined. Channel domains D1 - D3 show a progressive lateral offset toward the NW, punctuated by major SE directed avulsion events attributed to downfaulting on the basins margin.

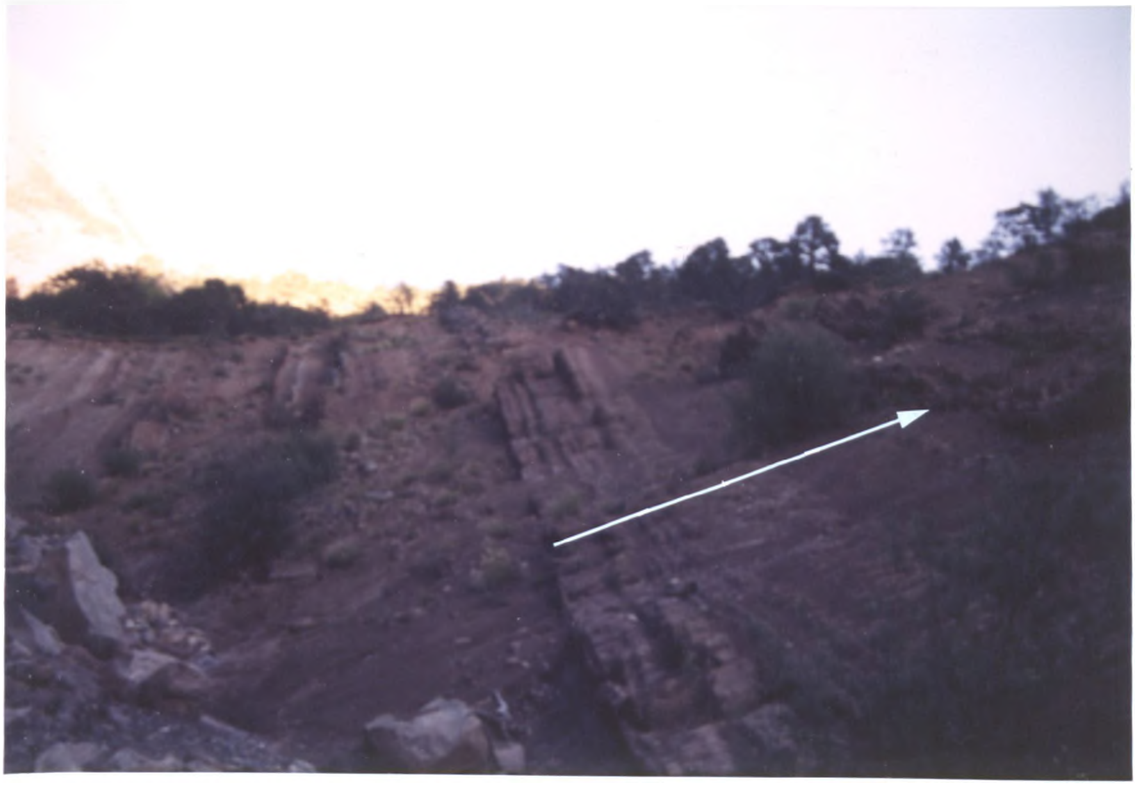


FIG 3.46 Tabular units of sheet sandstone within mudstones of the fine member, Alluvial Plain association. Note the fining upward trend of the unit in the foreground (arrowed).

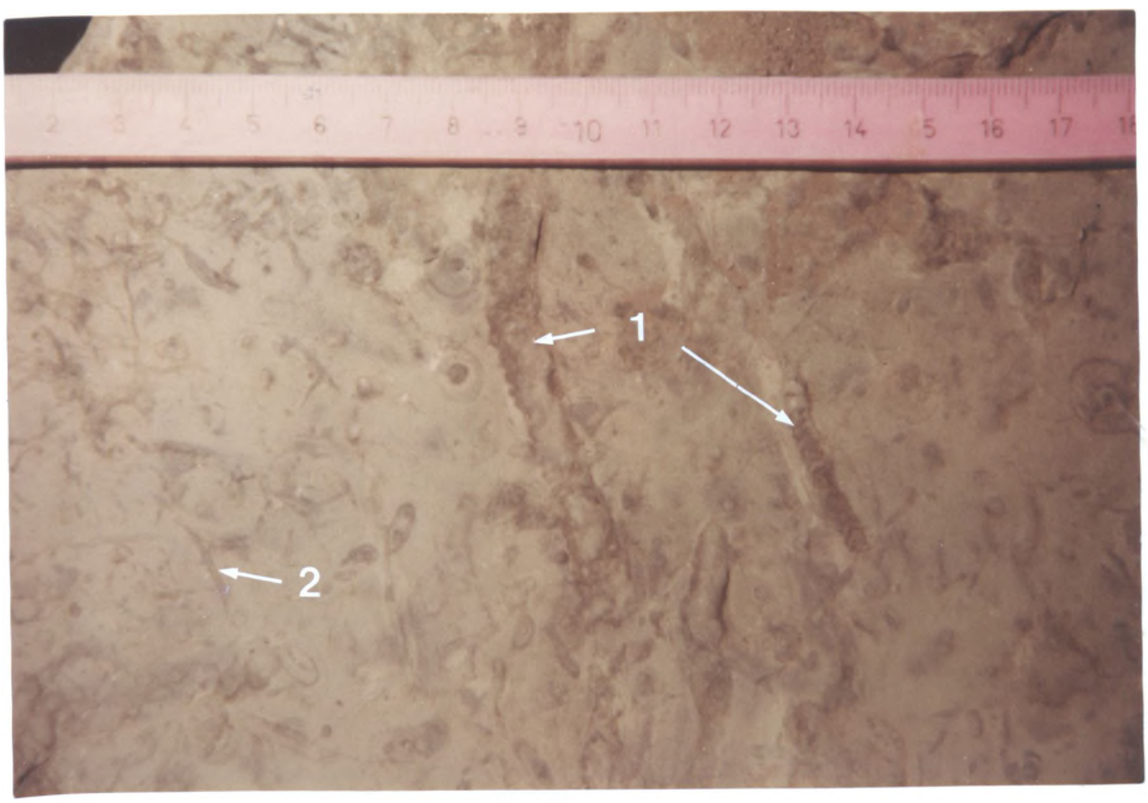
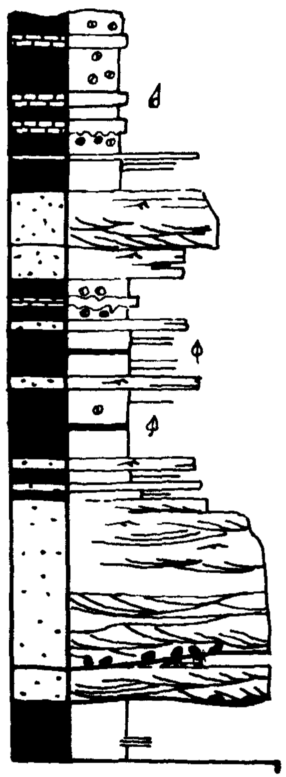


FIG 3.47 Bioturbation within fine member deposits of the Alluvial Plain association. Two types of burrow (1-2) are interpreted as being produced by adult beetles and beetle larvae respectively (See text).



DESCRIPTION	INTERPRETATION	
Fine Member Fining-upward, grey mudstones with carbonate nodules -> Massive, nodular limestone, and yellow mottle. Tabular units of sheet sandstone, small ribbon channels.	Poorly drained mudstones with gley mottling, calcretes, palustrine/ lacustrine limestones. Overbank levee, crevasse splay sandtones.	Interchannel Floodplain with Ephemeral Lakes.
Coarse Member Multistorey channel, ribbon body (type 2b-c), fining-upward fill dominated by trough x-stratified sandstone, inclined bedding.	Sinuous, mixed-load channel - lateral accretion bedding.	Main Distributary Channel.

FIG 3.48 Fining-upward sequence within the Distal Plain facies association.

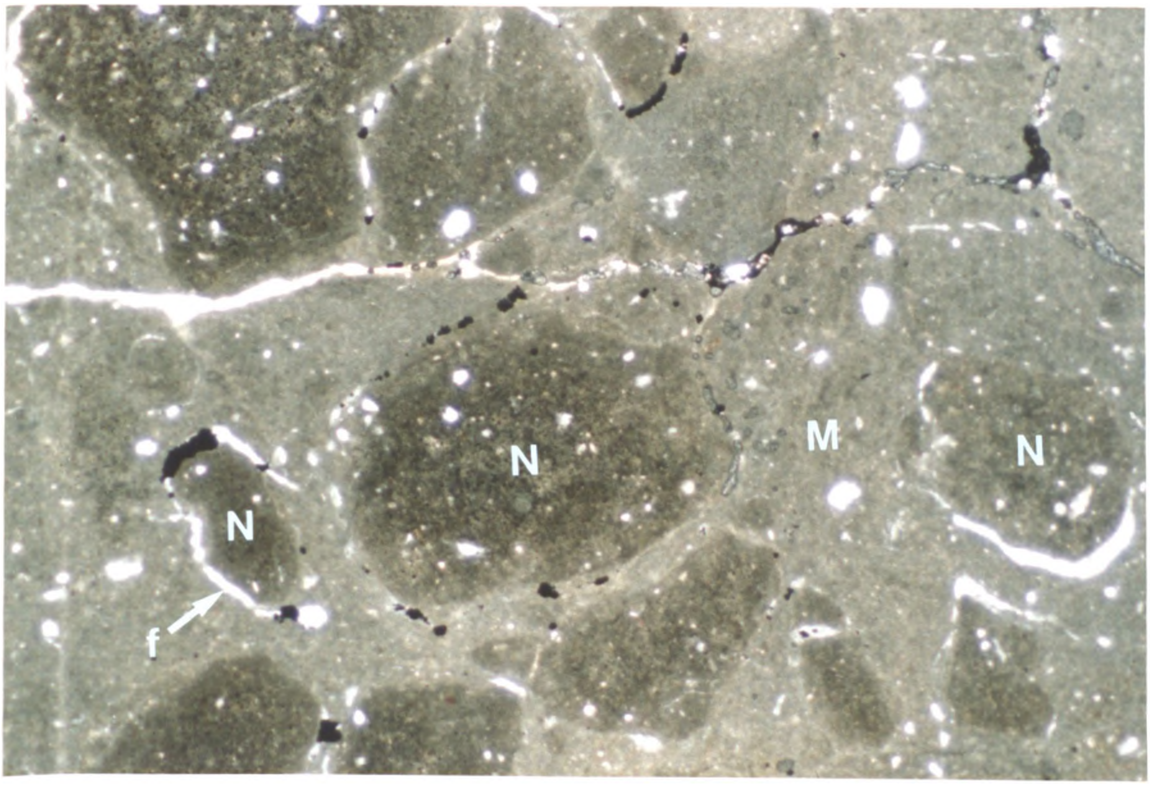


FIG 3.49 Thin section of pedogenic nodules in micritic palustrine limestone. Nodules (N) defined by curved and circular fractures (F) (micro-spar filled) in host micrite (M). X 15

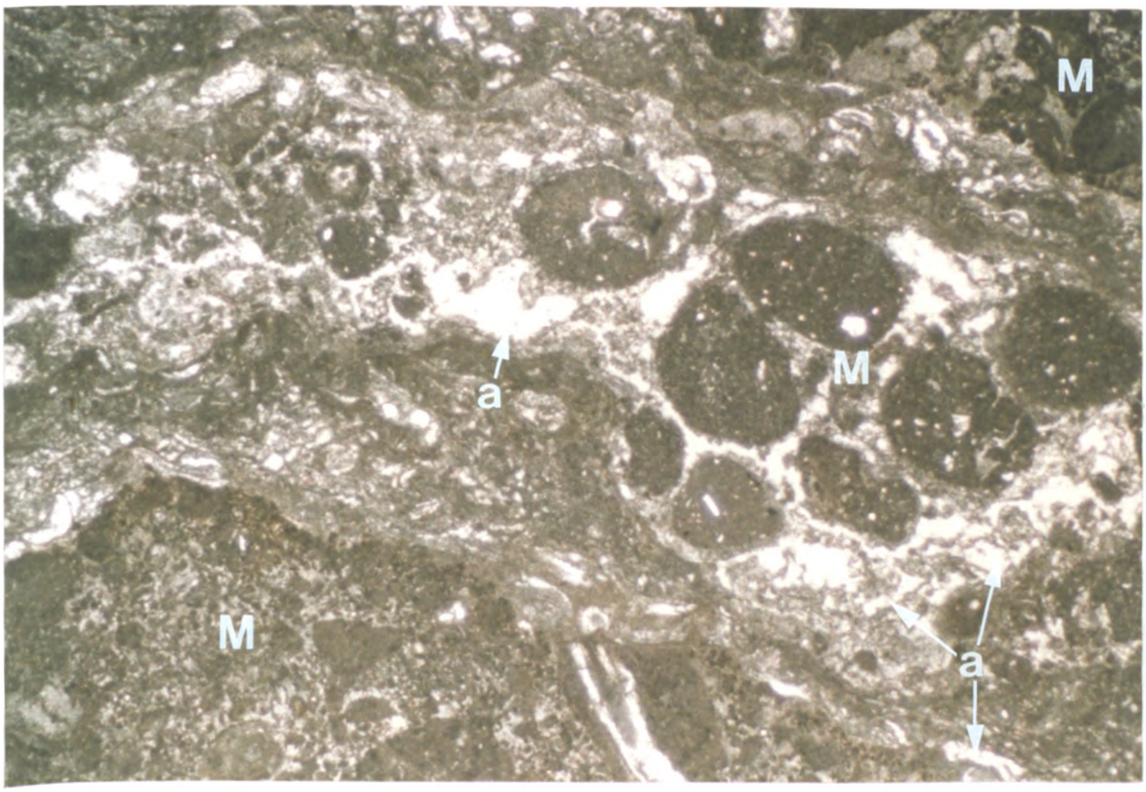
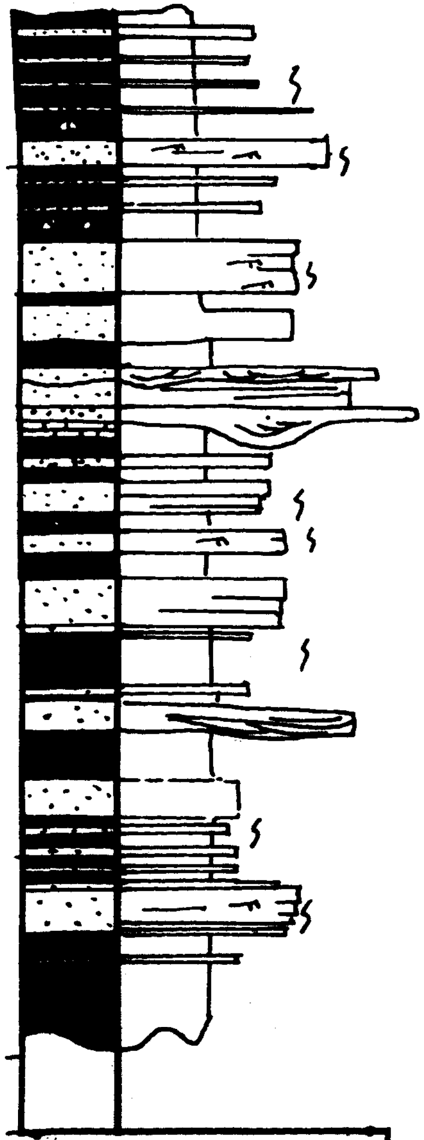


FIG 3.50 Thin section af algal laminated palustrine limestone. Intimate interlayering of peloidal micrite (M) and crenulated algal filaments (a) replaced by micro-spar.



DESCRIPTION	INTERPRETATION
Thinly bedded sandstones with sheet form, in tabular units of dm-m scale traceable for hundreds of metres. Beds have planar erosive base, normally graded. Thicker beds with ripple lamination and trough cross-stratification. Heavily bioturbated by beetle burrows	Sheet sandstones deposited from flows issueing from the mouths of active channels in distal portion of fans.
Reddish brown mudstones with weak grey mottle or homogenous, small calcrete glaebules, rarely, thin micritic limestone	Mudstones grey mottling characteristic of gley soil processes in flat area with impeded drainage.
Small single storey, ribbon channel sandstone with trough cross-stratified fill.	Ephemeral feeder channels

FIG 3.51 Graphic log of a representative section of the Distal Terminal Fan Association, Esparron.

FIG 3.52 Distal terminal fan facies association. Tabular units of graded, sheet sandstones within heterolithic mudstone. Note the isolated ribbon channel (single storey- type 2a; arrowed). Channel is 1.5m thick for scale.

FIG 3.53 Tabular units of graded sheet sandstones of the distal fan association extend for several hundreds of metres at Esparron. K - Cretaceous substrate. EFS - Esclangon fan system.

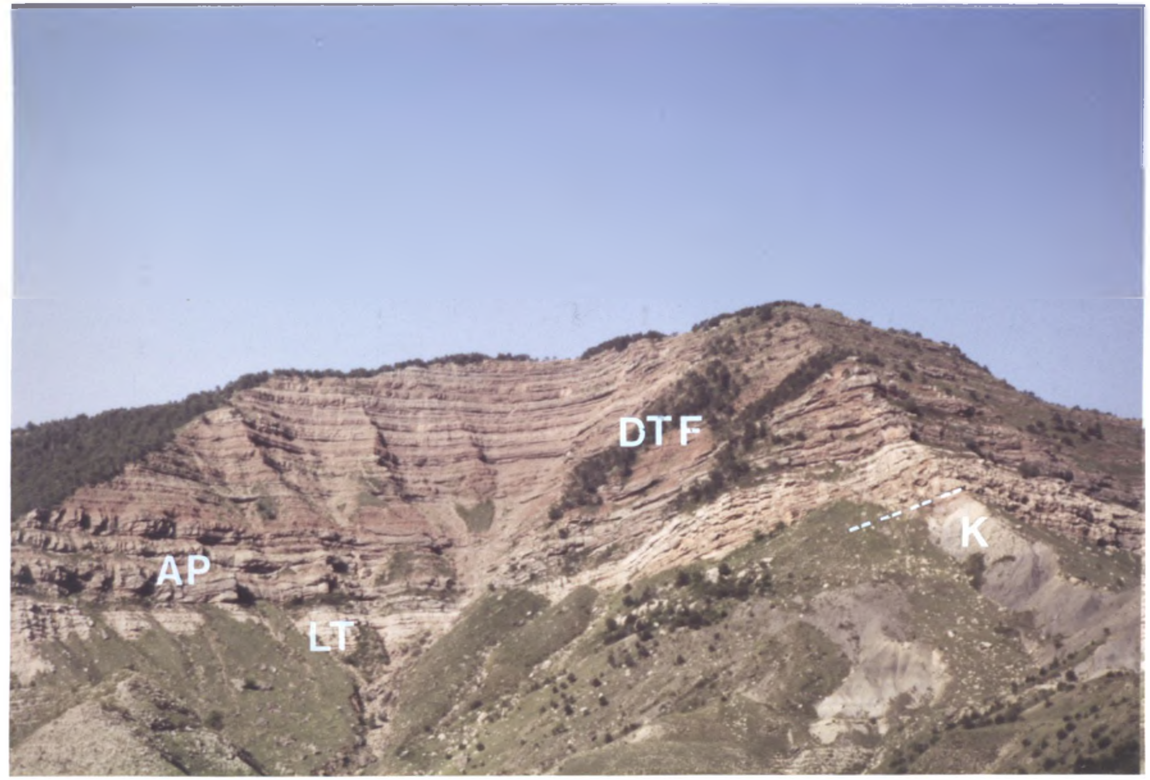
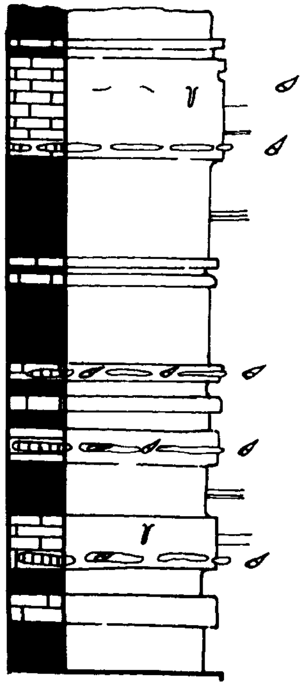


FIG 3.54 Large scale fining-upward mega-sequence of the Esclangon fluvial system as exposed between la Tour and le Pategue, Esclangon. K- Cretaceous. LT-La Tour Palaeovalley AP-Alluvial plain association DTF-Distal terminal fan association.



DESCRIPTION

Dm-m scale micritic limestone and calcareous mudstone. Limestone contains *Planorbid* gastropods, ostracods, characea, plant debris. Chert nodules develop along gastropod shell layers. Occasionally thin graded siltstones. Chert nodules.

INTERPRETATION

Biochemical micrite production in fresh water. Periodic deposition from dilute turbulent flows. Shell layers generated by periodic eutrophication. Silica precipitation post-depositional.

Hydrologically closed lake.

FIG 3.55 Lacustrine facies association.

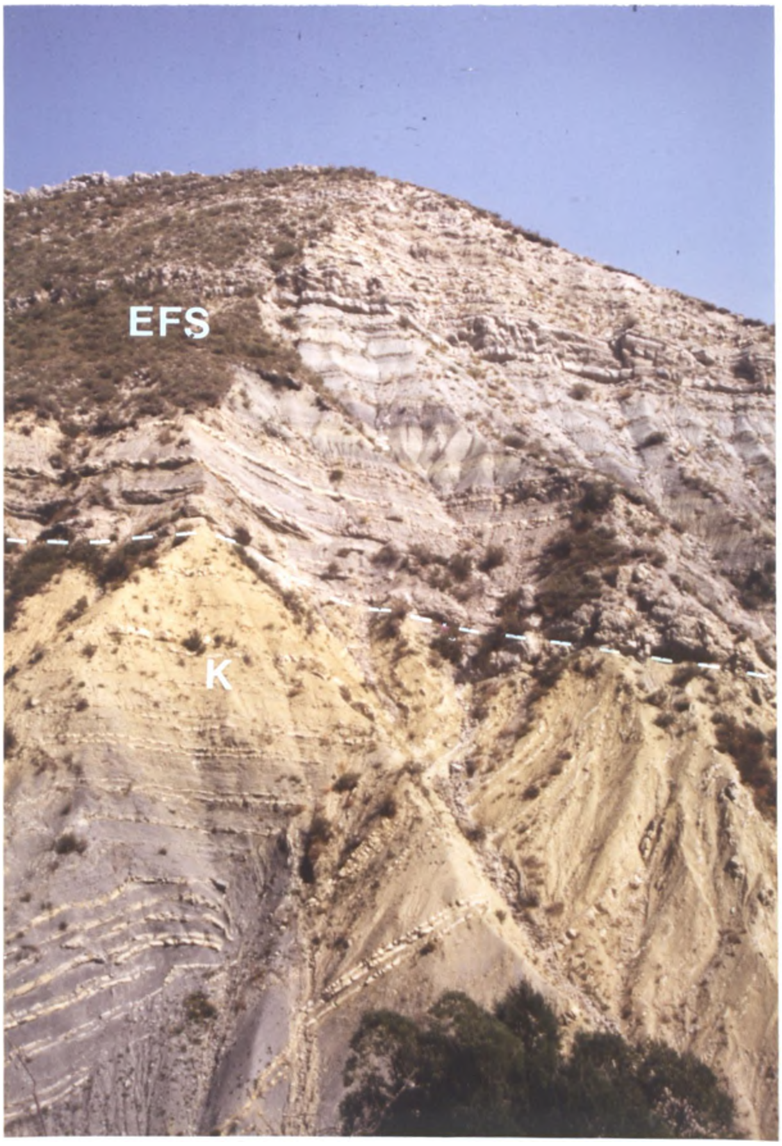


FIG 3.56 Limestone and calcareous mudstones of the Lacustrine facies association, Esclangon fluvial system (EFS), St.Geniez section. The lacustrine deposits unconformably overlie folded Cretaceous (K). Note the low angle of onlap of the lacustrine deposits onto the unconformity.

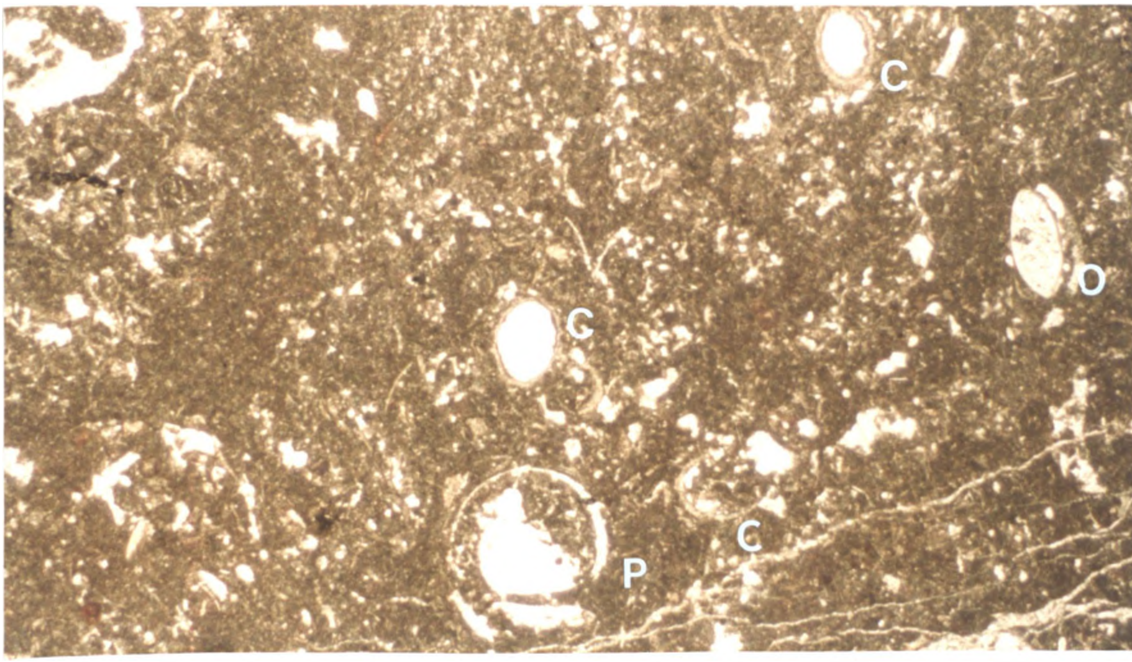
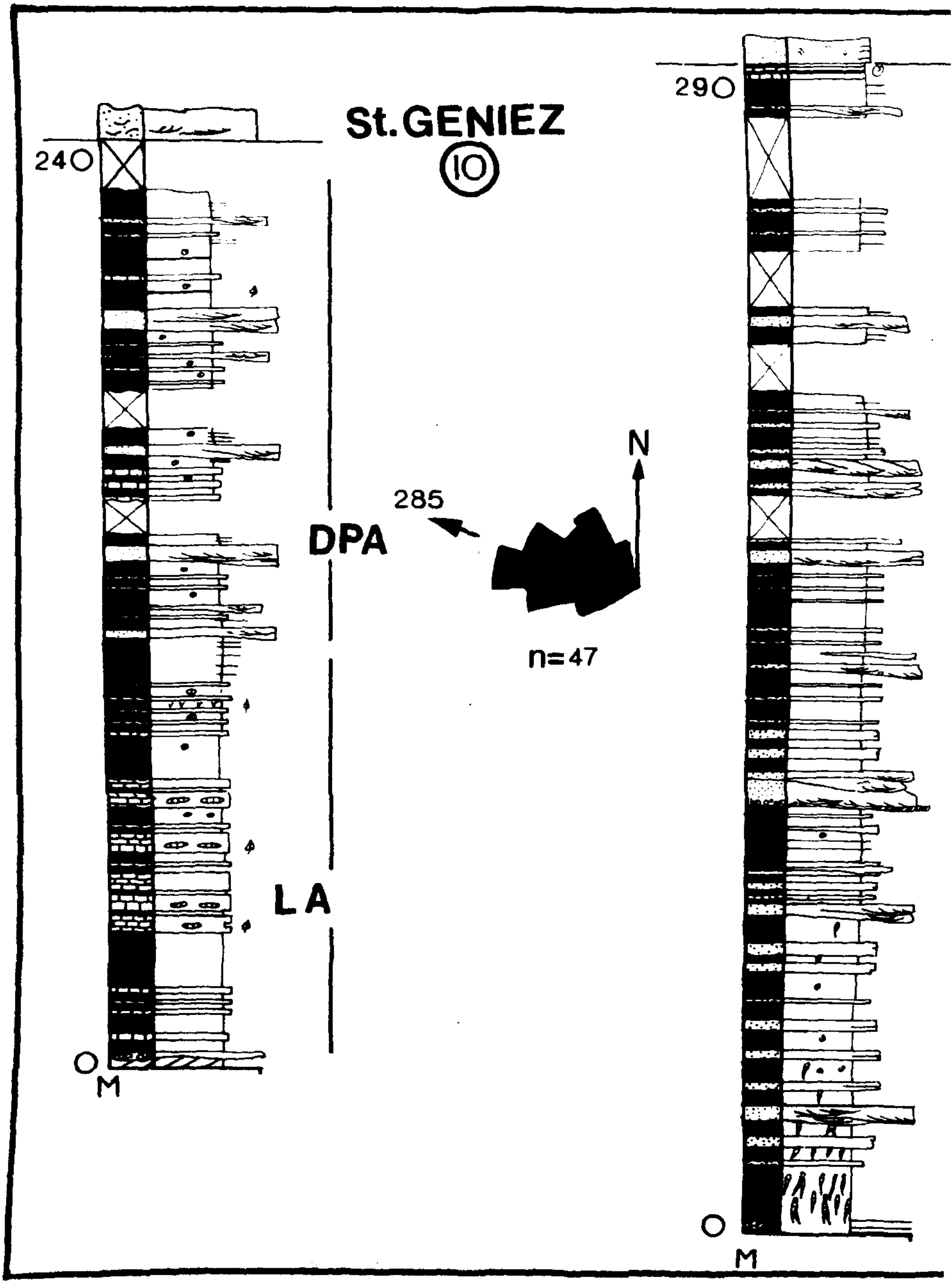


FIG 3.57 Thin section of lacustrine limestone showing micritic texture and presence of *Planorbid* gastropods (P), ostracods (O), and characea (C). X 10

FIG 3.58 Thin section of chert nodule within lacustrine limestone. *Planorbid* gastropods (P) replaced by silica enclosed within micro-crystalline chert (Ch), which replaces micritic limestone (M). Bedding is horizontal. X 10.


St. GENIEZ
10
240
290
DPA
LA
285
N
n=47
0
M
0
M

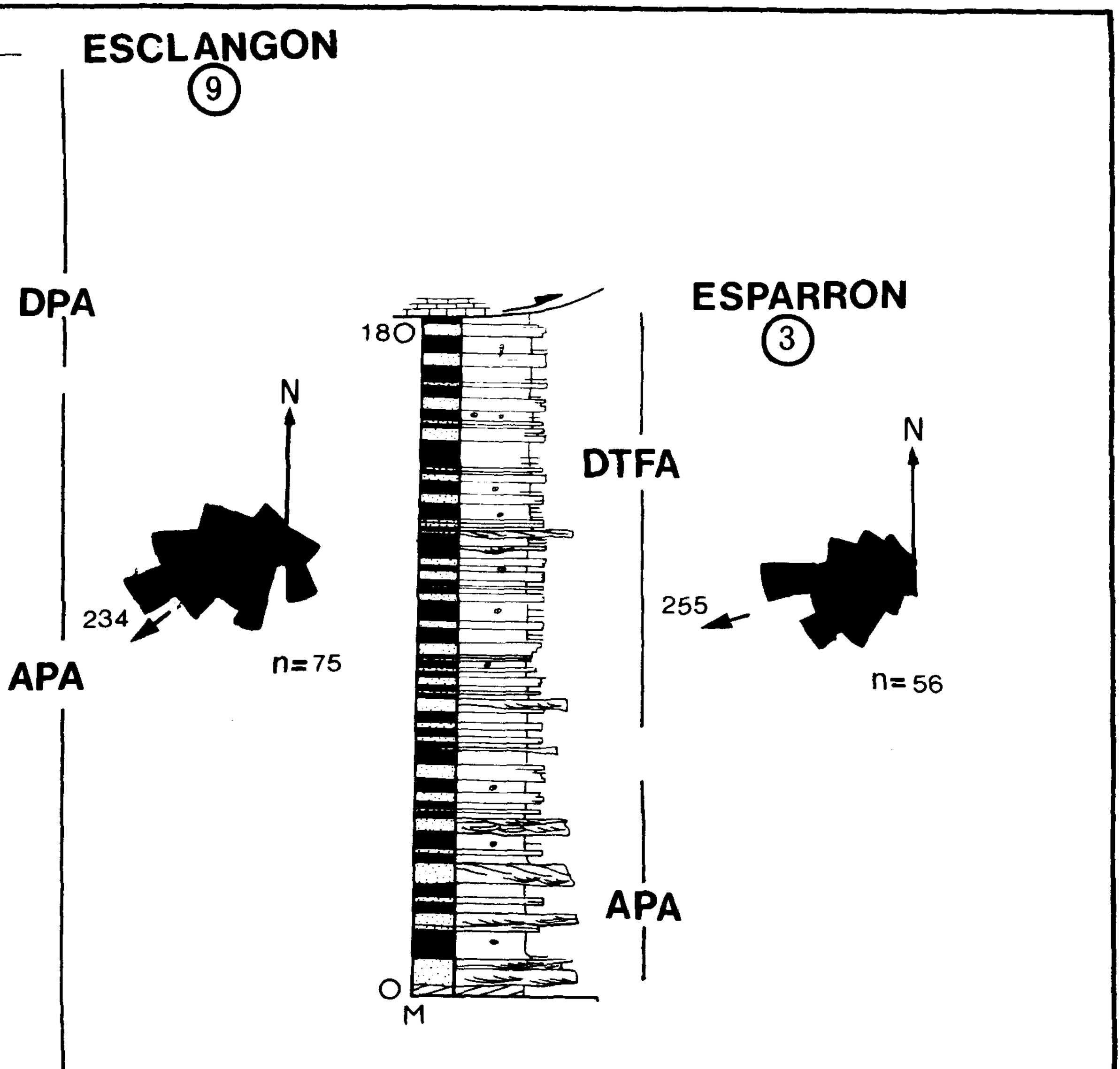


FIG 3.59 Mega-sequences of the Esclangon fluvial system, at Esparron (Marigeneste) (locality 4), Esclangon (locality 9), and St.Geniez (locality10).

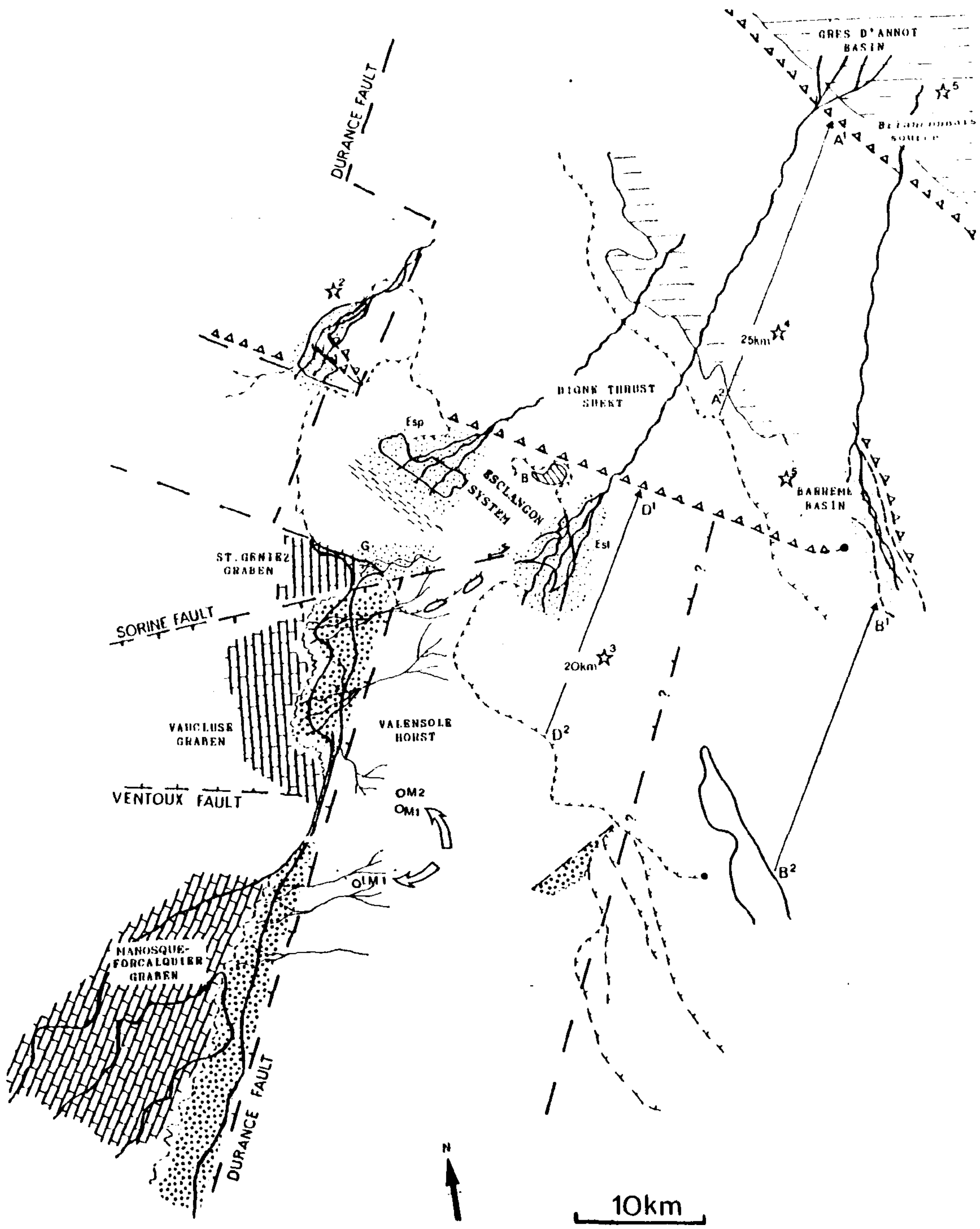


FIG 3.60 Tectono-sedimentary model of the fluvial systems of the Molasse Rouge formation. The Esclangon terminal fan system, a mixed load alluvial system sourced from the uplifted, alpine, Gres d'Annot basins, emerges from the Digne (alpine) thrust front and flows parallel to the extensional fault framework of the foreland.
In the foreland to the alpine thrust belt the Vancon fan is one of a series of small terminal fan systems which develop in the hanging-wall blocks of extensional faults, and which form a part of larger fan-systems fringing the regional extensional faults; in the case of the Vancon fans they form the northernmost part of a terminal fan system fringing the regional Durancian fault (See Fig 4.61). *2: Arnaud *et al.* 1974. *3 - 4: Data of thrust shortening from Graham in Elliott *et al.* 1985. 5:Information from Evans 1985, 1988.
A2->A1 B2->B1: reconstruction of Oligocene (1) position of Annot (A) and Barreme (B) basins.

ALPINE SOURCED CLASTICS
INTRA-BASINAL SOURCED CLASTICS
LIMESTONE
CARBONIFEROUS BASEMENT

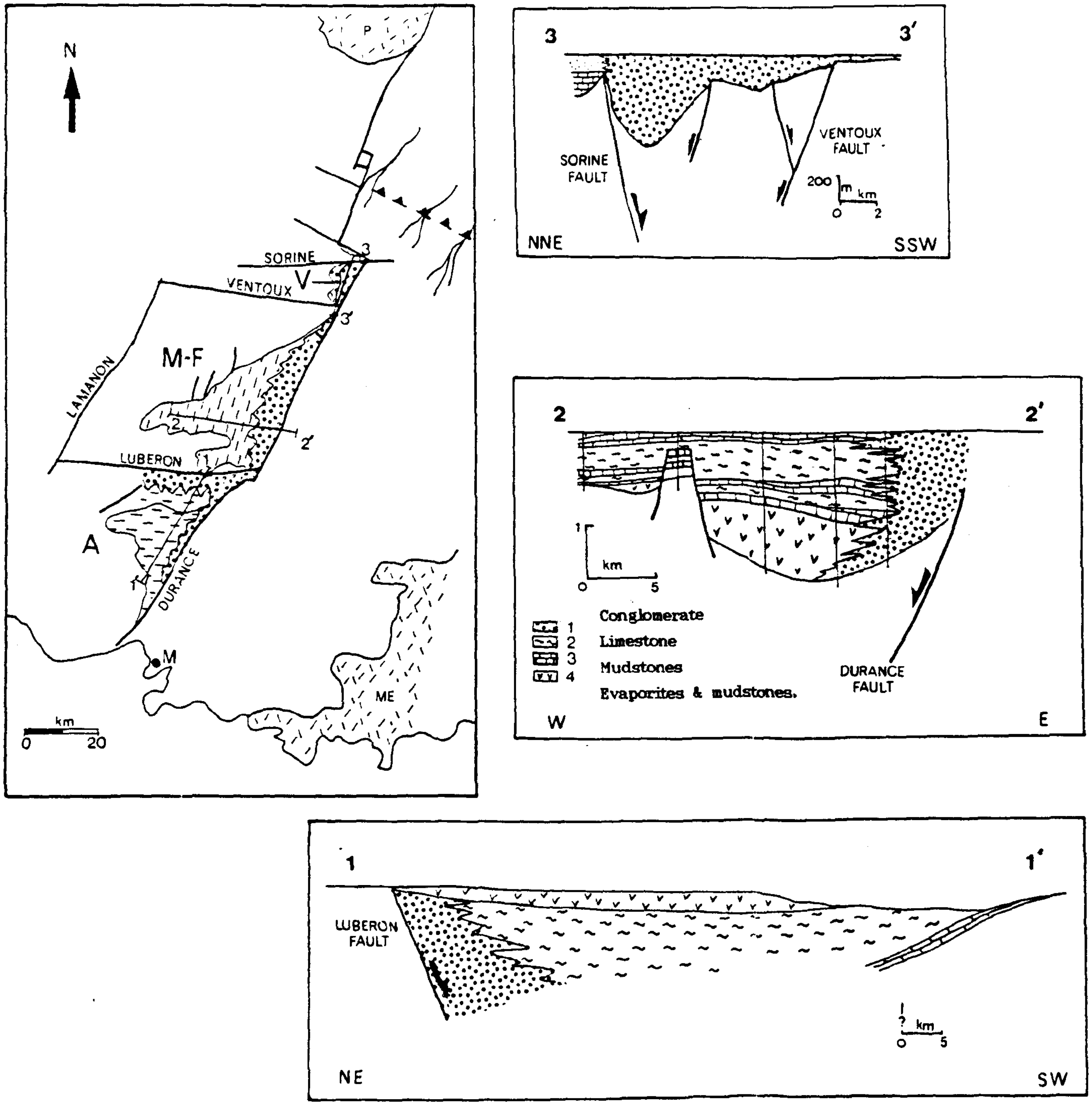


FIG 3.61 Terminal fan systems developed in the hanging-wall to the Durancian fault in the foreland to the S.W Alps. Note how the regional graben developed to the west of the Durance fault is transected by a series of ~ E-W trending high-angle faults (interpreted as transfer faults) which compartmentalise it into a series of discrete smaller basins, the Aix basin (A), Manosque-Forcalquier basin (M-F) and the Vaucluse basin (V). Alluvial fans are primarily sourced off Mesozoic formations in the footwall block to the Durancian fault *ie* intra-basinally sourced, with smaller fans in the Aix basin noted to develop off the footwall of the E-W trending Luberon fault.

1-1' - NE-SW section through the Aix basin (after Cavelier *et al.* 1984)

2-2' - E-W section through the Manosque-Forcalquier basin. (after Cavelier *et al.* 1984)

3-3' - NNE-SSW section through the Vaucluse basin.

Digne-Valensole basin the formation is attributed a Burdigalian - Helvetian age, and a simple two-fold, bio-stratigraphic division, into a Burdigalian, and Vindobonian member after Goguel *et al.* (1964). Beaudoin & Gigot (1971) recognise a similar Burdigalian - Helvetian division to the formation at Auribeau.

The upper limit of the formation has been further constrained in the north-east of the Digne-Valensole Tanaron-Esclangon) where it is given a Helvetian-Tortonian age (Haug 1891, and Beaudoin 1966). The biostratigraphy of the basin has been extended by the recognition in this study of lithostratigraphic members Bm1-Vf2. which are fully discussed in chapter 5.

In the Jabron basin the marine molasse is attributed a Burdigalian - Helvetian age by Goguel (1964).

4.3 Previous Work

The Marine Molasse Formation of the Provence region has not previously been the subject of any detailed sedimentological analysis.

Beaudoin & Gigot (1971) described a lithological section through the formation (Ravine de Bramefan, Auribeau), but concentrated their study on the morphology of current produced sole structures in the sequence, which, together with the discovery of birds foot-print trace fossils, they considered indicative of the deposits of a high energy, beach. Unfortunately their observations are not keyed into detailed graphic logs, making the application of their work to this study difficult.

The most recent study of the sedimentology of the marine molasse was that of Gigot *et al.*, (1975) who analysed the sequences in the north of the Digne basin, and on the basis of bimodal current ripple lamination, considered them to be 'tidalites'.

4.4 Classification of Sedimentary Sequences

The clastic successions of the Marine Molasse formation have been divided into **facies** on the basis of grain size, sand-mud ratio, sedimentary structures, degree of bioturbation, and macrofaunal content (after Johnson 1978). The **facies** recognised within the formation are summarised in Fig. 4.5, and described in more detail as a part of facies associations. After Reading

(1986), these **facies associations** are defined as groups of facies that occur together and which are considered to be environmemtally related. Where these facies associations have a vertical organisation they are termed **facies sequences.** The facies associations of the formation are summarised in Fig. 4.5.

4.5 Petrographic characteristics of the Marine Molasse formation.

The Marine Molasse sandstones of the Digne-Valensole basin and of the eastern part of the Jabron basin comprise moderately to well sorted, siliciclastic sandtones. In thin section these are of 60-80% sub-rounded, quartz (dominantly strained unimodal quartz) grains, with lithic fragments of limestone forming 10-30%, and with lesser amounts of chert and quartz arenite. Glauconite forms an important minor constituent varying from <5-10%. Plagioclase feldspar, muscovite mica and opaque minerals form a minor component.

In contrast, the Burdigalian successions in the western part of the Jabron basin (and also the Apt-Forcalquier basin - Jones *pers. comm*) comprise coarse bioclastic grainstones. These are formed almost entirely of comminuted bioclastic detritus, of bryozoan, bivalve, echinoid, sponge, coral barnacle and gastropod fauna. Quartz is of variable proportion ranging from 0-20% but typically being less than 5%. Minor constituents include micritic peloids, glauconite and opaque minerals, which form less than 5% of the lithology.

4.6 Facies Associations.

4.6.1 Introduction

The Marine Molasse formation predominantly comprises siliciclastic sandstones and mudstones which form a distinctive suite of shallow and marginal marine facies, organised into facies associations and facies sequences of tens of metres thickness. These show a higher level of organisation into transgressive - progradational cycles which range in thickness from 100 - 300m, and which are discussed in Chapter 5. The c ycles typically comprise a thin (dm - m scale) basal, transgressive facies or facies sequence which is succeeded by a thick (tens to hundreds of metres) progradational mega-sequence comprising vertically repeated, progradtional facies sequences and aggradatio nal facies associations of tens of metre thickness. More rarely transgressional megasequences are developed.

The facies associations and sequences of the formation are therefore described in terms of being:

(i) progradational or aggradational facies associations.

(ii) transgressional facies sequences.

4.6.2. Progradational - Aggradational Facies Associations.

4.6.2.1 Shelf Sandwave Facies Association

This facies association comprises coarse bioclastic, grainstone (as described in section 4.5) and is unique in being the only grainstone deposits of the Jabron and Digne-Valensole basins. The predominant structure of the facies association is large scale (0.5-3.0m), sets of shallow trough, cross-stratification, with dm scale set base relief. Foresets are tangential, and dip at angles ranging from 16 - 27'. The foresets are defined by textural variations and are of cm - dm scale. Thin, mm - cm scale, fine grained, well sorted sandstone intervals occur in the foresets with a cm - dm spacing (up to 80 cm), defining packages of foreset lamination, though with no apparent cyclicity.

Erosional discontinuity planes are rarely present and may be mantled by cross-lamination directed up the foreset.
Ripple cross-lamination is rarely present, preserved as toesets to, and as erosional lenses between, the larger sets.

Stratification is predominantly unidirectional, but with a subordinate, opposed mode defined by ripple lamination within the erosive pause planes, and within the ripple laminated co-sets. Cross-stratification is not organised above the level of single sets.

Interpretation

The cross-stratification of this association is typical of that produced by large scale, 3-D bedforms (Harms *et al.* 1982).

The grouping of the foresets by thin, fine grained intervals reflects a pulsing to the flow and is similar to slack water drapes in tidal stratification. The absence of mud or silt grade 'drapes' is interpreted as due to the lack of this grade material in the system. The erosional discontinuity planes record flow reversals and are interpreted as tidal, erosional pause planes (Terwindt 1971). The fine grained and well sorted nature of the bottomsets suggests deposition primarily from suspension.

The low lithic clast content of these deposits is very distinctive, and is interpreted as reflecting the clastic starvation of the depositional environment. Direct fluvial clastic input was obviously not occuring, with the consequence that intrabasinally generated bioclastic material provided the sole clastic source. Biogenic production rates must have been high in a laterally adjacent lower energy section of the basin (see section 5.2.4.1 for discussion).

Large scale tidal stratification may develop in inshore estuarine, or offshore shelf settings (Elliott 1986, Johnson 1986). The facies association lacks most of the diagnostic criteria of inshore tidal settings (see Fig 4.6 for review), and the absence of mud deposits, the abundance of open marine bioclasts, and the predominantly unidirectional character of the stratification, are more characteristic of an open shelf setting.

Tide dominated, carbonate sand facies, are characteristic of sub-tropical shelves, with mud-free deposits more typical of open, rather than protected shelves (Selwood 1986), with the main features of these areas being active sandwaves in the form of linear tidal sand ridges generated by mutually evasive tidal streams (Ball 1967, Sellwood 1986).

4.6.2.2 **Inshore Tidal Channel and Shoal Facies Association**

This facies association comprises siliciclastic sandstone dominated deposits within which, two facies are dominant.

The majority of the association comprises linsen, wavy and flaser bedding of the **heterolithic facies** within which fining upward sequences of the **channelised sandstone facies** (type 1) are developed. This association forms thick aggradational sequences (tens of metres) within which the above facies are randomly interbedded.

Channelised sandstone facies (type 1): This facies comprises large scale channelised fining upward sequences, ranging in thickness from 1 to 6.0m. The sequences frequently erode into each other, forming multistorey channel bodies of some 5-14m thickness.

Channel bases are lagged with silt pebbles, brackish, and open marine bioclasts including *Pecten* and *Ostrea* bivalves. The lower two thirds of the sequences typically comprise well sorted, coarse - medium grained sandstone, with trough, and planar cross-stratified sets 0.3-1m high. Larger scale, trough cross - stratified sets up to 3.5m high, and which may be composed of a poorly sorted, coarse bioclastic sandstone, are occasionally developed directly above the base of the larger scale sequences (Fig 4.10 - 4.11). Heterolithic facies develop as bottomsets to these larger strata (Fig 4.11). Set bases are commonly lagged with marine bioclasts and silt pebbles, with silt pebbles also occuring as pebble trains within foresets.

The upper part of the channelised sequences are dominated by heterolithic cross-laminated bedding, within which subordinate, cross-stratified sets occur. Planar stratified sands form a subordinate facies.

Channel fill sequences typically do not exhibit any evidence of lateral migration, but an example does occur in the Mens section (Fig 4.15) where large scale accretion surfaces, 2.5 m high and dipping at 10° - 16° are developed .

The cross-stratification of this facies is bidirectional and shows the abundant development of silt drapes, and erosional pause planes which define 'tidal bundles' (Figs 4.12 - 4.13; see also Figs. 4.7 - 4.8). Within the channelised facies, a few sets show the development of **tidal bundle sequences** in which tidal bundles, show a cyclicity to their thickness having a periodicity of 28-30 (Fig 4.14).

Bioturbation is low throughout the channelised sequences, being absent from their basal section but increasing to a low - moderate level in their upper parts where *Skolithos*, *Planolites*, *Aulichnites* are common.

Heterolithic facies (type 1): This facies predominantly comprises horizontally bedded linsen, wavy, and flaser current ripple lamination (Fig 4.16 - 4.19). The lamination is predominantly bidirectional, and may show the development of a 'herring-bone' style (Fig 4.19), though more often, lamination shows a preferred asymmetry (Fig 4.18). The bidirectional lamination commonly takes the form of opposed, climbing sets (Fig 4.17). Wave ripple lamination is common, but is subordinate, commonly reworking, and modifying the current ripples (Fig 4.18, and 4.20 - 4.22).

In large exposures the facies may occasionally be seen to form low angle sets, of parallel bedding, dipping at 5-15°, and which extend laterally for up to thirty metres (exposure width) (Fig 4.23).

The heterolithic facies is characterised by random, small scale (dm-m scale), coarsening, and, fining upward sequences (Fig 4.23) which take the form of:

(a) the alternation of linsen and flaser bedding over dm-m scale (Fig 4.25). This may show a cyclicity, with the ripple sets forming the microsequences having a thickness periodicity of the order of 20 - 30. More commonly however no such periodicity is developed and the sequences are best considered as random.

(b) small, dm scale, coarsening-upward sequences passing gradationally from linsen bedding to trough cross-stratification in which the ripple bedding forms the bottomsets to the overlying cross-stratification (Fig 4.24).

(c) Fining upward sequences (cm-dm scale) from an erosive based, sandstone bed (3-15cm), into heterolithic, ripple bedding are also common (Fig 4.32). These beds have sharp, groove cast bases(Fig4.26), overlain by planar lamination, or more rarely hummocky cross-stratification, and passing up into wave, or current ripple lamination, commonly having a climbing form (Fig 4.27 - 4.28). These sandstones are typically separated by heterolithic bedding but may amalgamate and dominate metre scale levels within the facies.

(d) Units (dm-m scale) dominated by planar and ripple laminated sandstone occur within the heterolithic facies. The ripple lamination is predominantly bidirectional, but wave ripple lamination is also common. Primary current lineation within the planar laminated sandstone has a variable orientation. Isolate sets of trough cross-stratification within these sandstone units have a low-angle form, with swept-out toe-sets, and planar laminated topsets.

Bioturbation throughout the heterolithic facies is typically low with the trace fossils *Skolithos, Planolites, Diplocraterion* and *Ophiomorha* burrows) and *Aulichnites* (tracks) present.

Interpretation

This facies association displays all of the diagnostic features of inshore, sub-tidal deposits, namely, bidirectional cross-stratification on a variety of scales, the intimate interbedding of large and small scale stratification, and the development of heterolithic facies (de Raaf & Boersma 1969, Terwindt 1971, and 1981) (see Fig 4.6).
Within the **heterolithic facies** the development of bi-directional climbing cross-lamination is perhaps the most unusual feature of this facies, and is interpreted to reflect the migration of suspension clouds (storm generated ?) by tidal currents.

The low angle sets of heterolithic bedding are interpreted to represent the deposits of a depositional slope, and are analogous to the bedding developed on the inner, accretional margins of estuarine inter-shoal channels (Fig 4.29 from Nio *et al.* 1980).

The small scale sequences (a-d) within the heterolithic facies are characteristic of inshore tidal deposits and reflect fluctuating tidal current strengths (Terwindt 1971, & 1981).
The development of a periodicity of the order of 20-30 to some of the ripple set sequences (type a), suggests that some may have developed under neap-spring tidal cycles (di-urnal regime). This is supported by the development of analagous sequences within the cross-stratification of the channelised sandstone facies (see below this section).

The graded sandstone sheets of this facies (sequence c) are interpreted to be the deposits of waning, turbulent flows, with uni-directional flows generating current ripple lamination, and oscillatory wave currents depositing the wave ripple lamination.
The developement of hummocky cross-stratified beds indicates that

storm generated combined flows (oscillatory and unidirectional) were developed in this setting (see sandstone sheet facies section 4.6.2.6 for fuller discussion). Analagous beds in nearshore and offshore, wave influenced environments (Johnson 1978), including tide dominated, outer and inner estuarine shoal environments (Terwindt 1971, 1981) are interpreted as the deposits of storm generated currents. Their predominance at particular levels within the heterolithic facies suggests that the depositional setting was an open marine one, subjected to variable degrees of storm reworking. Fining-upward micro-sequences involving these graded beds are interpreted to record the waning of a storm event over a few tidal cycles. In a possible modern analogue, Koshiek et al (1986) demonstrated an abrupt increase, and then progressive decrease, in sand transport over an estuarine shoal for the two days during and following a storm (Fig 4.30).

The planar laminated and cross-stratified sandstone units (sequence d) developed within the facies are similar to the deposits detailed from nearshore, intertidal shoals (Reineck & Singh 1980; Jade Estuary, Hayes 1980), shoal bars (Hubbard *et al.*), and high energy, estuarine sandflats (Dalrymple 1985). The planar lamination is interpreted as the product of upper flow regime depostion under either, or both, shoaling waves, and tidal currents under shallow flow depths. The cross-stratification is attributed to tidal flow under deeper (high tide ?) flow conditions.

The fining-upward sequences of the **channelised sandstone facies** are interpreted to be the deposits of tidal channels flanked, by the lower energy, heterolithic, interchannel areas. The large scale cross-stratification of the sequences records deposition from 3-D, and 2-D, sandwaves. Evidence of a marine setting is given by the marine bioclasts and trace fossils, and is supported by abundant evidence for tidal currents in the bidirectionality of the stratification, and the development of erosional and non-erosional pause planes. The tidal bundle sequences having a periodicity of 25-28 are similar to those detailed by Visser (1980) and by analogy are interpreted to have been generated under neap-spring tidal flows in a semi-diurnal tidal regime.

The fining-upward trend of the sequence taken in conjunction with the development of accretion bedding reflects decreasing tidal current strength in the tidal channel during their filling by lateral migration.

The variation in channel thicknesses reflects deposition in channels which ranged in depth up to some 6m. Modern day inlet, and inshore, tidal shoal environments show a similar, spatially distributed, range of channel depths and flow characteristics. Similar scale and range, channelised fining-upward sequences, have beeen described from inshore tidal shoal channels of the southern N. Sea (Oomkens *et al.* 1960, Reineck & Singh 1980, Nio *et al.* 1980, Nummedal *et al.* 1981) and tidal inlet channels of the Atlantic coast (Kumar & Sanders 1974, Greer 1975, Hubbard. et al. 1979).

In summary the evidence described, particularly the development of wave and storm stratification, is considered to indicate that this association developed in an open marine, inshore, sub-tidal shoal and channel environment.

4.6.2.3 **Coarsening-upward Offshore to Inshore Tidal Channel and Shoal Sequence.**

This association comprises a coarsening upward sequence, passing from **muddy sandstone facies** into the sandstone dominated, **heterolithic facies**, wich is erosionally overlain by a **channelised sandstone facies** (Fig 4.31- 4.33). The latter two facies have been detailed in the preceeding section, and the descripion will concentrate on the muddy sandstone facies.

Muddy sandstone facies.

This comprises mud and silt rich, fine grained siliciclastic sandstone which is intensively bioturbated such that its internal stratification is almost completely obscured. At outcrop scale the facies has a massive appearance with a diffuse planar stratification (Fig 4.33). On detailed examination irregular patches of heterolithic bedding, with linsen, wave and current ripple lamination are preserved (Fig 4.34).

Two types of shell layer are common within the facies, occuring randomly with a decimetre to metre (40 - 2m) scale spacing.

Type 1 comprises concentrations of whole or little fragmented shells, which lie parallel to bedding with a convex upward attitude (Fig 4.35). Individual strata range in thickness from a single shell thick to some 15cm and may form composite strata of up to 1.2m thickness. The layers have a diffuse base with no significant increase in grain size relative to the enclosing lithology. The shells comprise *Pecten*, or irregular echinoids, with *Cardium*, and *Pholadomya* shells also present.

Type 2 shell layer is characterised by concentrations of fragmented marine bivalves, gastropods, and echinoids above a sharp erosive base. The bivalves include *Pecten*, *Dosinia*, *Cardium*, *Pholadomya* and*Ostrea*. Strata range in thickness from some 2-25cm, with average thickness being some 10cm. The matrix of the strata is a medium grained sandstone of coarser grain size than the enclosing litholology. The strata are commonly disrupted by vertical cylindrical burrows, commonly containining in-situ *Pholadomya* bivalves (Fig 4.36).

Lateral continuity of shell beds is constrained by outcrop width, having a minimum value of some 300m.

The trace fossils *Diplocraterion*, *Skolithos* together with sub-vertical bivalve burrows containing *Pholadomya*, and *Dosinia* are ubiquitous throughout the facies.

In a variant to the facies, small dm-metre scale (1-2m) units of heavily bioturbated fine grained sandstone are developed within the muddy sandstone. They have gradational bases and tops, and show a coarsening-upward basal section overlain by a fining upward top. Internally they show bidirectional current ripple lamination.

Heterolithic facies: This facies has been described in some detail in the preceeding section, and in these sequences it shows a vertical organisation above a gradational base with the muddy sandstone facies (Fig 4.31). At the base planar, linsen and wavy bedding alternate on a cm-dm (~10-60cm) scale. These heterolithic facies predominantly show bidirectional, current ripple lamination but with wave ripple lamination being common, and with the current ripple sets commonly showing evidence of wave modification (Fig 4.18 & 4.20-.21)

As the sequence coarsens upwards, wavy-flaser bedding becomes dominant.

Centimetre scale (5-15cm), graded sandstone beds with planar lamination and groove casts (Fig 4.32) occur interbedded within the facies with a random, cm - dm scale, and form small scale (5-15cm) fining-upward sequences as detailed in section 4.6.2.2.

The upper part of the main sequence sees a further coarsening upward with the interbedding of dm scale sets of bi-directional, cross-stratification within the heterolithic bedding.

Bioturbation is of moderate to low levels within this facies, showing a progressive vertical decrease in its level of intensity. The trace fossil assemblage comprises *Skolithos, Planolites, Aulichnites, and Diplocraterion* .

Channelised sandstone facies: A channelised sandstone body forms at the top of these sequences which erodes down into the heterolithic facies (Fig 4.3.1). The channel base has 1-4m scale relief over large exposures, with channel bodies traceable for up to 250 m perpendicular to palaeoflow. Channel fills are 3 - 6m thick, and have a gradational fining upward trend as detailed in section 4.6.2. Channel bases are lagged with bioclasts of *Ostrea, Dosinia, Meretrix, Pecten, Cardium, Pholadomya* and an abundance of silt pebbles. The sandstones of the basal section of the channel fill may comprise a coarse, bioclastic sandstone comprising the shells detailed above.

The top of sequences are commonly marked by a progressive, fining-upward passage from the channel facies, through a thin (dm-m) level of the heterolithic facies, back into muddy sandstones. This thin sequence is commonly replaced by a thin bed of heavily bioturbated, sandstone.

Interpretation

The **muddy sandstone facies** reflects low energy conditions and slow sedimentation rates in which bioturbation is of such a high rate that biogenic structures overprint physical sedimentary structures.

The concave upward orientation of shells within Type 1 shell layers reflects their hydrodynamically stable position under the influence of waves or currents (Clifton 1971). Negligible transport is interpreted to have occured during deposition given the low degree of abrasion of the shells, and the continuity in grain size of the matrix with the enclosing deposits. Analogous coquinoid sandstones have been described as 'swell coquinas' by Brenner & Davies (1973), their formation being attributed to wave swell reworking of the substrate producing concentrations of epifaunal bivalves.

The erosive based type 2 shell layers are indicative of deposition from a turbulent, waning flow, with the abraded nature of the bioclasts indicating that they have undergone considerable transport. The predominance of vertical bioturbation through the strata, is interpreted to represent the shifting of epifaunal suspension feeders to the new sediment-water interface following the rapid deposition of the coquina. Brenner & Davies (1973) have described similar shell layers from shelf deposits, and attributed them to deposition from offshore directed, storm currents.

The trace fossil assemblage is of a softground, nearshore marine environment (Ekdale *et al.* 1986) developed in shoreface and tidal flat settings, with the abundance of Ophiomorpha and Diplocraterion/Arenicolites burrows typical of the latter setting (Evans 1965, Warme 1969).

In summary, low wave and tidal energy prevailed in this depositional setting, with the presence of the swell lag coquinas and small wave ripples indicating that storm waves were periodically effective. Analogous deposits have been described from the 'Upper Offshore' facies of the low wave energy Sapelo Island (Howard & Reineck 1972), and the 'Offshore' facies of the high wave energy Californian coast (Howard & Reineck 1981).

The developement of metre scale units of current-ripple laminated sandstone within the offshore facies indicates that low flow regimes were periodically established in the offshore, with the bidirectional nature of the lamination indicating that these were tidal currents.

The **heterolithic facies** characterised by bi-directional cross-lamination and rapid lithofacies changes is typical of the deposits of inshore tidal environments (Oomkens & Terwindt 1960, Terwindt 1971 & 1981, Reineck & Singh 1980), and has been interpreted as such in the preceeding section.

The graded sand sheets were deposited under a waning, storm generated, current, with the sandstone sheet to heterolithic, fining-upward micro-sequences, interpreted to reflect a waning of the storm event over a few tidal cycles (see Fig 4. 30 from Koshiek et al. 1986)

The **channelised facies**, as discussed in the preceeding section, are analogous to inshore, and estuarine channels of the southern N. Sea (Oomkens *et al.* 1960, Reineck & Singh 1980, Nio *et al.* 1980, Nummedal *et al.* 1981) and estuarine, inlet channels of the Atlantic coast (Greer 1975).

In summary the gross coarsening-upward sequence is considered to reflect a progressive increase in the proportion of tidal facies and of the strength of the tidal current regime. A comparable large scale coarsening-upward sequence has been generated in a progradtional model of the tidal inlet shoal of Ossabaw Sound (Greer 1974: see Fig 4.37).

A schematic summary of the interpreted depositional settings of the facies of this association are detailed in Fig 4.38.

4.6.2.4 **Upper Estuarine Facies Association**

This facies association involves repeated fining upward sequences from a **channelised sandstone facies** (type 2) into heterolithic, and mudstone dominated facies (Fig 4.39 - 4.42).

Simple channel-fill sandstone sequences are of 1.5 - 5m thickness, and may form multistorey bodies of up to 8m thickness (Fig 4.39 - 4.40). The sandstone bodies have tabular geometry and are laterally persisent across outcrops of up to some 50m width (Figs 4.39 - 4.40).

The sequences have a sharp erosional base with a thick lag of intraformational silt pebbles, *Ostrea* and *Crassostrea* valves and wood material, as well as extraformational pebble and cobble. Metre scale slump blocks occur at some channel bases (Fig 4.41). The basal section of channel fills are dominated by trough cross-stratified sets (20-40cm) of medium grained siliciclastic sandstone. The sets have thick, silt pebble lags and internally show the development of silt drapes, erosional pause planes, and tidal bundles (see Fig 4.7). This cross-stratified facies pass fines upward into an association of current ripple stratified sandstone and heterolithic bedding.

In flow perpendicular sections, channel fills exhibit large scale sets of accretion bedding dipping at 10 - 16° into the channel (Fig 4.39-4.40). This bedding may be laterally interupted by large scale erosion surfaces having a low angle, convex upward, form. The accretion bedding passes up dip, out of the channel, into horizontal or shallowly dipping heterolithic facies (Fig 4.40).

In a variant to the channel sequences, the upper section of the fill may be dominated by planar laminated sandstones (Fig 4.42 (i). Bidirectional current ripple lamination, and small scale trough cross-stratification, commonly having a low-angle 'swept-out' form, are interbedded within this laminated sandstone. Bidirectional climbing ripple sets and small sets of wave ripple lamination may also occur.

Bioturbation is extremely low with the occasional *Planolites* and *Skolithos* in the upper part of the channel.

Stratification within the channel fills is bidirectional, but typically with a predominance of a particular orientation.

The overlying **heterolithic facies** (type 2) shows a fining upward trend passing from wavy-flaser to linsen bedding over some 1 - 4m (Fig 4.40 - 4.42). Ripple stratification is bidirectional, commonly displaying a herringbone pattern, with sets occasionally having small basal groove casts. Small wave ripple sets are interbedded within the heterolithic facies. Thin graded planar laminated sandstones with current and, less frequently wave ripple laminated tops occasionally occur. Interference ripple sets and runoff modification to some ripple sets is apparent on bedding plane exposures. Comminuted plant debris commonly defines laminations within the ripple sets.

Thin, monospecific shell layers of *Turretelid* and/or small *Cardium*(<0.3cm) occur. Thin bioclastic rich sandstone strata containing fragments of *Pecten, Ostrea, Dosinia, Cardium, Turretelid* and echinoid also occur.

Small scale, heterolithic channel fills (0.5 - 1.5m thickness), with *Ostrea* shell lags are present within the facies (Fig 4.47). The fill may either concordantly drape the channel or form inclined stratification dipping into the channel at some 10–14°. Shallow channels only 20 - 40cm deep and a few m's width also occur. The fill of these channels is dominated by bioclastic debris of *Ostrea, Cardium, Turretelid* in a structureless silty sand matrix.

Bioturbation within the facies is moderate with *Ophiormopha, Diplocraterion, Skolithos* and *Planolites* recognised. Arthropod tracks are seen on the upper surface of sand strata.

An overlying **mudstone facies** is marked by the development of palaoeosol mottling (Fig 4.40 - 4.42). It consists of pale grey (N6), mudstone with linsen ripple lamination at the base, which fines up into mudstone with, thin streaks of fine sand. Thin, graded sandstones containing fragments of cardium and turretelid also occur. The muds have an ochrous mottling (10YR714 - 10YR616) with the development of paleosol profiles with small calcrete glaebules. Thes may achieve the maturity of calcrete tubule or coalesced glaebular horizons.

This facies is transitionally overlain by black (N3 - N1), organic rich, mudstones, of the **black mudstone facies.** Physical sedimentary structures are rare with mm thick silt streaks and the occasional cm thick graded siltstone layer. The mudstones contain whole and fragmented *Planorbid* shells and occasionaly whole large oysters (*Crassostrea*). The muds contain palaeosol profiles (gley) at specific horizons, their degrees of maturity varying from isolate glaebular to coalesced glaebular form. Also presnt in this facies are thin (cm - dm scale) pale grey, (N6) micritic limestone beds with a nodular overprinting, and with a rich *Planorbid* fauna.

In a variant to this sequence, the heterolithic facies is overlain by muddy sandstones of the **brackish bay facies** (Fig 4.39) (described in detail in section 4.6.2.5) as opposed to the mudstone facies.

Interpretation.

The bidirectionality, pause planes and tidal bundles of cross-stratification of the channelised facies is consistent with their interpretation as the fills of tidal channels. Rotated slump blocks are common features in the tidal channels of modern day intertidal flats, where they develop by gravitational collapse of channel margins during emergence at low tides (Reineck & Singh 1980). The development of a brackish marine fauna and woody material lag is characteristic of estuarine channel fills.

Fining-upward, channelised, tidal sequences with point bar deposits capped by muddy sediments are considered by Barwis (1978) to be characteristiuc of estuarine tidal channel fills. Analogous tidal channel fills showing sandy and clayey levees passing out into tidal mud flats levee development are descibed from estuarine intertidal flats on the Netherlands coast by Roep (1988). Erosional surfaces within the point bar accretion bedding probably reflect erosion during exceptional discharge conditions (see Bridges & Leeder 1976)

The predominance of planar laminated sandstone at the top of some of the sequences, records that upper flow regimes were attained by tidal currents during the late phases of channel fill. Barwis (1978) noted that in the intertidal section of estuarine channel point bars, megaripples became washed out into planar bedding. Analogous planar sandstones have also been described from high energy estuarine sand flats of Cobequid Bay (Dalrymple *et al.* 1985).

The succeeding **heterolithic facies** are interpreted to be the deposits of shallow subtidal - intertidal, mixed flats (Evans 1965, Reineck & Singh 1980; see Fig 4.48). Low energy tidal conditions are attested to by the fine grained nature of the deposits and the commonly starved form of the ripples. Groove casts on the bases of the ripple sets may have been generated by tidal currents. As detailed by Allen (1965) erosional furrows in cohesive mudstone may be generated by flows having a mean velocity of > 20 cm/sec, which falls well within the stability field (25 - 60 cm/sec) of current ripple sets (Harms *et al.* 1982).

The thin graded bioclastic strata in this facies probably developed under high tidal, or storm conditions with transportation of coarse material out of the tidal channels and onto the mud flats. In contrast the monospecific faunal concentrations probably reflect the in-situ reworking and concentration of epifauna by wave action during storms.

The small heterolithic channel fills are interpreted to be ths fills of small, meandering tidal channels / gullies on the tidal flats. Reineck & Singh (1980), and Bridges & Leeder (1976) detail similar channel fills from estuarine tidal mudflats. The shallow bioclast filled channels appear analogous to the tidal drainage scours described by Evans (1965) in his 'Higher mud flat zone'.

The **mudstone facies** with its lack of fresh water fauna, but extensive pedogenic modification, is characteristic of the deposits of supratidal salt marshes (Evans 1965, Reineck & Singh 1980, Elliott 1986; see Fig 4.44). The thin bioclastic strata were probably introduced into the facies during flood events e.g. storm events, spring tides.

The gastropod fauna and high organic content of the **black mudstone facies** is indicative of its development under a fresh water body, whilst the oyster fauna are characteristic of a brackish environment with salinities of some 24 $^{0}/_{00}$ (Hudson 1963). The palaeosols are interpreted as gley soils which develop under high, but fluctuating, groundwater levels (Buurmaan 1983). Periods of relatively high carbonate content, or low clastic deposition rates permitted the deposition of freshwater limestones. This facies is therefore interpreted as the deposits of shallow, ephemeral, and periodically vegetated, ponds or lakes.

In summary, the facies of this association show all of the characteristics of a wave protected, upper estuarine environment, with the fining-upward sequences attributed to the lateral migration of estuarine channels across intertidal flats.

The variant to these sequences which involves the brackish mudstones is interpreted to record the passage from an estuarine channel to a shallow brackish, (lagoonal ?) bay environment. Such environments laterally coexist in protected tidal settings, and these sequence may have been simply generated by the lateral migration of facies belts.

4.6.2.5 **Coarsening-upward, Lagoonal Bay to Flood Tidal-delta Sequence**

These coarsening-upward sequences are of up to 12m thickness, and involve the vertical passage from a **brackish mudstone facies** into a **channelised sandstone facies** (fig 4.45 - 4.46).

The **brackish mudstone facies** at the base of the sequences comprises a fine, grey mudstone with a monospecific epifauna of *Turretelid* gastropods (Fig 4.47). These may be concentrated into thin shell layers with concentrations of *Ostrea* bioclasts also being present. Finely comminuted plant debris and larger woody fragments are also common in the mudstone.

The mudstones coarsen upwards, passing into intensively bioturbated muddy sandstones with relict patches of bidirectional linsen stratification and wave ripple lamination (Fig 4.48) . Thin, graded coarse sandstone beds with bioclastic fragments of Ostrea, Cardium, Dosinia, Meretrix, Turretellá are interbedded (Fig 4.49). Bioturbation of the facies is by *Ophiomorpha, Diplocraterion* and *Skolithos* .

This facies passes gradationally up into a **heterolithic facies** (type 3), which at its base comprises linsen and wavy-flaser current ripple bedding (Fig 4.50). Ripple lamination is bidirectional. Centimetre scale graded sheet sands, with planar lamination, current ripple laminated tops and marine bioclast lags are randomly interbedded.

The upper section of the facies comprises an association of medium scale (10 - 20cm thick), trough and planar cross-stratified siliciclastic sandstones, and thinly bedded heterolithic facies. The sets of cross-stratification commonly have bioclastic and silt pebble lags, siltstone drapes, and erosional pause planes. The bioclasts are of *Cardium, Ostrea, Dosinia, Meretrix, Pecten, Turretella*. Stratification has a predominant orientation (landward), although opposed (seaward) sets do occur.

The degree of bioturbation decreases vertically within the facies with *Ophiomorpha,Planolites, Skolithos* and *Diplocraterion* present at the base of sequences.

In flow parallel exposures, the internal organisation of the heterolithic facies is apparent (Fig 4.45 - 4.46). The facies is deposited on large scale, inclined surfaces which pass down dip into the **brackish mudstone** facies. These surfaces dip landward at 8 - 15 °. Individual strata are of 10 - 50 cm scale (Fig 4.50), and of sigmoidal form, with a lateral extent of some 60m from their proximal to distal terminations. The bottomsets of this stratification comprise the heterolthic deposits, with the 'foresets' or upper part of the stratification comprising the cross-stratified dominated part of the heterolithic facies.

The inclined strata are arranged into sigmoidal units, of tens of metres' width, by major sigmoidal erosion surfaces (Fig 4.45). These surfaces are erosional in their upper part passing down dip into conformity with the bottomsets. The inclined strata may be traced passing up-dip into sandstone channels of the **channelised sandstone facies** where, it becomes apparent that the major erosion surfaces defining the sigmoidal units pass into erosive channel bases. A series of vertically and laterally stacked sigmoidal units, offset northward (landward) may be seen within a coarsening- upward sequence at St.Symphorien (Fig 4.45).

The top of these coarsening-upward sequences is marked by the passage into an erosive based **channelised sandstone facies.** The channels are traceable across outcrop for several hundred metres having m scale erosional relief. Complete fining-upward channel fill sequences are 3 - 5m thick and show a transition from a basal section dominated by trough and planar stratified sets (20-40cm) to an upper section of heterolithic deposits or planar laminated sandstones (Fig 4.46 (iii)). Sets have high concentrations of silt pebbles at set and foreset bases, and also contain woody material and plant material ('coffee grounds'). They also show bi-directional stratification, siltstone drapes, erosional pause planes and the developement of tidal bundles. In the occasional example tidal bundle sequences are developed (Fig 4.51) in which the bundles have a thickness periodicity of the order of 25-30.

In flow perpendicular sections the channels have large scale, lateral accretion bedding of up to some 4m height and 60m lateral extent, dipping at some 10 -16° (Fig 4.45 - 4.46) Channels have a basal lag with *Ostrea* shells admixed with more open marine bioclasts.

The channel fills show a clear predominance of onshore directed palaeocurrents.

Interpretation.

The faunal assemblage of the basal mudstones of these coarsening-upward sequences is indicative of brackish bay environment (Park, 1959), with *Ostrea* being tolerant of reduced salinities of some 24‰ (Hudson 1963)). The coarsening upwards within the mudstone is interpreted to record inreasing proximity to the overlying flood delta, with sandstone being supplied to a bay environment from a flood tidal delta, and also perhaps storm washover strata.

The coarsening-upward character of the **heterolithic facies** together with its develop ment of large scale sigmoidal dipping surfaces are considered characteristic of its developement as a delta body. The bidirectionality of the stratification of the facies is interpreted to reflect its tidal generation, with the landward dominated orientation of the stratification characteristic of a flood tidal delta setting (Fig 4.52 from Hayes 1980).

The vertically increasing scale of stratification, grain size, and the proportion of sandstone in these vertical sequence is interpreted to reflect increasing tidal current strength. The linsen facies at the base of the sequence reflects conditions in which the maximum tidal current velocities were typically of the order of 0.45 ms^{-1} (Terwindt 1981). Interbedded graded sheet sands reflect periodic higher energy waning flow events, probably developed during exceptionally strong tides or, storm events.

The upper part of the heterolithic facies sequence records the development of 2D and 3D sandwaves.

The tidal nature of the fill of the channel bodies at the top of the sequence is apparent from the bidirectionality of the strata, and, the abundance of reversing flow structures. Comparable lateral accretion bedding has has been detailed from the inner margins of estuarine tidal channels (Nio *et al.* 1980). McCrory *et al.* (1986) described a similar channelised, tidal estuarine facies with sand dominated lateral accretion bedding, from the Milk River formation, Alberta.

The position of this channel facies overlying the flood delta facies clearly indicates a close link between them, which is proven by the lateral coexistence of the facies in certain sections. Analogous channel sequences, dominated by flood orientated cross-bedding, are developed in inlet mouth, flood channels which pass landward into lagoonal tidal-deltas (Fig 4.52 from Hayes 1980). By analogy these channelised sequences are considered to be inlet mouth flood chanels, which acted as the feeder channels to the flood tidal-deltas.

The proximal-distal (or vertical in facies sequence) tidal velocity gradient, apparent in this sequence, is analagous to that detailed from modern flood tidal deltas.

Hine (1975) detailed that in the proximal, intertidal portion of flood deltas in the mesotidal Chattam Harbour Estuary, where tidal currents achieved maximum velocities of 0.85 ms^{-1}, a flood dominated association of 'linear and cuspate megaripples' and current ripples developed. Velocity studies on the intertidal proximal portion of flood tidal deltas of North Inlet, S.Carolina recorded maximum tidal velecities of some 0.80 ms, and the develop ment ofa facies association dominated by ripple sets, but with 'megaripples' and 'sandwaves' also present.

Wedge shaped, tidal sandbodies analagous to this facies are described from the Pano formation, S. Pyrenees (Cuevas Gozola (1985). The formation comprises a sandstone body divided into a series of laterally, and vertically, stacked wedge shaped units, interpreted to be flood tidal delta deposits, overlying lagoonal deposits. The stacked nature of the wedge units was considered to reflect 'pulsed' phases of expansion of the delta into the lagoon. Analogous wedges, or sigmoidal units detailed from this facies association may also reflect such 'pulsed' phases of flood delta expansion. Alternatively each wedge might represent a separate lobe within a multi-lobate, flood delta as described by Hine (1975) from the Chattham Harbour Estuary. Lobes prograde at different rates, and with lateral shifting of the tidal feeder channels the delta lobes are subject to periodic shifting or abandonment.

Analysis of the Shoreface facies of this formation (section

4.6.2.6) indicates that the wave energy levels were those of a low - moderate energy shoreline, whilst analysis of the inshore tidal shoal facies (section 4.6.2.2) suggests a mixed tide-wave influenced inshore system. This contrasts with the wave dominated shorelines on which flood tidal-deltas characteristically develop (Hubbard *et al.* 1979). A transgressive shoreline model (after Morton & Donaldson 1973: see also section 5.2.3.6 for supporting evidence from basin-wide sequence analysis) is thus envisaged for the development of the flood tidal delta facies sequences. During transgressive periods shoreface erosion of barrier island deposits is envisaged to have resulted in an increase in the sediment supplied to the seaward margin of the inlet by longshore drift, and subsequently through the inlet by wave augmented flood currents causing the developement and landward migration of large flood-deltas.

4.6.2.6 **Offshore to Foreshore Facies Sequence**

These coarsening-upward sequences consist of the gradational, passage from the **muddy sandstone** facies into a heterolithic **sandstone sheet facies** and finally into the sandstone dominated, **planar sandstone facies** (Fig. 4.53). The planar sandstone facies also forms thick (tens of metres) 'aggradational' sequences rather than coarsening-upward sequences.

The **muddy sandstone facies** has already been described in detail in section 4.6.2.3.

The **sandstone facies** in the middle part of the coarsening-upward sequences comprises erosive based, fine-grained siliciclstic sandstone beds with a sheet-like geometry, 5 - 30 cm thick and at least 50m wide (outcrop width). Toward the base of the sequence these beds are interstratified with the muddy sandstone facies whilst higher in the sequence the beds become interstratified with heterolithic, wave- and current- ripple laminated beddding

The sandstone beds are graded and have groove cast bases above which planar or hummocky stratification is developed. The planar stratified beds show the simple passage from planar lamination into current or wave ripple lamination, commonly in the form of climbing sets.

The hummocky cross-stratified beds, range in thickness from some 10–30cm with hummocks showing a range of wavelengths from some 0.10–1.0m, typically being some 0.20–0.60m. The beds show a vertical organisation with hummocky cross-stratification at the base of the bed passing up into planar lamination with a wave or current ripple laminated top. The laminae of the hummocky stratification are defined by alternating light (coarse) and dark (fine, mica and plant debris rich) layers.

Bioturbation throughout the facies is characteristically by *Diplocraterion* with vertical *Skolithos* also common.

The **planar sandstone facies** at the top of these coarsening - upward sequences has a gradational base, and in its lower part comprises some 30 - 50% planar lamination. Interbedded are planar cross-stratified sets (<10% of the facies) 5 - 15 cm high. These sets commonly have low angle foresets and planar topsets with p.c.l. The sets are interpreted to be directed onshore, though with a high degree of variance.

Thin (2 - 5 cm) graded sand sheets, with planar lamination, and more commonly, truncated hummocky cross-stratification forms some 5 -10% of the facies, with the latter occuring as isolate sets, or cosets up to 1.8 m thick.

A variation to the previously described hummocky cross - stratification occurs in this facies in which a wave ripple laminated basal section is developed. This shows a vertical sequence passing from symmetrical, trochoidal ripples into progressively larger wavelength, and amplitude, sinusoidal ripples which are erosively overlain by hummocky cross-stratification (Fig. 4.54).

In coarsening-upward sequences the upper part of this facies is dominantly of planar lamination (some 70–80%) (Fig. 4.55 - 4.56). It shows a well developed grain size sorting and the development of light (mica poor) and dark (mica rich) mineral layering. In detail it forms decimetre scale, sets of parallel lamination defined by low angle erosion planes (<5°) and traceable for tens of metres across the exposure. The upper surface of these beds are commonly reworked by symmetrical and asymmetrical, wave ripples (Fig. 4.55 - 4.56). Symmetrical wave ripples are of variable dimensions with wave lengths ranging from 5.0 - 22.6 cm and amplitude from 0.2 - 2.3 cm and are predominantly of sub-vortex type (Allen, 1979).

Asymmetric wave ripple sets are interpreted as being directed onshore with the sets showing ranges in their orientation of the order of 20 - 50^0. Interference ripples are present. Primary current lineation is ubiquitous within the planar laminated sandstone and is perpendicular to interstratified wave ripple

Thin veneers of the convex-upward shells of the bivalves *Dosinia* and *Cardium* are also common on the the upper surface of beds (Fig. 4.57).

Shallow swaley cross - stratification (Leckie & Walker 1982) is also recognised within the facies, developed above scours of cm-dm (up to 35cm) depth and dm-m (up to 2m) width, and low angle slopes of some 10-15^o (may attain angles of 28^o) (Fig. 4.59). Scour surfaces become planar outside the swale passing into conformity with planar lamination. The swaly scour is draped concordantly by sandstone laminae which thicken into the swale, progressively flattening its topography and passing into planar lamination. Swaley stratification may occur isolated within planar laminated sandtone, or may erosively amalgamate to form co-sets (amalgamated hcs) of 0.30-1.2m thickness (Fig. 4.59).

In a few examples, swales show an elongate, gutter-like form (plan view), and a fill which has a low angle of dip (6-10^o) down the scour axis. The axes of these gutter scours are orientated sub-parallel to the orientation of wave ripples in the facies, and are interpreted as being directed offshore.

Finally, two examples of steep-sided channels, 1 - 2.0m thickness, and tens of metres width were found within the facies (Fig. 4.61). These show fining upward fills passing from a thin basal interval of trough cross-stratification into wave rippled sandstone. (Fig. 4.61). The channel axis, and the trough stratification are directed offshore with the overlying ripple sets orientated oblique to the channel axis.

There is a low level of biogenic reworking to the facies with *Diplocraterion* (Fig. 4.56), *Skolithos* and *Thalassinoides* burrows present. One remarkable example of a wave rippled surface bearing numerous bird footprints is also present (Fig. 4.60).

Interpretation.

The **muddy sandstone** facies at the base of the sequence is interpreted as low energy, offshore deposits (see section 4.6.2.3).

The overlying **sheet sandstone** facies records the alternation of low (fairweather) and high (storm) energy conditions, with the heterolithic beds reflecting wave ripple formation during fair weather conditions.

During high energy conditions graded sand sheets with hummocky cross-stratification are interpreted to have been deposited from combined flow, storm generated currents flowing offshore, perpendicular to oblique to the shoreline as given by wave ripple orientation. The stratification sequence of the hummocky cross - stratified beds is comparable to the idealised, storm-fair weather Hummocky-stratification sequence of Dott & Bourgeouis (1982). The alternation of heavily bioturbated wave rippled muds and silts with the coarser storm sand beds at the base of the facies is characteristic of the Offshore - Transition zone facies on wave dominated shorelines (Howard & Reineck 1981; Elliott 1986 for review).

The predominance of parallel laminated sandstone in the **planar sandstone facies**, particularly its upper part, is characteristic of deposits from the shoreface to foreshore zones of many wave dominated shorelines (Howard & Reineck 1987).

The interbedded cross-stratification is interpreted to be the deposits of 2-D dunes generated under shoaling wave action on a shoreface, as documented by Clifton (1976).

The hummocky cross-stratified beds with a wave ripple laminated basal section are interpreted to record the progressive build-up of a storm event (see Allen's (1980) storm model), accompanied by rapid suspension fall-out. The transition from trochoidal to sinusoidal ripple is analogous to that detailed by Allen (1981) and reflects the passage from the vortex to 'post-vortex' ripples, with the coincident increase in the wavelength of the ripple sets suggestive of a simultaneous increase in the wave orbital diameter. The symmetrical nature of the ripples indicates that a unidirectional current was not generated during the early phase of the storm.

Comparison of the lower part of this facies with modern day shorefaces suggests that it is transitional between low and high energy types. On low wave energy shorefaces (Howard & Reineck 1972), (Howard & Reineck (1981) intermediate scale cross stratification is absent (unlike this facies), whilst high wave energy shorelines (Clifton & Hunter 1971, Howard & Reineck 1981, Elliott 1986) are characterised by a high proportion of dune deposits.

The channelised sequences within the facies are interpreted to be have resulted from scouring by an offshore directed, turbulent current which deposited trough-stratification during its waning phase. The fine grained upper part of the channel fill reflects a return to lower flow regime (fair weather) conditions with the abandonment and infilling of the channel. These channels are interpreted to have been developed during storm conditions when shoreline set-up produced enhanced rip-currents, or, geostrophic currents which eroded the upper shoreface.

The predominance of parallel lamination at the top of the facies is characteristic of the deposits of foreshore zones (Elliott 1986), and the trace fossil assemblage, particularly the birds footprints, support such an interpretation. The planar lamination with primary current lineation and shear sorted mineral layering are characteristic of upper flow regime bed transport during wave generated swash - backwash flows (Clifton 1969). Analogous parallel laminated facies are developed in the Upper Shoreface of the low wave energy Sapelo Island (Howard & Reineck 1972), where similar shell layers are attributed to storm reworking of shoreface shell material.

After Leckie & Walker (1982), the swaley stratification is interpreted to be the shallow water equivalent of hummocky cross - stratification. The elongate 'swaly' scours bear close similarity to the 'gutter cast' structures described from ancient sequences by Cant (1980), Kreisa (1981) and Kessler & Gollop (1988), and are interpreted to have formed by storm or rip current scour accompanied (during their waning stage) by supension fall out.

Analogous strutures are also produced on the foreshore of Sapelo Island by the draping of scours by sand settling out of wave wash (Wunderlich 1972).

Summarising this coarsening upward sequence is interpreted to record the progradational passage from offshore to a low to moderate wave energy, storm influenced, shoreline.

4.6.2.7 **Offshore, Shoreface to Inshore Tidal Channel Sequence.**

In this variant to the Offshore-Foreshore sequence, the shoreface, or foreshore, deposits of the **planar sandstone facies** are erosively overlain by cross-stratified siliciclastic sandstones of the **channelised sandstone facies** (type 1 as described in section 4.6.2.2) (see Fig 4.62).

The base of the channelised facies is marked by a reliefed erosion surface lagged with bioclasts, and pebbles. Channel sequences show a fining upward fill dominated by, bi-directional trough cross-stratification. The top of these sequences is marked by the passage into ripple laminated deposits of the inshore shoal, **heterolithic facies.**

Interpretation.

The lower part of this sequence records the progradational passage from an Offshore to an upper Shoreface-Foreshore setting (see section 4.6.2.6), whereas the channelised facies are interpreted as inshore tidal channel fills (see section 4.6.2.2).

Coarsening-upward, offshore to shoreface sequences, truncated by inshore tidal channels are characteristic, progradational sequences of barrier and tidal inlet systems, of mixed, wave-tide shorelines (Elliott 1986). The lack of planar laminated deposits, as detailed from the top of 'classic' tidal-inlet, channel sequences (Kumar & Sanders 1974) may be interpreted as the abandonment of the outer part of the channel, seaward of the inlet mouth, or barrier shoreface where such facies are typically generated.

4.6.2.8 **Foreshore-Shoreface and Distributary Mouth Bar Facies Association**

This association comprises well sorted, fine grained siliciclastic sandstones of the **planar sandstone facies** within which, m - tens of metre's thick (2-13m) units of coarse sandstone and conglomerate of the **pebbly sandstone facies** and **channelised conglomerate facies** are developed (Fig. 4.63).

The **planar sandstone facies** has been described in detail in section 4.6.2.6.

The **pebbly sandstone facies** comprises non-channelised, coarse sandstone dominated units, with graded conglomerates, and cross-stratified and ripple laminated, coarse-medium sandstones (Fig. 4.63 - 4.64).

The conglomerate beds are 10 - 40cm thick with planar erosive bases, and are laterally traceable for at least 30 m (exposure widths). They are normally graded, with beds having a clast supported, cobble-pebble conglomerate base fining up into a coarse to medium grained sandstone (Fig. 4.65). The clasts are well rounded, poorly sorted, and typically of 3-6cm diameter, but range up to 18cm. The sandstone in the upper part of beds may be structureless, planar laminated, or trough cross - stratified (Fig. 4.66). The tops of beds may be current ripple laminated.

These conglomerate beds are typically bioclast free, but may contain fragmented shells of the bivalves *Ostrea, Pecten* and *Dosinia*.

Interstratified within these conglomerate beds are current ripple laminated sandstones, graded, granule layers, and sets (5-25cm) of trough cross-stratified bioclastic sandstone in co-sets up to 2m thick. These cross-stratified sandstones contain the abraded shells of marine bivalves *Pecten, Dosinia, Cardium, Pholadomya* and *Ostrea*. Shell beds in which the shells are bioclast suported with a tightly packed, imbricate fabric may also be developed within the facies.

Isolated beds of planar, and swaley stratified sandstone of the planar sandstone facies overlying a cobble or shell lagged erosion surface are also randomly interbedded within the pebbly sandstone facies (Fig. 4.64).

Bioturbation in this facies is low with *Skolithos, Planolites* and to a lesser degree *Diplocraterion* present.

Palaeocurrents measured from cross-stratification and imbrication from the conglomerate and coarse sandstone beds are essentially unidirectional in an inferred offshore direction (Fig. 4.63 - 4.64). The associated bioclastic sets are also predominantly offshore directed, but with subordinate opposed sets, particularly of ripple lamination, present.

The **channelised conglomerate facies** comprises erosive-based, fining-upward channel fills developed within the pebbly sandstone facies. The fills 2-3m thick and are dominated by tabular, and shallow trough-shaped beds of graded, imbricate conglomerate (Fig. 4.63 and 4.67). Conglomerate beds are of up to 30cm thickness, with clasts being up to 15cm diameter (see section 4.6.2.9. for further details).

Interpretation.

The **planar sandstone facies** which forms the 'background' facies of this association are interpreted as the deposits of an upper shoreface-foreshore environment (see section 4.6.2.6).

The **pebbly sandstone facies** units are therefore also interpreted to have been deposited in such an environment with support coming from the marine trace fossil assemblage and marine bioclasts within the facies.

The graded conglomerate beds represent the deposits of waning turbulent, flows in which the conglomerate was transported as bedload traction carpets. The structureless sandstone of the upper part of many of these beds may have been deposited as graded suspension deposits, whereas the stratified sandstone of other graded beds were deposited by a unidirectional traction current, waning from an upper (planar lamination) to lower (ripple field) flow regime.

In shallow marine settings, conglomerate may be transported offshore by storm generated currents, (Wright & Walker 1981, Leckie & Walker 1983, Bourgeouis & Leithold 1984), as well as fluvial effluent (Wright 1977). However, the low to moderate wave energy envisaged for this shoreline is unlikely to have generated currents capable of transporting the conglomerate, and consequently the preferred interpretation of the beds is as the deposits of fluvial flows issuing onto the shoreface from a distributary channel.

The interbedded, planar and swaly stratified sandstones with lagged bases are interpreted to have formed under wave action. Storm waves are envisaged to have scoured and winnowed the substrate, concentrating the conglomerates on scour surfaces, with the fine sandstone deposited from suspension under wave and combined storm-wave currents as the storm waned. A similar interpretion has been given to conglomerate - sandstone beds, of the shoreface of Fire Island (Kumar & Sanders 1976), and, ancient barrier foreshore and channel mouth bar facies (Kleinsphen *et al.*, 1984).

The thick, imbricate shell layers of the facies are attributed to the concentration of shell material under prolonged wave action. Shell banks with the same imbricate fabric, are developed under wave action at mean low water level on the coast of Essex (Greensmith & Tucker 1969).

The **channelised conglomerate facies** is interpreted as the fill of a fluvio-distributary channel, with the imbricate conglomerates being the deposits of longitudinal bars (see section 4.6.2.9 for further discussion).

Anadon *et al.* (1985) described a similar facies association from proximal, fan - delta mouth bar sequences, with conglomerate deposition attributed to flashy stream floods issued into a marine setting.

An association of fluvial mouth bar conglomerates, and well sorted barrier shoreface-foreshore sandstones has also been detailed by Kleinspehn *et al.*, (1984). As discussed by these authors (*op. cit.*), the alternation of shoreface and fluvio-distributary units may be attributed to avulsion of fluvial channels, progradation of mouth bars, or lateral migration of spits and barriers.

4.6.2.9 Fluvio-distributary Facies Association.

This facies association is unique to the north-east of the Digne-Valensole basin where it forms a large scale coarsening upward, (mega-) sequence of 100m thickness (Fig. 4.71). At the base of the sequence, tabular units of ripple laminated sandstone are developed within the **muddy sandstone facies**, with the upper part of the mega-sequence comprising repeated fining-upward units of conglomerates and coarse sandstone of the **channelised conglomerate facies** and bioturbated **brackish muddy sandstone facies** (type 2) (Fig. 4.68-4.69)

At the base of the mega-sequence the sandstone units within the **muddy sandstone facies** comprise m scale thick (1-3m), tabular units of fine-medium grained sandstone, and subordinate conglomerate, which may be traced laterally for up to 400m. The units have a gradational base and top and typically show a coarsening upward base overlain by a fininig-upward top. The units are heavily bioturbated, notably by *Diplocraterion* and *Skolithos* and with large vertical burrows some of which contain *Pholadomya* bivalves in burrowing position.

Current ripple lamination is the most abundant physical sedimentary structure of the facies , with subordinate wave ripple lamination present. Decimetre scale (10 -40cm) sets of trough cross-stratification may be present as isolated sets, rarely forming co-sets. The cross-stratification and lamination is predominantly unidirectional, but with the occassional set of opposed ripple lamination. Thin (<3 cm) graded, pebble-granule to coarse sandstone beds with current ripple laminated tops are commonly present within these sandstone units (<10%).

The interbedded muddy sandstone facies is as described in section 4.6.2.3.

The mega-sequence gradationally coarsens upward, with thin graded gravel beds progressively increasing in size and proportion, and pasing into an association of metre scale, fining-upward conglomerate units of the **channelised conglomerate facies** interbedded within **brackish muddy sandstone facies.**

The conglomerate units are traceable across outcrop for some 300 - 700m, having a tabular sheet like appearance, and with some sheets having a shallow erosional relief of the order of 1 - 1.5m (Fig. 4.71 - 4.72). Laterally units pass out into the muddy marine sands within which they are interstratified. The conglomerate units form fining-upward sequences with a basal section dominated by tabular beds of massive, or imbricate, graded and horizontally stratified conglomerate (Gm lithofacies). Beds may take the form of extremely shallow and wide troughs, 10-40 cm deep and 3 - 15m wide.

The conglomerates are cobble to pebble grade, having a coarse sandstone matrix. The conglomerate is polymodal, and clast supported, with clasts being well rounded, and commonly imbricate with an a(t) b(i) fabric (Harms *et al.*, 1982) (Fig. 4.74).
Sets of planar cross-stratified conglomerate, (Gp lithofacies) of 0.8-1.4m set size are also present. Thin lenses of of well sorted pebble conglomerate and sandstone are interbedded within the coarse conglomerate.

Conglomerate units fine upward into trough cross-stratified and Current ripple laminated sandstone.

Bioturbation in this facies is generally restricted to the sandstone at the top of the conglomerate units where *Ophiomorpha* and, *Skolithos* burrows are present. The conglomerates, and sandstones comonly contain fragmented bioclasts of *Ostrea*, often as whole single valves (Fig. 4.75), and less frequently, *Cardium, Dosinia* and echinoid fragments.

The conglomerate units are sepearated by dm-metre scale intervals of **muddy sandstone facies** which comprises heavily bioturbated, muddy sandstones with in-situ *Dosinia* and *Pholadomya* bivlaves, and thin, diffuse shell layers of the restricted marine *Turretella* gastropods and *Ostrea* bivalves. *Ophiomorpha* burrows are recognised within this facies. Thin graded pebble beds within this facies may be traced back into laterally adjacent conglomerate channels (Fig. 4.71 - .72).

At the very top of the mega-sequence, the conglomerate units become interbedded within grey mudstones containing a monospecific *Turretelid* fauna, and showing the developement of gley mottled horizons, and calcrete profiles.

Palaeocurrent trends from the conglomerate association show a mean WSW direction (inferred as being offshore directed) (fig 4.71). Data from the associated coarse sands is generally concurrent with this but with a subordinate NNW - ENE set derived from ripple stratified sands in the upper section of units.

Interpretation

The deposition of this coarsening-upward mega-sequence in an open to marginal marine environment is clearly indicated by (1) the developement of offshore facies at the base of the sequence (2) the vertical and lateral passage of the conglomerate facies into mudstones with a brackish marine fauna (3) brackish marine trace fossils, and macro-fauna within the sandstone and conglomerate units.

At the base the current ripple laminated and trough cross-stratified sandstone units record the periodic develop ment of lower flow regime conditions in the offshore environment. The develop ment of graded conglomerate beds records the deposition from waning turbulent flows, whose competence and frequency apparently increased as the the sequence coasened upwards. The restricted developement of bidirectional cross-lamination within these units suggets that weak tidal currents were operative.

The coarse grained, conglomeratic nature of the upper part of the megasequence requires flows of sufficient strength to transport conglomerate bedload, and generate shallow channel relief suggesting proximity to a fluvial channel system. The erosional base, and stratified nature of the conglomerates, are therefore interpreted to be the product of fluvial currents issuing into a shallow marine environment.

The massive, sheet conglomerates (Gm) which dominate the conglomerate facies are characteristic of the bedload deposits of longitudinal bars, or, diffuse sheets (Smith 1974), with the planar cross-stratified conglomerates generated as large, flow transverse, bedforms or bars (Collinson 1986), or on the downstream margin of longitudinal bars (Rust 1978). Fluctuations in discharge are recorded by the preservation of pebble conglomerate, and cross-stratifed sandstone lenses within the facies, which are typical of low flow stage deposits (Steel & Thompson 1983).

In conlusion, the sandstone unit facies at the base of the mega-sequence are interpreted as distal distributary mouth bar deposits with the granular and sand material being primarily deposited from fluvial flood currents. Anadon *et al.*,(1984) detailed similar sand and granule units from the Fluvio - Distributary system of the Eocene, Monserrat fan - delta.

The channelised gravel facies is interpreted as having been generated in a proximal, distributary mouth bar setting, at the transition from channelised to unconfined sheet flows. The sharp base to units, and their subsequent fining-upward profile are taken to represent the lateral migration and abandonment of fluvial-distributary channels.

Palaeocurrent data suggests that these strongly fluvial influenced sequences, were subjected to a low degree of tidal reworking during their upper 'abandonment' phase.
The low degree of tidal reworking, and the absence of wave, or storm deposits, suggests that this facies association was deposited in a low tide, and wave energy, (partially enclosed ?) shallow sea. The grey mudstones with a restricted marine fauna, and pedogenesis at the top of the sequence are similar to inter- to supra-tidal, estuarine, mudflat deposits (Tucker 1973).

This conglomerate and muddy sandstone facies association bears close similarity to the 'proximal marine association' of Miocene coastal alluvial fans in S.W Turkey (Hayward 1983). An analogous facies, interpreted as a 'Proximal mouth bar conglomerate association' is described by Anadon *et al.*, (1985) from the Eocene Monserrat Fan - Delta System.

4.6.2.10 **Alluvial Plain Facies Association**

This association comprises 2-4m thick channelised siliclastic sandstones, interstratified within ochrous mudstones and sheet sandstones. These form repeated, erosive-based, fining-upward sequences, passing up from a coarse grained **channelised sandstone facies** to a mudstone dominated **tabular sandstone and mudstone facies** (Fig. 4.76).

The basal section of the channel fills is of decimetre scale (0.20 - 0.50m), trough cross-stratified, coarse to medium grained sandstones with thin beds of rippled sandstone, siltstone and mudstone. Vertically, set size and grain size decreases with heterolithic ripple laminated sandstone, siltstone, and mudstone dominant in the upper section of the channel. At their margins channels may show a progressive infilling by fine grained mudstones and siltstones, giving a channel 'plug' (Fig. 4.77).

Perpendicular to palaeoflow, the channel bodies may be traced for 20 - 50m, having a single storey, tabular, or simple ribbon form, or more commonly multistorey ribbon bodies being developed

(Fig. 4.77). Lateral accretion surfaces 10 - 25m wide, dipping at 12 - 20 ° are developed within both the ribbon and tabular channels. The accretion bedding extends to within a few decimetres of the channel base and may show the developement of dessication cracks in its upper part.

The channel sequence is transitionally overlain by ochrous mottled mudstones and siltstones interbedded with sheet sandstones which form tabular bedded units of dm-m scale, 2m maximum thickness (Fig. 4.78). The sandstone beds are graded with and current riple laminated. Small ribbon sandstone channels 0.30-0.80m thick and 1-3.5 m wide are interbedded within these sheet sandstone units (Fig. 4.78). These channels exhibit lateral accretion bedding and fine mudstone plugs (Fig. 4.79). These sandstone units can be traced back to the channels from which they extend for several tens of metres as channel 'wings'.

The mudstone dominated intervals comprise ochrous (10YR6/6) mudstones, with a grey (N6) mottle (gley) and contain thin sandstone and siltstone streaks. At the top of complete fining upward sequences, the mudstone becomes a homogeneous or weakly mottled, reddish brown (10R5/4) with calcrete glaebules, and forms distinctive horizons traceable for at least 250m. Dm-m scale beds organic rich, mudstone of the black mudstone facies (see section 4.2.6.4) may also be developed at the top of the sequence.

Bioturbation in the fine member, mudstones and sheet sandstones, is of a low to moderate level, with sub-vertical cylindrical burrows, of 0.2 - 0.8cm diameter, having a simple 'y' branching and meniscate structure. Structureless horizontal burrows of similar dimension are also present.

In a further variant to the sequence it may fine up into a heterolithic, tidal flat facies. At the transition from the tidal facies, to this facies, dm scale, sheet sandstone units are interbedded within grey, pedogenically mottled mudstones of the supratidal flat mudstone facies.

Cross-stratification in the channels of this facies association is unidirectional, with channel fill orientations showing a variance of some 80°. Within the sheet sand facies directional data has a higher degree of variance and is directed at an oblique to perpendicular direction to that of the channel.

Interpretation

Fining upward, channelised sandstones interstratified within pedogenically mottled fine grained mudstone and siltstone facies are characteristic of most sandy fluvial sequences (Collinson 1986). The fining upward, channelised facies are characteristic of the vertical sequence of a laterally migrating, mixed bed-load channel (Allen 1974, Schumm, Galloway 1982). Repeated fining-upward sequences record repetitive channel establishment.

The accretion bedding is the product of lateral accretion on the point bar of a meandering channel (Allen 1965). The well developed accretion bedding, with its thin siltstone, mudstone intervals and dessication cracks, indicate that point bar accretion was discontinous and suggests that the channels alternated wet and dry (Puigdefabregas & Van Vliet 1978). The mud filled channels are interpreted as abandoned channel fills.

The sheet sandstones with their erosive bases, and ripple drift lamination are characteristic of the deposits of waning overbank flows, with channelisation producing the small, interbedded ribbon channels.

The mottling of the mudstones is characteristic of pseudogley soils which are restricted in their development to flat landscapes, or to depressions with impeded drainage (Buurmann 1980) (See also paleosol facies 1 interpretation. Chap4).

At the top of the sequence, the homogeneous, and gley mottled mudstones reflect oxidisation in a relatively well drained area. The upward trend from gley to pseudo-gley, or oxidised mudstones, together with the grain size decrease suggests passage from low lying, poorly drained, to well drained areas with distance from the channel. The developement of black mudstones at the top of some sequences is interpreted to record the establishment of a shallow, ephemeral lake or marsh in the interchannel area.

Where sheet sandstones of this facies association are associated with tidal flat facies they may be interpreted as crevasse splay deposits prograded into an interchannel bay area.

Analogous, marginal marine, facies associations, comprising mixed-load, distributary channel bodies, fine grained floodplain deposits and swamp facies are described from the coastal plain of the Niger delta (Weber 1971).

4.6.2.11 **Alluvial Fan Facies Association**

This association comprises interbedded mudstones and sandstones of the **sheet sandstone and mudstone facies** and conglomerates of the **ribbon channel conglomerate facies.**

The association forms large scale coarsening-upward sequences (mega-sequences) of ~ 100m thickness (Fig. 4.80) in which lagoonal bay mudstones (brackish mudstone facies; section 4.6.2.4) pass up into sheet sandstone and mudstone facies into a ribbon channel conglomerate facies..

At the base of the mega-sequence the sheet sandstone and mudstone facies comprises mudstone and sandstone forming repeated, m - tens of metre scale, coarsening-, and less commonly fining-upward sequences.

The sandstone dominated units, comprise 2-25cm thick, graded sandstone beds (Fig. 4.82). These are typically structureless, though commonly have a current rippled top, and very rarely, a wave rippled top. Discrete, or, multiple laterally connected, lensoid scours filled with graded sandstone, or pebble conglomerate, are developed along the length of sandstone beds (Fig. 4.81, and 4.82-4.83). These scours may be plugged by mudstone rather than sandstone (Fig.4.83). Syn-sedimentary soft-sediment deformation is ubiquitous within these units in the form of convolute lamination and ball-and- pillow structures (Fig. 4.85).

The sandstone units are seperated by mudstone dominated units. Thin beds of organic rich, black mudstone with *Planorbid* gastropods are occasionally developed in these units. At the base of the mega-sequences these black mudstone beds contain *Crassostrea* shells.

Bioturbation of the facies is of a moderate level, with simple vertical, and branching burrows preserved.

As the mega-sequence coarsens upwards ribbon conglomerate bodies of between 0.5 - 3.0m thickness and 5 - 15m width are developed (Fig. 4.80, and 4.86). This **ribbon conglomerate facies** is arranged as a series of fining upward sequences, which pass from channelised conglomerate bodies into sheet sandstones and mudstone facies containing glaebular calcrete profiles.

The channels have a single, and multistorey form and a fill of sub-rounded, boulder-pebble grade conglomerates. The fills show a simple fining-upward sequence dominated by dm thick, horizontally bedded imbricate conglomerate (Gm facies). Thin lenses of pebble grade conglomerate and coarse sandstone are interbedded within these coarse conglomerates. Current ripple laminated medium sandstone and siltstone forms the top of the channel fill. Tabular beds of graded conglomerate and sandstone extend from the ribbon bodies as channel wings.

Palaeocurrent data from the mega-sequence indicates that the graded sandstones of the lower sheet sandstone facies, and the channels of the upper ribbon conglomerate facies have a parallel, south-westward orientated, palaeoflow.

Interpretation

The mudstone and sheet sandstone facies of this asociation bear resemblances to the floodplain facies of the alluvial plain facies association (section 4.6.2.10), but, a number of differences make it worth considering in its own-right.

The graded sandstones are interpreted to be the deposits of waning turbulent flows, with localised scouring by the flow generating the small ribbon sandstone scours. The mud filled scours may have been generated by sandstone starved flows, or, by sediment by-passing during scour with mud deposition being from the dilute 'tail' of the scouring flow.

The ball-and-pillow deformation, which is ubiquitous in the facies, is considered by (Allen 1982) to be generated by Rayleigh-Taylor instability between rapidly deposited sandstone and mudstone layers. Such deformation structures are developed in a variety of environmental settings, but are probably commonest in deltaic deposits (Allen 1982).

The parallel palaeoflow orientation of the sheet sandstones of this facies and those of the overlying channelised conglomerates suggest that the former developed at the downstream termination of the channels. The lack of evidence of sub-aerial deposition *eg.* desiccation cracks, pedogenic reworking, suggests that they developed in a subaqueous setting, and this is tentatively supported by the interbedded black mudstones which may be interpreted as lacustrine (with Planorbid gastropods) and brackish bay/lagoon (with Crassostrea) deposits. However, the lack of wave ripple reworking, so common in lacustrine facies, appears to contradict this.

A possible anologue of this lower facies are the lacustrine, sandstone lobe deposits developed at the distal termination of alluvial distributary channels (Cabrera *et al.*, 1985).

The large scale conglomerate channels of the association are interpreted to be the fills of braided, distributary channels. The massive, and imbricated conglomerates of the fill are characteristic of the deposits of longitudinal, medial and diagonal bars of braided river channels (Hein & Walker 1977, Rust 1978). The developement of calcrete profiles in the interchannel deposits reflects deposition in a continental setting. Similar facies have been described from the mid-fan fringe of alluvial fans by Cabrera *et al.* (1985).

4.6.3 **Transgressive Facies and Facies Associations.**

4.6.3.1 **Shoreface conglomerate facies**

This facies consists of thin units of pebble - boulder grade conglomerate overlying a planar erosion surface (Fig 4.87). The conglomerate is commonly only a single, or double clast thick, but may be up to 70 cms thick where it infills localised shallow scours of metre width.

Clasts are well rounded, and range in size from a few cm's up to 80 cm, with cobble size clasts being typical. They are extraformational, being of Jurassic - Cretaceous carbonates and cherts, vein quartz and intra-alpine granites. The carbonate clasts are often heavily bored, with borings penetrating from all sides of the clasts (Fig 4.88 - .89). Marine bioclasts (bivalves, corals and sharks teeth) may be dispersed through the lag. This lag facies separates underlying, shallower or marginal marine facies associations from overlying muddy sandstone offshore facies (see section 4.6.2.3; see chapter 5 for further details of context).

The conglomerates pass abruptly or, show a gradational passage into muddy sandstones of the offshore facies. In the latter case thin trains of pebbles, or, isolated pebbles are developed within the muddy sandstone for up to 1.5m above the level of the erosion surface (see Fig 4.87).

Interpretation

This facies is interpreted to represent a conglomerate lag generated, by wave surf winnowing on the foreshore and upper shoreface, during erosional, shoreline retreat. Analogous, conglomerate horizons have been interpreted as transgressive lags by Clifton (1981). The thicker, gradational lags develop where sedimentation rates, are sufficiently high that fine grained sediments are able to accumulate with the conglomerate during the period in which the transgression passes through the area (Clifton 1981).

Shoreface lags in ancient sequences are considered characteristic of transgression under conditions of moderately rapid sea level rise, low subsidence rates, and low sediment supply (Elliott 1986).

4.6.3.2 **Condensed Limestone Facies.**

This facies is developed within the muddy sandstone offshore facies (section 4.6.2.3) and comprises a bioclastic limestone unit of some 4m thickness.

This limestone forms by the progressive decrease in the spacing of the swell and storm shell coquinas within the muddy sandstone facies which amalgamate to produce a heavily bioturbated carbonate unit. This condensed concentration of shelly material contains identifiable bioclasts of *Pecten, Cardium, Ostrea, Mya,* together with echinoid, bryozoan, corral and fish teeth and plates. The matrix to the bioclastic supported framework decreases from a fine sandstone at the base of the unit to a fine carbonate siltstone-mudstone at the top.

Interpretation

This carbonate unit is interpreted to have developed during a transgressive event, and to reflect a significant and prolonged period of clastic starvation of the offshore environment. During transgressive events siliciclastic sediment is commonly trapped in the nearshore and alluvial environments resulting in a starving of siliciclastic supply to the offshore (McRae 1972). During such a period bioclasic material material would form a high proportion of the material available, resulting in the accumulation of a condensed bioclastic level. Analogous transgressive carbonate sandstone levels have been described by Oomkens (1971) from the Niger delta.

4.6.3.3 Glauconitic Sandstone Facies

This facies comprises units of glauconitic sandstone within heavily bioturbated offshore muddy sandstones (see section 4.6.2.3). The facies is developed above a cobble horizon of the shoreface lag facies (section 4.6.3.3) which is overlain by heavily bioturbated muddy sandstones of the Offshore facies (section 4.6.2.3). Fine grained, sandstone units 1 - 2.4m thick occur within the muddy sandstone. These units are tabular, with a gradational base and are traceable as low ridges for up to 300m (Fig 4.90). The tops of the units are gradational or may be marked by a thin bioclast and pebble lag. The units may show a fining upward sequence or more symmetrical coarsening and then fining upward sequences. The sandstone units are intensively bioturbated and typically devoid of physical sedimentary structures (fig 4.90), but may contain remnant patches of wave and current ripple lamination. They have a high proportion of glauconite, occuring as micron-mm scale granules (up to 1mm), and larger, (1-3mm) ovate concentrations within burrows, giving the units a greenish hue in outcrop. The intensive bioturbation of the facies is by *Diplocraterion, Skolithos* and by large, vertical burrows which contain *Pholadomya* and *Dosinia* bivalves in burrowing position.

Interpretation

The conglomerate lag facies has already been interpreted as the product of shoreface erosional retreat, with the passage into the muddy sandstone facies recording the transgressive establishment of an offshore environment. The developement of sandstone units in the facies records periodic increases in energy conditions in the 'offshore' setting. The developement of bi-directional stratification within the sandstone units suggests that tidal currents were important in their genesis, with wave ripple lamination indicating that waves periodically reworked the substrate.

The abundant glauconite in the units is unique to these sequences and indicates that they formed under low rates of deposition in fully marine conditions (Pettijohn *et al.* 1972, Odin & Matter 1981). Analagous glauconitic sandstones of the Hampshire Basin developed under fully marine offshore conditions of low, water turbulence and oxygen levels, during phases of shoreline transgression (Plint 1984).

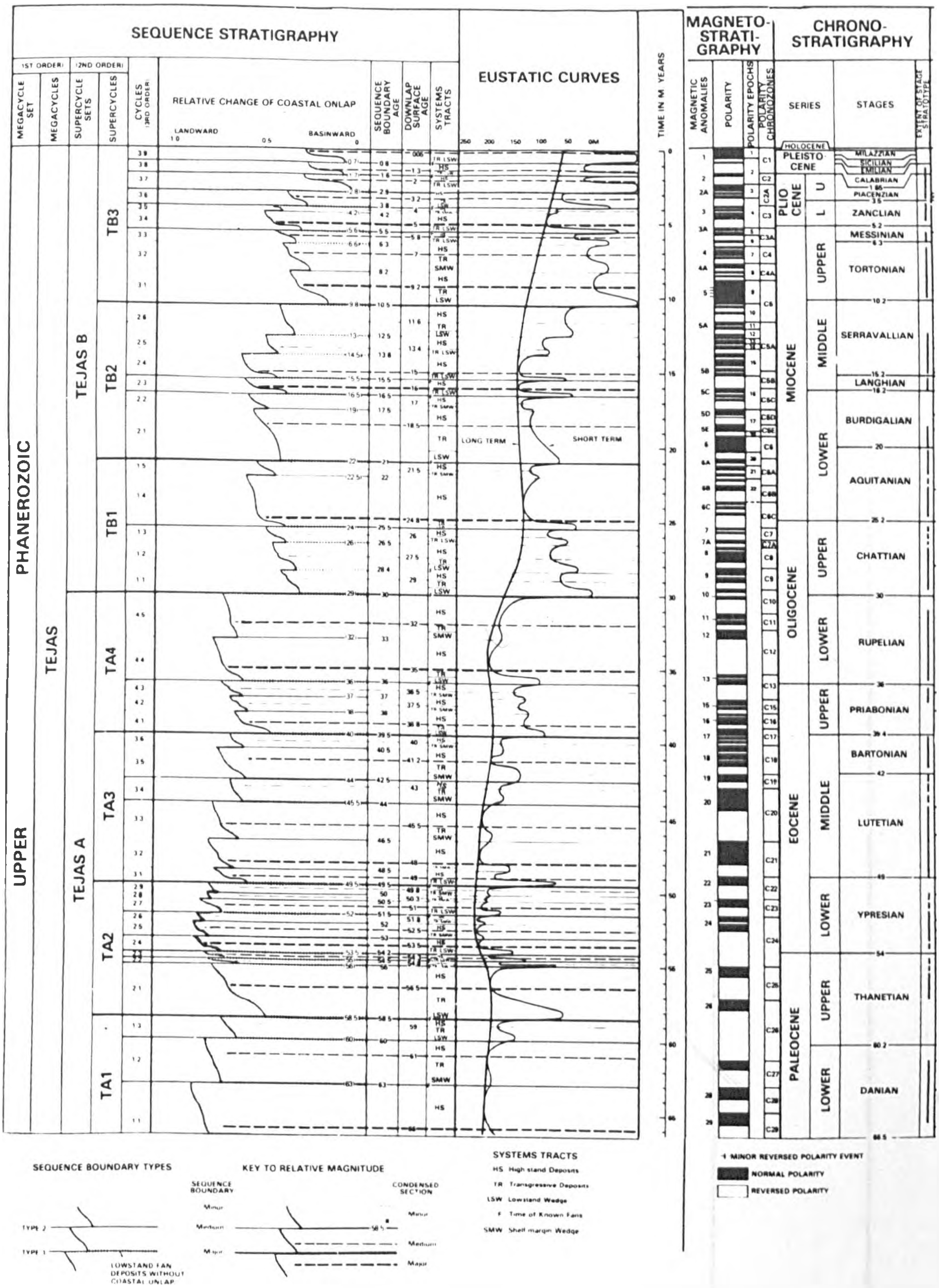


FIG. 4.1 Cenozoic eustatic sea levlel curve (from Haq *et al.* 1987).

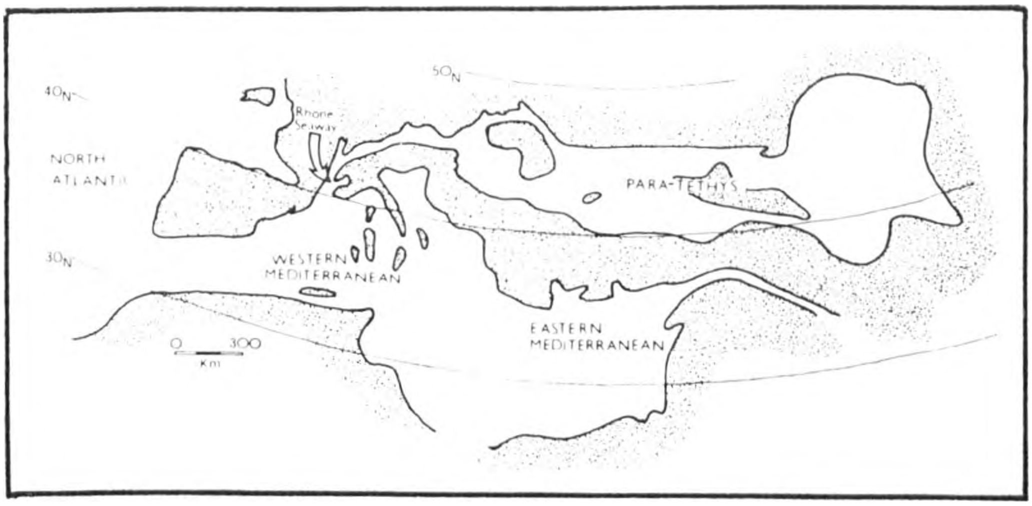


FIG. 4.2 Palaeogeographic map of the Lower Miocene (after Auboin *et al.* 1986)

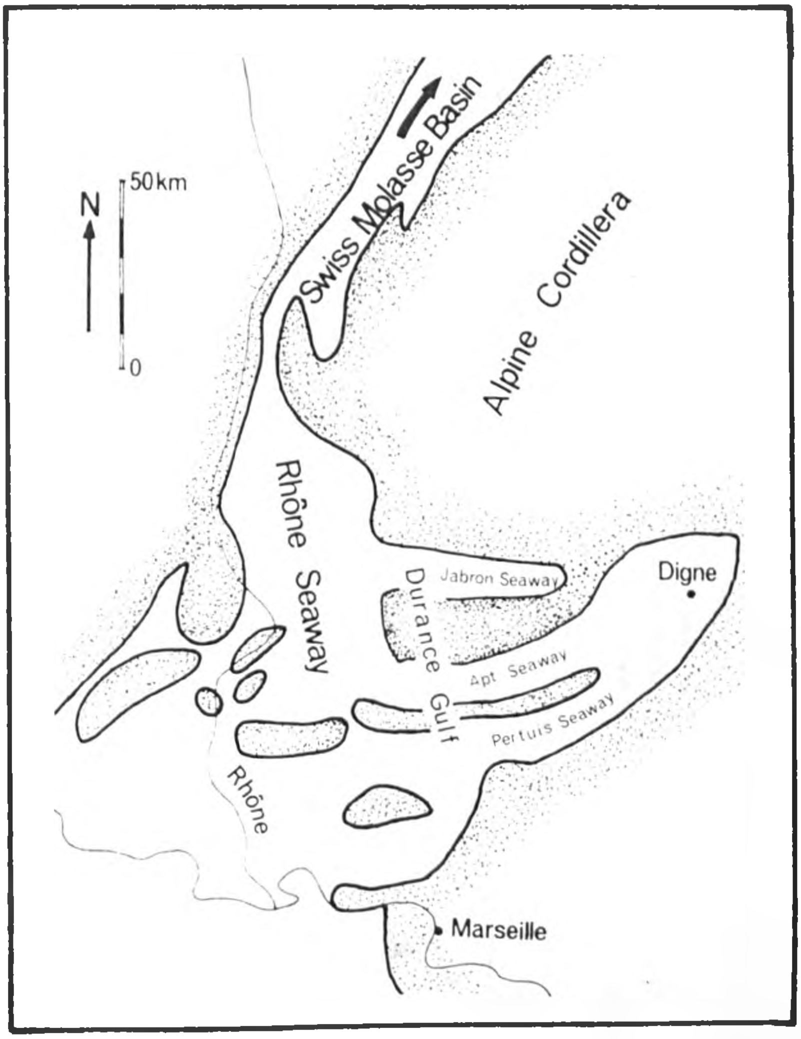


FIG. 4.3 Palaeogeographic map of the Burdigalian of S.E France (after Demarq 1984)

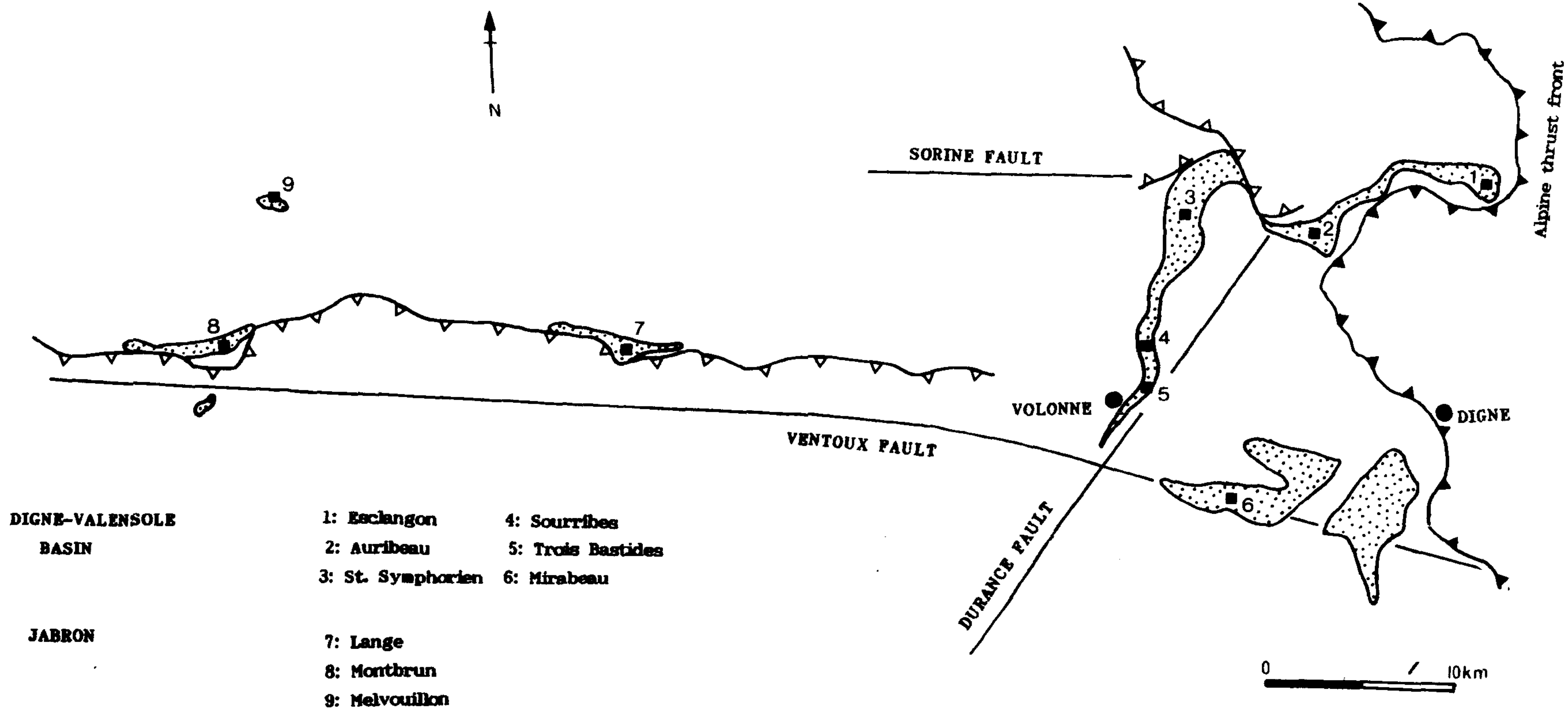


FIG 4.4 Outcrop map of the Marine Molasse formation detailing the studied sections.

PROGRADATIONAL FACIES ASSOCIATION	FACIES	INTERPRETATION
SHELF SANDWAVE	Trough cross-stratified (0.5 - 3m) bioclastic grainstone.	Open shelf tidal sandwaves.
INSHORE TIDAL CHANNEL AND SHOAL	Channelised sandstone (type 1) Fining-upward sequences (1-6m) dominated by bi-directional cross-stratification (0.3-3.5m scale). Flow reversal modification common. Lateral accretion bedding. Heterolithic (1) Heterolithic, bi-directional ripple lamination. Wave ripple sets, graded storm beds. Small scale sequences common.	Inshore tidal channel. Tide dominated inshore tidal shoal, wave and storm influenced.
OFFSHORE TO INSHORE TIDAL CHANNEL AND SHOAL	Muddy sandstone Intensively bioturbated muddy sandstone with relict patches of ripple lamination. Heterolithic (type 1) as detailed above Channelised sandstone as detailed above.	Offshore, above storm wave base. Inshore tidal shoal Inshore tidal channel
UPPER ESTUARINE	Channelised sandstone (type 2) Channelised fining upward sequences (1.5-5.5m), commonly display lateral accretion bedding, bi-directional trough cross- stratification (0.2-0.4m scale) and heterolithic ripple bedding. Heterolithic facies (type 2) Fining-upward from flaser-wavy to linsen bidirectional ripple bedding over ~4m. Plant debris, small, heterolithic filled channels (0.5-1.5). Mudstone facies Grey mudstone with yellow-brown mottling. Calcrete nodules. Black mudstone facies Organic rich black mudstone with fresh water gastropods oysters, thin lime- Brackish mudstone see below.	Estuarine tidal channel. Sub-, inter-tidal mixed flats Supratidal flats Lacustrine, palustrine, lagoonal swamp. Brackish bay, mudflats.

FIG 4.5 Facies and Facies Associations of the Marine Molasse formation.

PROGRADATIONAL FACIES ASSOCIATION	FACIES	INTERPRETATION
LAGOONAL BAY TO FLOOD TIDAL-DELTA	Brackish mudstone (1) Grey mudstone with relict patches of linsen bedding. Restricted marine fauna, plant debris. Thin graded sandstone with bioclasts	Brackish lagoonal bay.
	Heterolithic facies (3) Coarsening-upward sequence passing from linsen-wavy bedding to inter-bedded wavy-flaser and trough cross-stratification. Large scale inclined bedding dips parallel to palaeoflow.	Flood tidal-delta.
	Channelised sandstone facies As channelised facies (1), flood dominated, lateral accretion bedding common.	Tidal delta 'feeder' channel.
OFFSHORE TO FORESHORE	Muddy sandstone facies see above	Offshore
	Sandstone sheet Graded planar and hummocky cross-stratified sandstone interbedded with heterolithic ripple bedding, or bioturbated muddy sandstone.	Transition zone
	Planar sandstone Planar laminated sandstone, with wave ripple laminated tops, amalgamated hummocky cross-stratification, gutter cast scours.	Shoreface-foreshore.
OFFSHORE - SHOREFACE-INSHORE TIDAL CHANNEL	Muddy sandstone see above	Offshore
	Sandstone Sheet see above	Transition zone
	Channelised sandstone (1) see above	Inshore tidal channel.

PROGRADATIONAL FACIES ASSOCIATION	FACIES	INTERPRETATION
FORESHORE-SHOREFACE - DISTRIBUTARY MOUTH BAR.	Pebbly sandstone Graded, dm-scale conglomerate beds interbedded with trough cross-stratified coarse sandstone. Contains oyster shells. Subordinate fine grained planar laminated sandstone.	Distributary mouth bar
	Channelised gravel Fining-upward sequence of horizontally bedded, imbricate cobble conglomerate. Also planar stratified conglomerate. Contains oyster shells.	Fluvio-distributary channel - mouth bar.
	Planar sandstone see above	Shoreface-foreshore
FLUVIO-DISTRIBUTARY	Channelised conglomerate see above	Fluvio-distributary channel-mouth bar.
	Brackish mudstone (2) Intensively bioturbated muddy sandstones with restricted marine fauna. Thin beds of graded pebble conglomerate.	Brackish bay
ALLUVIAL PLAIN	Channelised sandstone facies (3) Channelised fining-upward sequences with uni-directional trough cross-stratification. Lateral accretion bedding, mud filled channels. Dessication cracks.	Coastal plain, fluvio-distributary channels,
	Tabular sandstone and mudstone Tabular units of thinly interbedded graded sandstone and pedogenically mottled mudstone. Calcrete profiles. Floodplain	Floodplain
ALLUVIAL FAN	Sheet sandstone and mudstone Units of graded sheet sandstone and mudstones Soft-sediment deformation common. Small (dm scale) gravel filled ribbon scours.	Terminal channel lobes, distal terminrl alluvial fan
	Ribbon channel gravel Fining-upward channelised sequences of cobble conglomerate, single, multi-storey channels.	Terminal fan braided distributary channels

TRANSGRESSIVE FACIES	DESCRIPTION	INTERPRETATION
Shoreface gravel	Bored boulder-pebble concentration above planar erosion surface.	Shoreface transgressive lag
Condensed limestone	Bioclastic limestone in offshore muddy sandstone facies	Transgression -clastic starved offshore/shoreface.
Glauconitic sandstone	Glauconitic sandstone horizons within offshore muddy sandstone facies	Transgression - sediment starved offshore.

SEDIMENTARY FEATURES	TIDAL PROCESS
Directional bi-modality in small-scale to large scale cross-laminated and stratified sets. One direction commonly dominant giving a "deficient bimodality"	Reversing tidal currents, deficient bimodality reflects segregation of tidal flow with areas of flood and ebb dominance.
Frequent occurence of discontinuity planes in cross-stratification namely (i) mud-drapes (ii) erosional pause-planes (iii) reactivation surfaces, which define (iv) tidal bundles. Tidal bundles may show lateral thickness sequences.	Reversing tidal currents result in repeatedly interupted progradation of mega-ripple, bedform modification by subordinate tidal current. Tidal bundle sequences record spring-neaptide cycle.
Widespread and quantitatively important developement of heterolithic ripple bedding.	Reversing tidal currents generate slack-water drapes.
Small-scale and large-scale stratification are intimately interstratified. Furthermore rapid vertical and lateral facies changes, with general lack of sequential organisation.	Variable strength of tidal currents due to differences in HW and LW levels by daily and spring/neaptide and/or wind influences on the tidal volumes and to shifts in tidal channel axis.

FIG 4.6 Diagnostic structural features of inshore tidal deposits (after de Raaf & Boersma 1971, Terwindt 1971 & 1981, and Visser 1980).

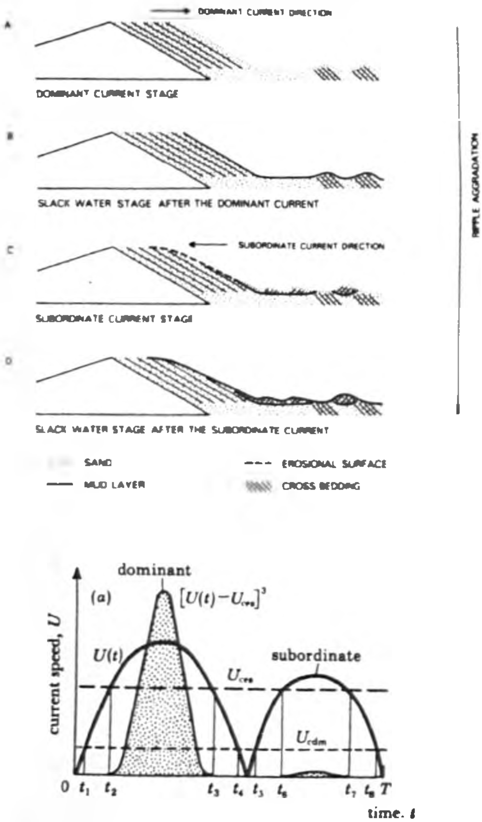


FIG 4.7 (i) Schematic interpretation of structures within cross-stratification generated during an ebb-flood tidal cycle having a pronounced velocity assymetry (from Visser 1980)
(ii) Sand transport and mud depositiopn during an assymetric tidal cycle as detailed in (i) (from Allen 1980).
U_{css} - Threshold of sand transport
U_{cdm} - Threshold of mud transport

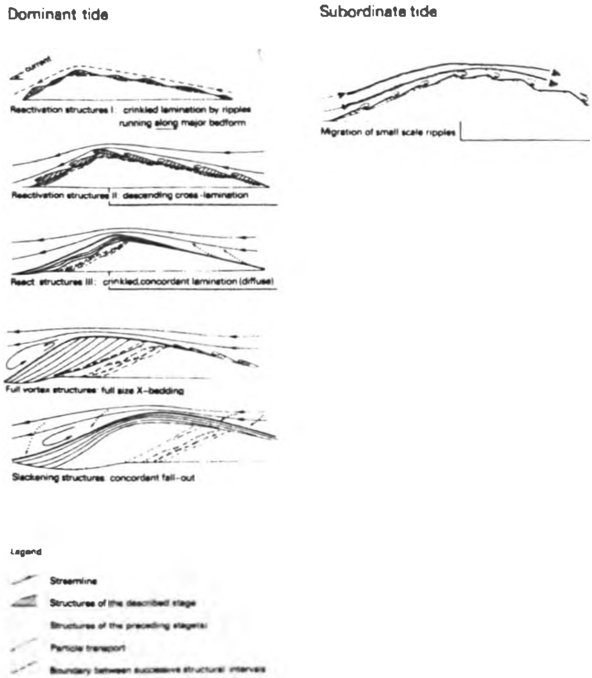


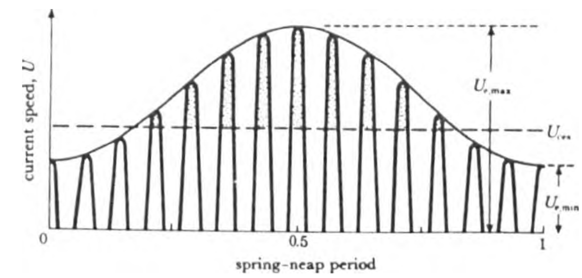
FIG 4.7 (ii) Formation of structural sub-intervals within a tidal bundle.

Dominant tide;- A: Weak current during acceleration phase of dominant tide after slack water phase.

B: Peak flow phase of dominant tide.

C: Deceleration phase of dominant tide

Subordinate tide;- A: Generation of upslope and obliquely climbing ripples on top of pause planes generates erosional pause planes. (from Terwindt 1981)

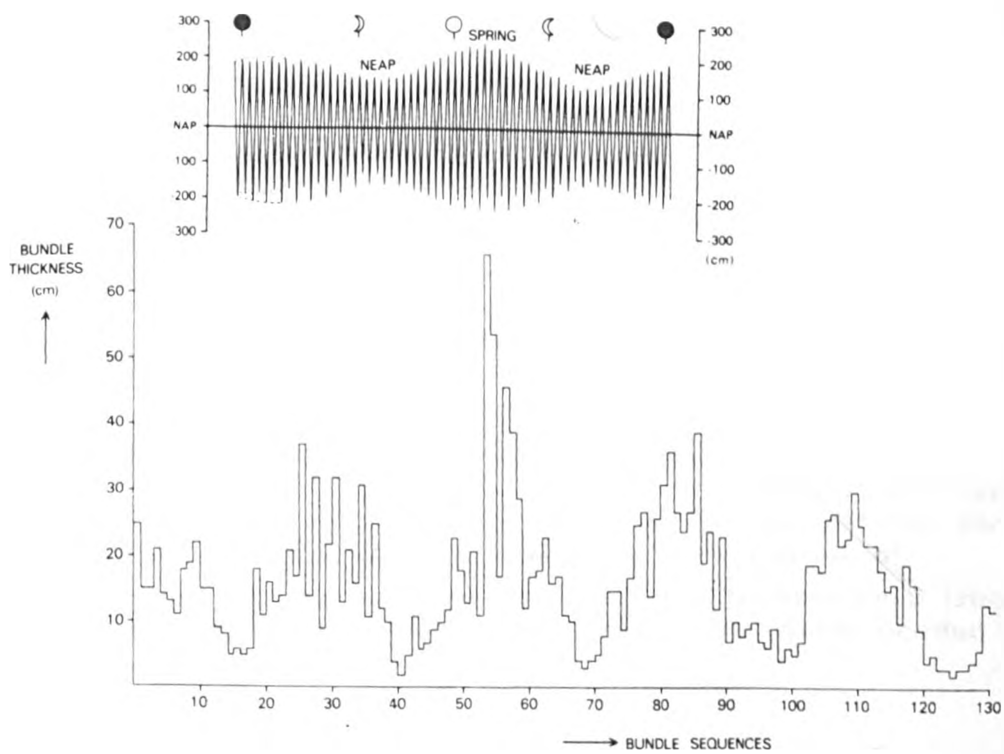


(i) Tidal velocity curve displays the idealised distribution of current speed over one full spring-neap cycle (diurnal). Sand transport is shown by the stippled portions of the curve. (from Allen 1980)

U_{ces} - sand transport threshold

U_{emax} - maximum velocity (spring)

U_{emin} - minimum velocity (neap



(ii) Neap - Spring tidal cycles within cross-stratification from a meso-tidal estuarine shoal, Dutch coast. A graph of the general tidal pattern of the estuary is included (from Visser 1980).

FIG 4.8 Tidal bundles within tidal cross-stratification.

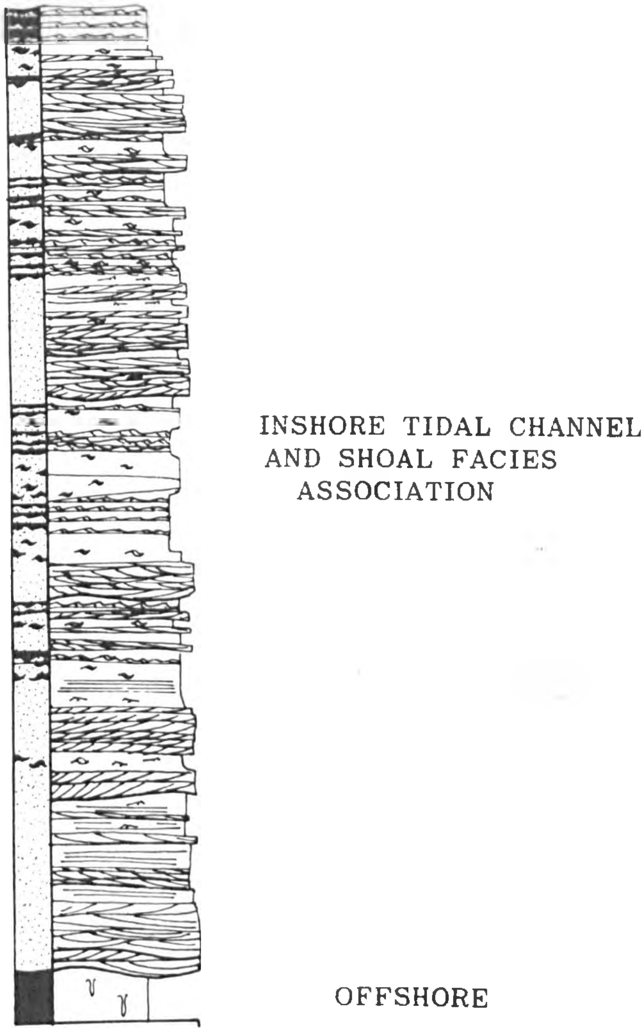


FIG 4.9 Inshore tidal channel and shoal association. Simplified graphic log of a section of inshore tidal channel and shoal facies associations as developed in the Vm1 member, St. Symphorien section. Note the metre scale alternation of channelised sandstone facies dominated by bi-directional trough cross-stratificatied sandstone and heterolithic ripple bedded facies.

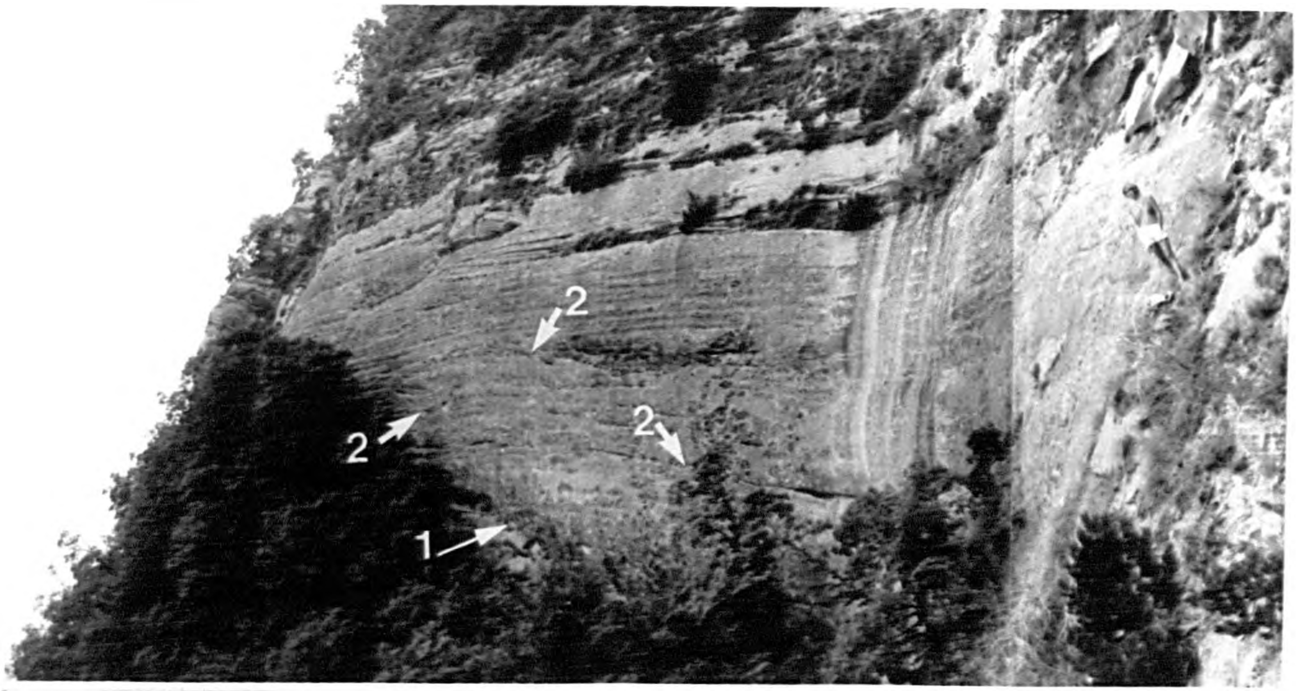


FIG 4.10 Channelised sandstone facies; large scale sets of trough cross-stratification overlying a channel erosion surface (1). Set boundaries (2) are at a low angle and erosional. Exposure is perpendicular to palaeo-flow. St.Symphorien section. Person 1.6m for scale.

FIG 4.11 (i) Channelised sandstone facies; large scale trough cross-stratification downlapping onto a channelised erosion surface at the base of a fining-upward sequence. The stratification passes out into well-developed toe-sets as detailed below.

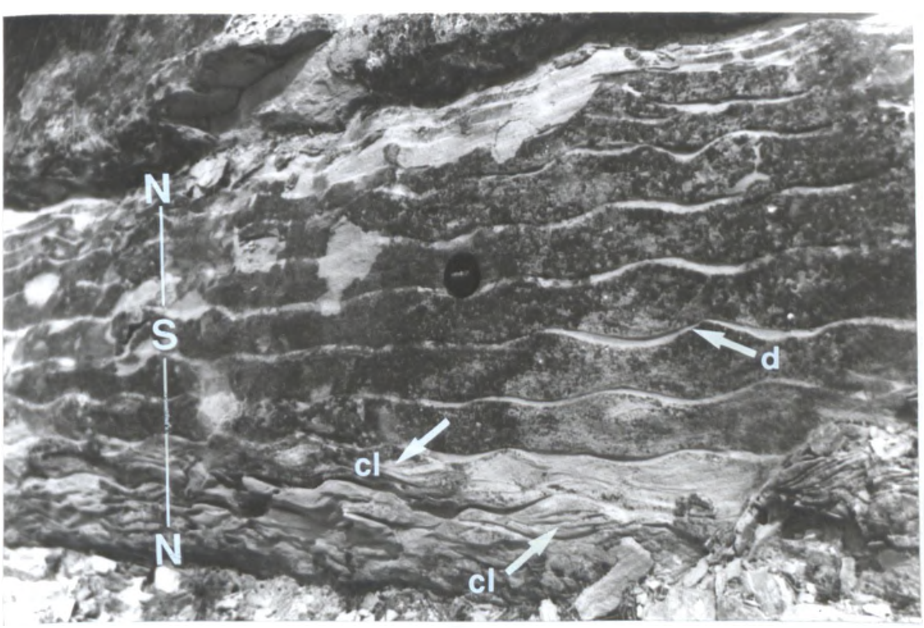


FIG 4.11 (ii) Detail of toesets (arrowed) of large scale trough cross-stratification of Fig 4.11 (i). View oblique to flow. Toesets form tidal bundles comprising ripple laminated form sets (dominant flow) with double siltsone drapes (d) (up to 1.5 cm thick). seperated by a thin (mm scale) layer of fine sandstone. The tidal bundles show a thickness periodicity of 26 (see Fig 4.14i) interpreted to define a semi-diurnal neap-spring-neap tidal cycle (N-S-N). Note the developement of climbing-ripple lamination indicating high rates of suspension fall-out in the lee of the bedform. Lens cap 5cm diameter for scale.

FIG 4.12 (i) Sigmoidal trough cross-stratification (i) showing the developement of swept-out toesets with thick double-siltstone drapes (d) (non-erosive pause plane). The lower siltstone drape (post-dominant flow slack water) is restricted to the toesets passing up dip into a ripple mantled erosive pause plane (p) - this reflects the erosional modification of the lee side of the bedform by the subordinate tidal current. The restriction of the upper siltstone drape (post-subordinate flow slack water phase) to the toesets and lower foreset suggests erosion of drape during the acceleration phase of dominant tide. Lens cap 5cm diameterr for scale. A tidal bundle sequence measured at this locality (see Fig 4.14ii had a periodicity of the order of 25 indicating the operation of a semi-diurnal tidal regime.

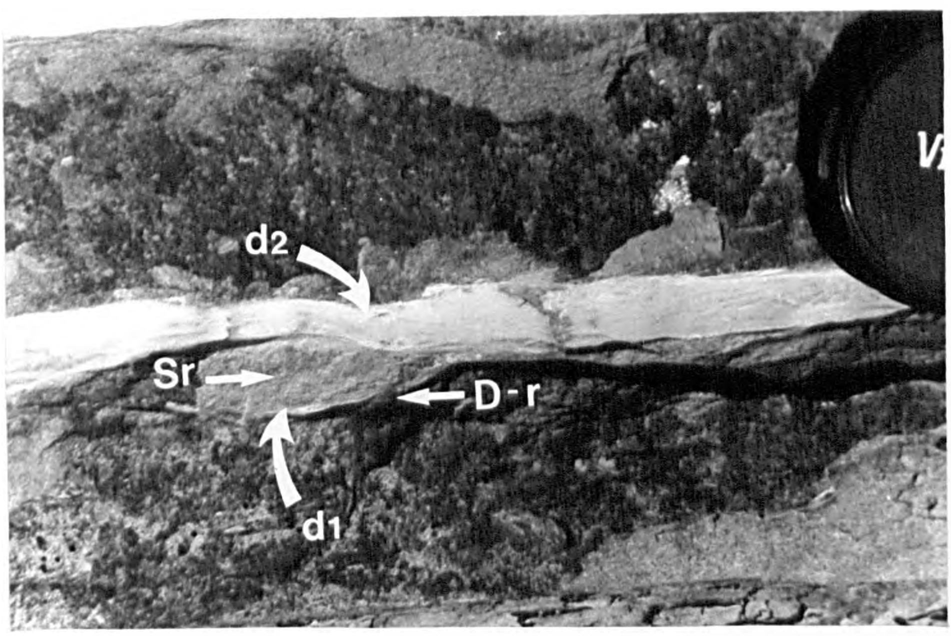
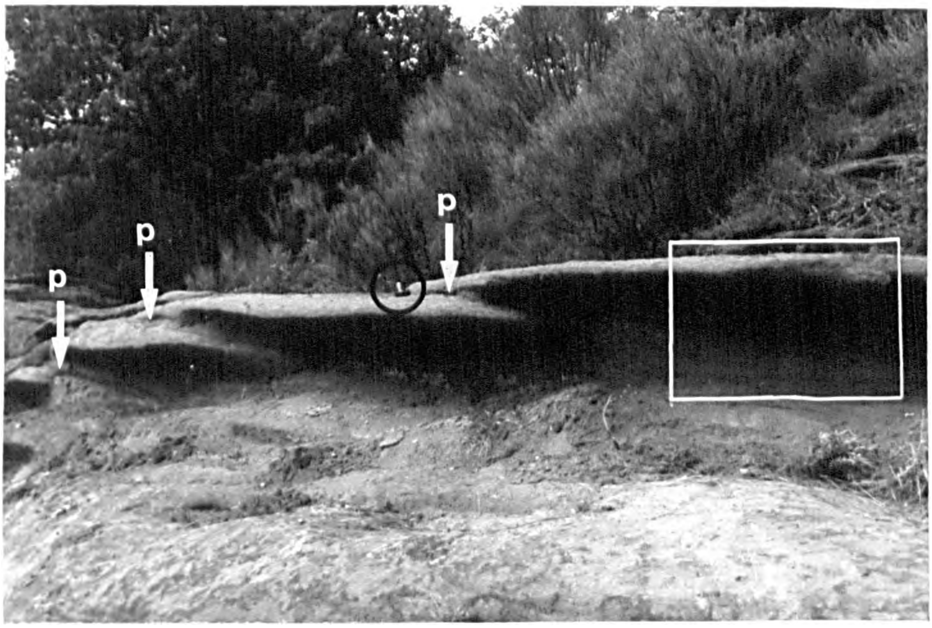


FIG 4.12 (ii) Close up view of double siltstone drape in toe-sets. Dominant flow toeset with ripple laminated top (D-r) and flaser siltstone drape (d1) overlain by opposed set of subordinate flow ripple lamination (S-r) and thick siltstone drape (d2). Lens cap 5cm diameter for scale.

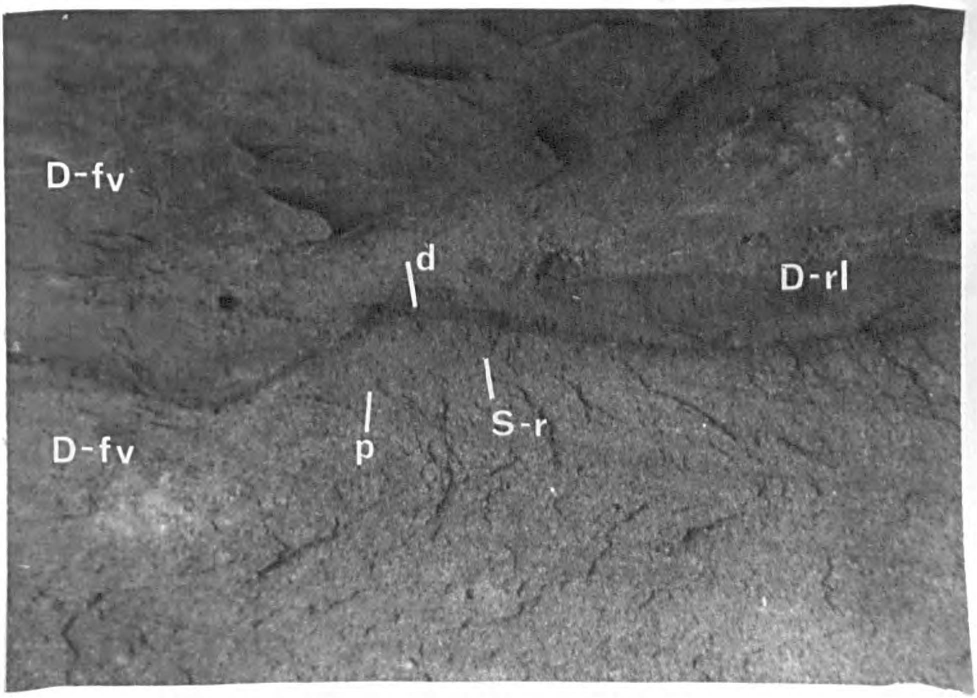
FIG 4.13 (i) Simoidal trough cross-stratification of the channelised sandstone facies. Tidal bundles clearly defined by prominent low-angle erosional pause planes (p). Trowel 30 cm for scale.

FIG 4.13 (ii) Detail of tidal bundle structures from withn area outlined in 4.13 (i). Dominant flow, full vortex stage sigmoidal stratification (D-fv) is interupted by subordinate flow erosional pause planes (p). Tidal bundles have widths of 1.2 - 2.4m.

FIG 4.13 (iii) Close up of tidal bundle outlined in Fig 4.13 (ii) Dominant flow full-vortex stage sigmoidal stratification (D-fv), truncated by erosional pause plae (p) mantled with subordinate flow ripple lamination (S-r) and slack-water siltstone drape (d). Overlying ripple lamination (dominant flow reactivation lamination D-rl) records acceleration phase of dominant flow prior to full vortex stage (D-fv).


p
p
p
p


D-fv
p
D-fv


D-fv
d
D-rl
S-r
p
D-fv

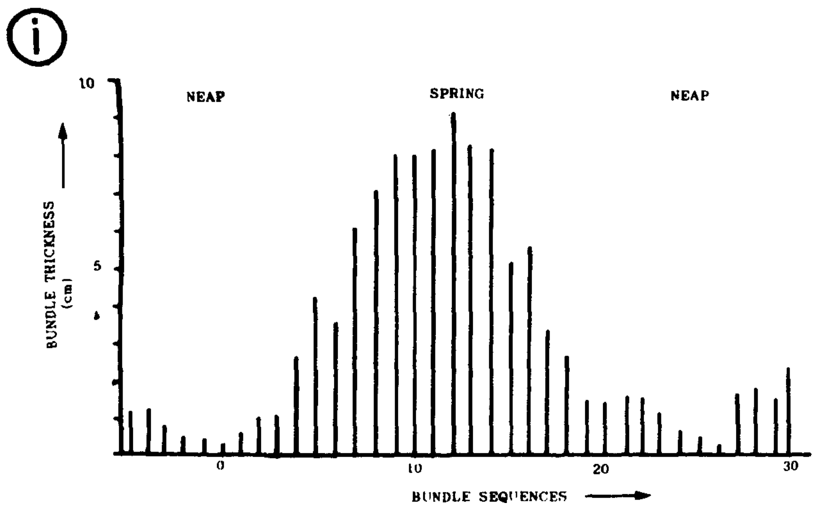
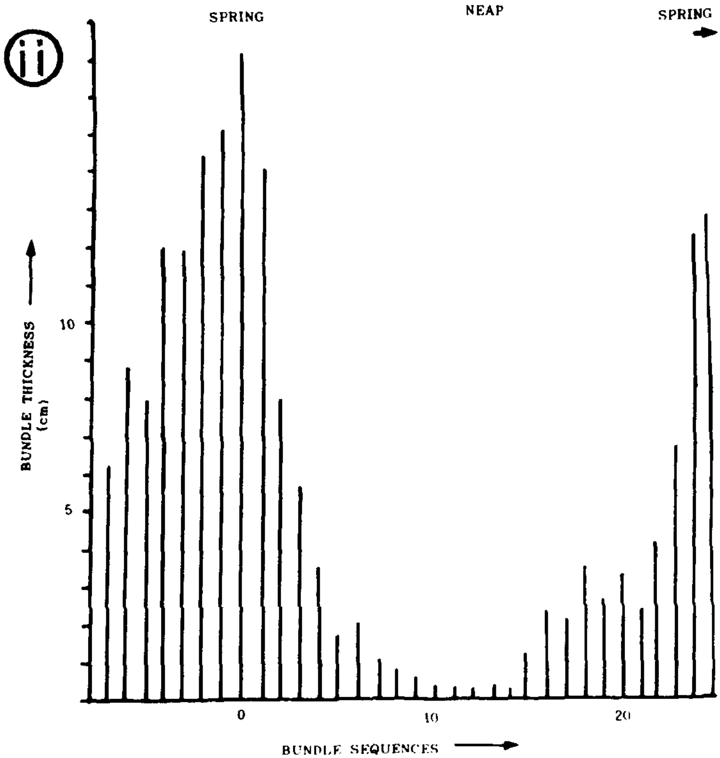
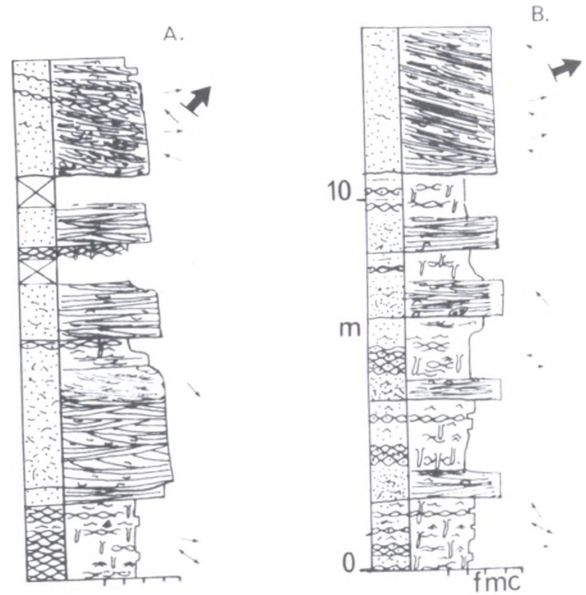


FIG 4.14 Tidal bundle sequences within strongly assymetric trough cross-stratified sets of the channelised sandstone facies reveals cyclic pattern to bundle thickness having periodicity of the order of 25-28, interpreted to record deposition under a semi-diurnal tidal regime. (i) St. Symphorien section, Bm2 member (see Fig 4.11) (ii) St. Symphorien section, Bm2 member, (see Fig 4.12).


ii
SPRING
NEAP
SPRING
BUNDLE THICKNESS (cm)
10
5
0
10
20
BUNDLE SEQUENCES



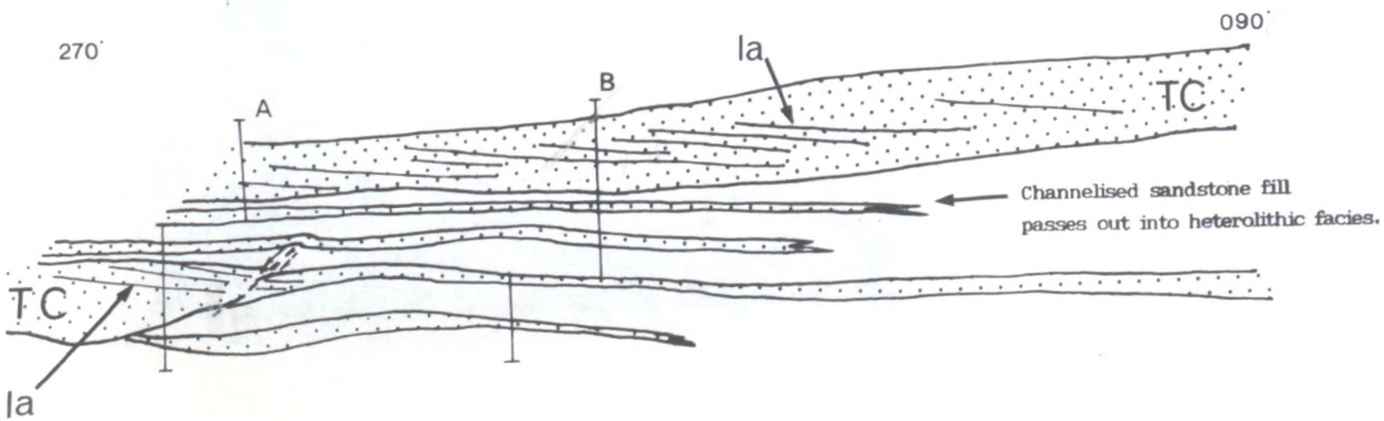


FIG 4.15 Photograph and interpretative sketch of tidal channel sandstone bodies (TC) enclosed within heterolithic facies of the inshore tidal facies association. The lower tidal channels show the developement of trough cross-stratified overbank 'wings' which pass out into the heterolithic facies. Note the well developed lateral accretion bedding (la) within the channels.

FIG 4.16 Linsen bedding within the heterolithic facies. Note the bi-directional lamination of these starved ripple sets. 1cm scale bar.

FIG 4.17 Wavy bedding within the heterolithic facies. Note the developement of bi-directional climbing ripple lamination (palaeoflow arrowed). 1cm scale bar.

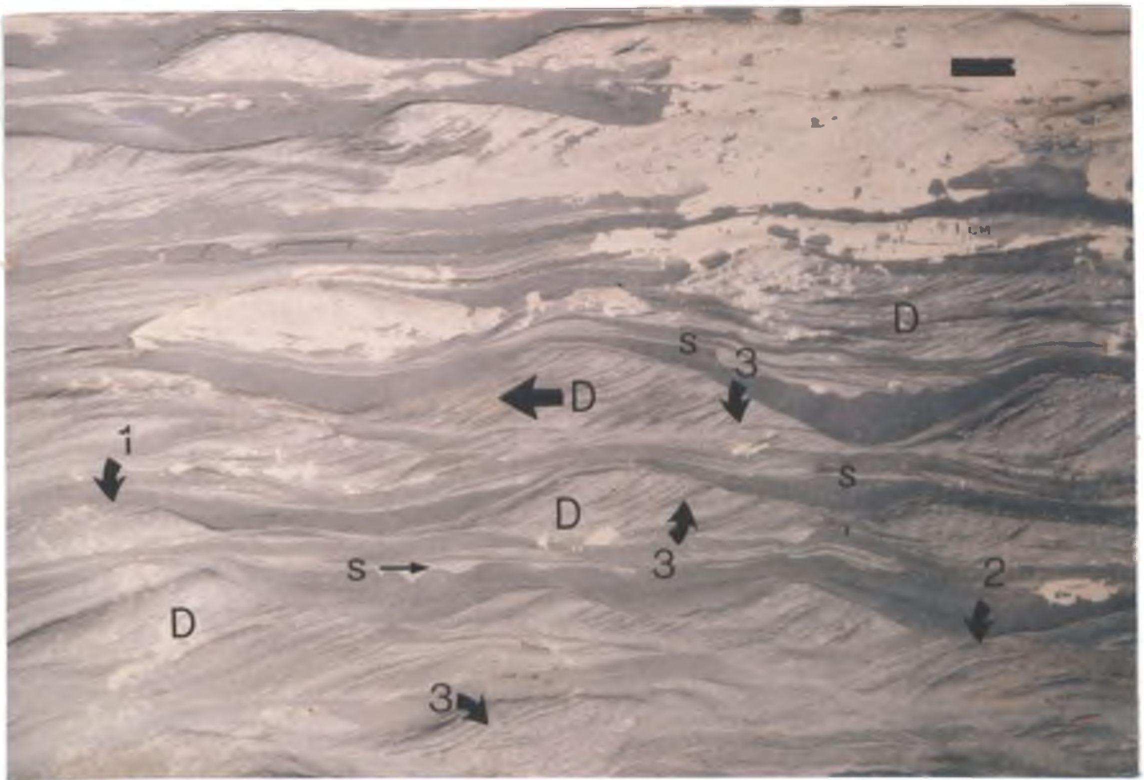


FIG 4.18 Bidirectional ripple laminated wavy-flaser bedding. Note the developement of alternating dominant (D) and subordinate (s) flow ripple laminated sets. Note also the common developement of symmetrical ripple forms (1), form discordant lamination (2), and internal erosion surfaces defining bundles of lamination (3), which are all characteristic of wave ripple lamination (de Raaf *et al.* 1977) and are interpreted to record wave modification of the tidal current ripple sets.

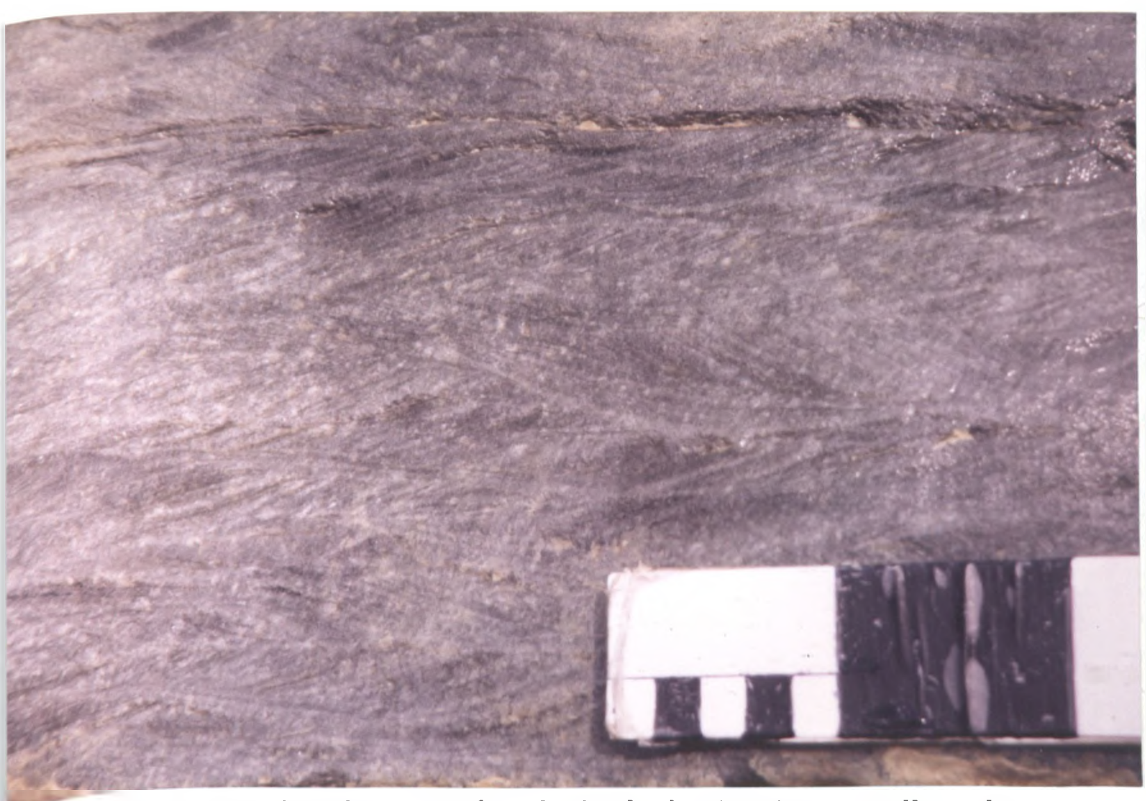


FIG 4.19 Herring-bone style of ripple lamination. Small scale bars 1cm.

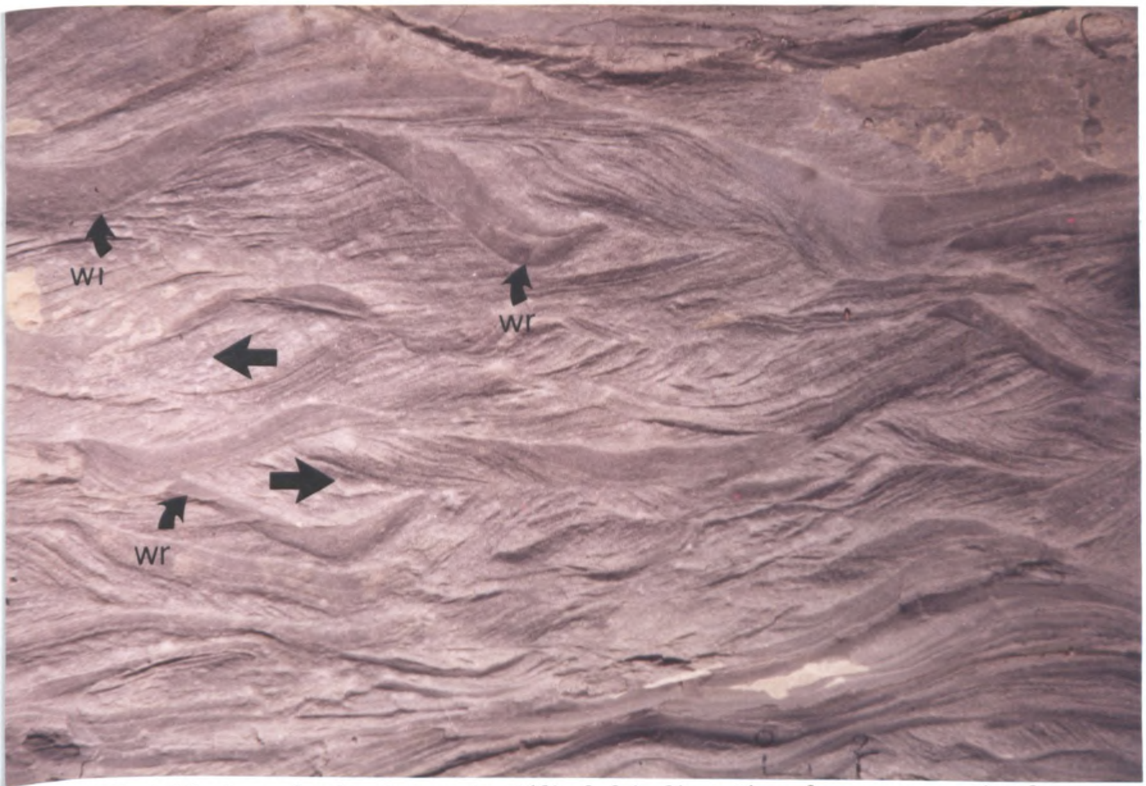


FIG 4.20 Complexly interstratified bi-directional current ripple lamination (arrowed) and wave ripple lamination (wr). Wave ripple sets have scoop shaped erosive bases draped by parallel undulating lamination.

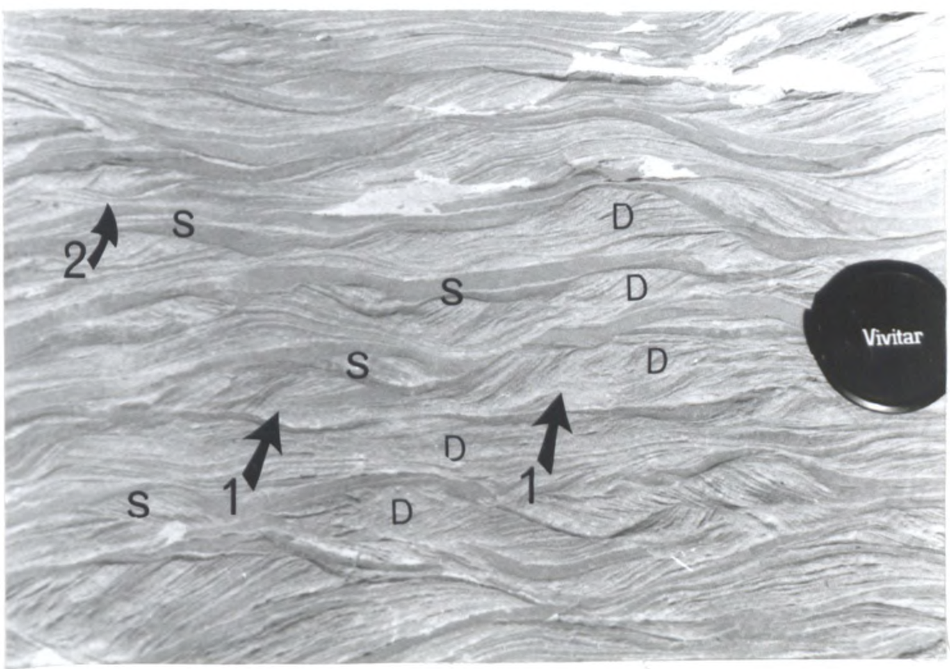


FIG 4.21 Bidirectional current ripple lamination showing wave modification. Bidirectional ripple sets show an alternation of dominant (D) and subordinate (s) flow sets. Internal erosion surfaces (1), draped with parallel lamination (2) define bundles of lamination within a number of the dominant flow sets suggesting wave modification of the ripples during their formation.

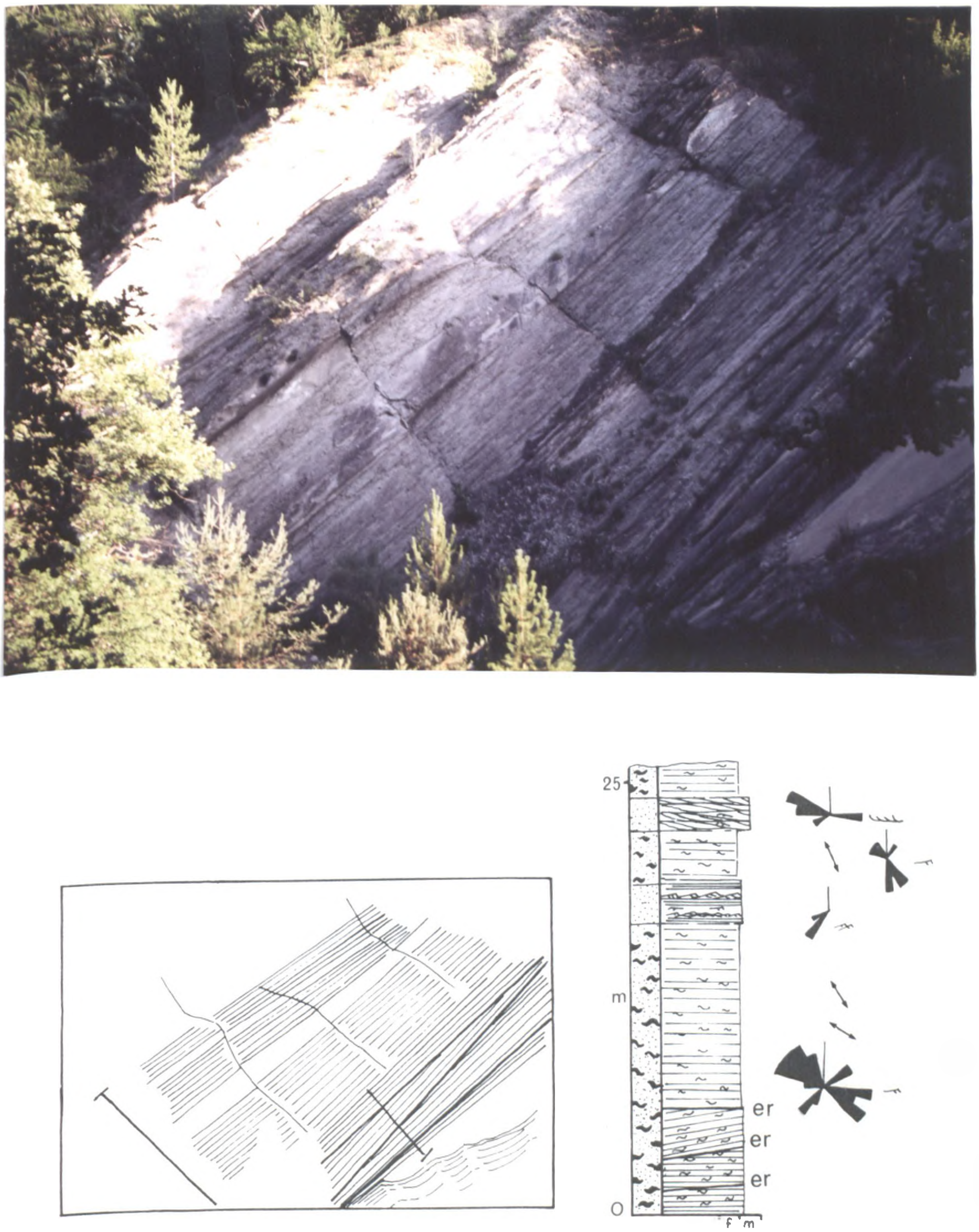


FIG 4.22 Cliff exposure of heterolithic facies showing the developement of low angle sets of heterolithic flaser (wavy) bedding defined by erosion surfaces (er) dipping at 5-8° and attributed to deposition on the inner accretional margins of inshore tidal channels.

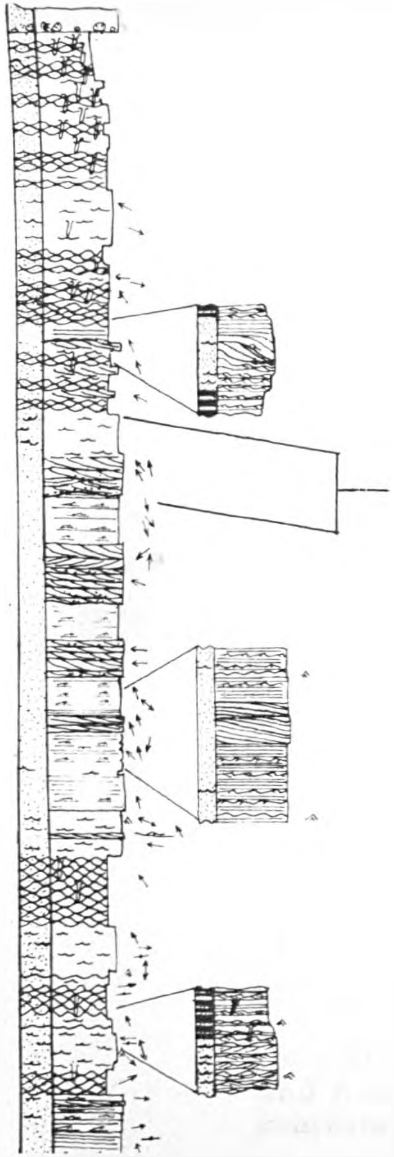


FIG 4.23 Small scale vertical sequences within the heterolithic facies of the inshore tidal channel and shoal facies association, St. Symphorien section, member Bm2.

Coarsening-upward sequence; linsen bottomsets -> cross-stratification

INTERPRETAION: Increased tidal current.
megaripple migration.

Fining-upward sequence; silt base lag -> trough cross-stratification -> flaser and planar lamination -> linsen bedding.

INTERPRETATION: Small tidal channel fill.

Coarsening-fining upward sequence; flaser / planar and current-ripple laminated sandstone -> trough cross-stratification -> flaser bedding.

INTERPRETATION: Increased tidal current.

Coarsening-fining upward sequence; linsen -> flaser -> linsen bedding.

INTERPRETATION: Increased tidal current.
neap-spring cycle (?) where shows thickness **periodicity** of ~ 28.

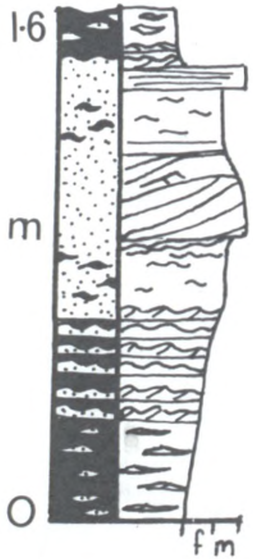


FIG 4.24 Trough cross-stratified form sets isolated within wavy and flaser bedding of the heterolthic facies generate small scale coarsening-upward sequences.

FIG 4.25 Small scale, random fining-upward sequence within the heterolithic facies passing from flaser through linsen bedding into silt streaked mudstone.

FIG 4.26 Groove cast base to graded storm sandstone sheet. Lens cap 5cm diameter for scale.

FIG 4.27 Dm scale unit within the heterolithic facies dominated by graded sandstone beds (1-6) interpreted to be the deposits of waning, storm generated currents. Beds have a sharp groove cast base, with the grooves commonly accentuated as load structures. They fine upward passing from planar or low angle lamination (p) into offshore directed climbing ripple lamination (cr) with the top commonly showing current ripple lamination (r). Beds are seperated by siltstone intervals with opposed flow (onshore directed) current ripple lamination (o-r). Scale bar 5cm length.

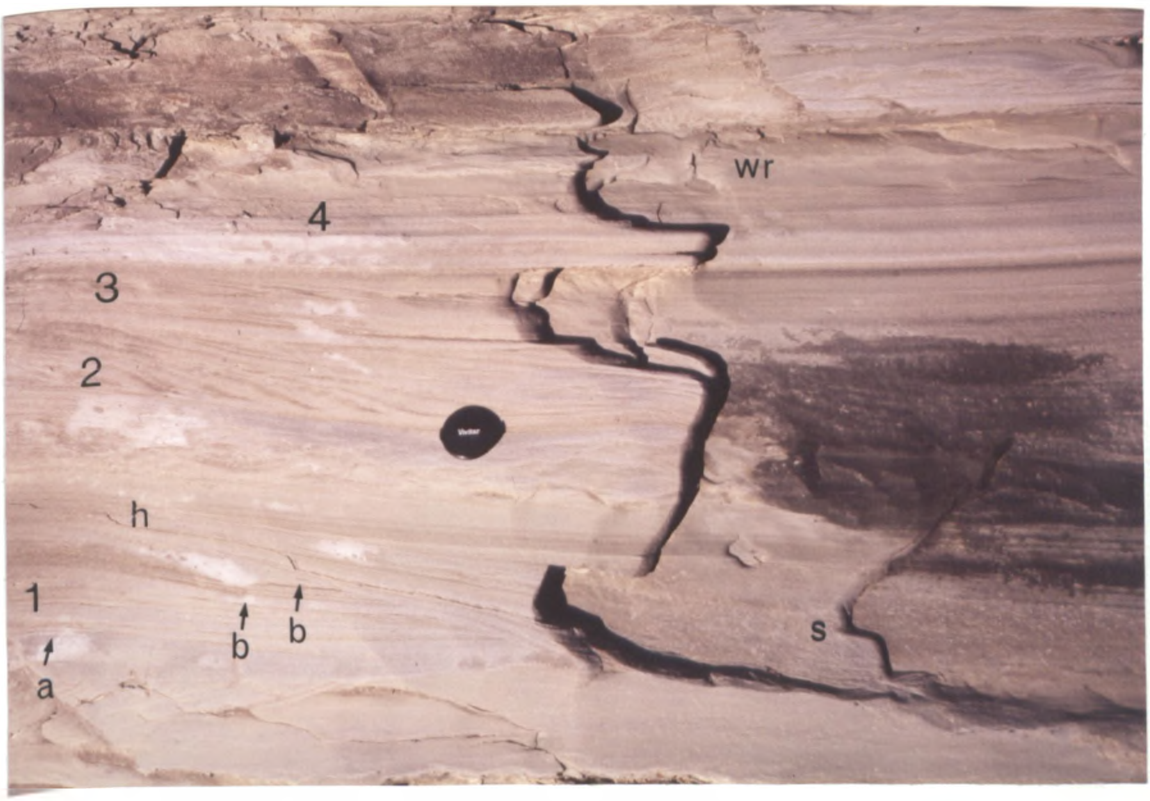


FIG 4.28 Graded, storm sandstone beds (1-4) with planar and climbing-ripple lamination, overlain by interval of bidirectional, wavy bedded ripple lamination (wr) within the heterolithic facies. The lower sandstone bed (1) has an erosive base (a) above which swaly and hummocky cross-stratification defined by second order erosion surfaces (b) is developed. Fine sandstone laminae thicken into swales (s) and thin onto the hummocks (h). Lens cap 5cm diameter for scale.

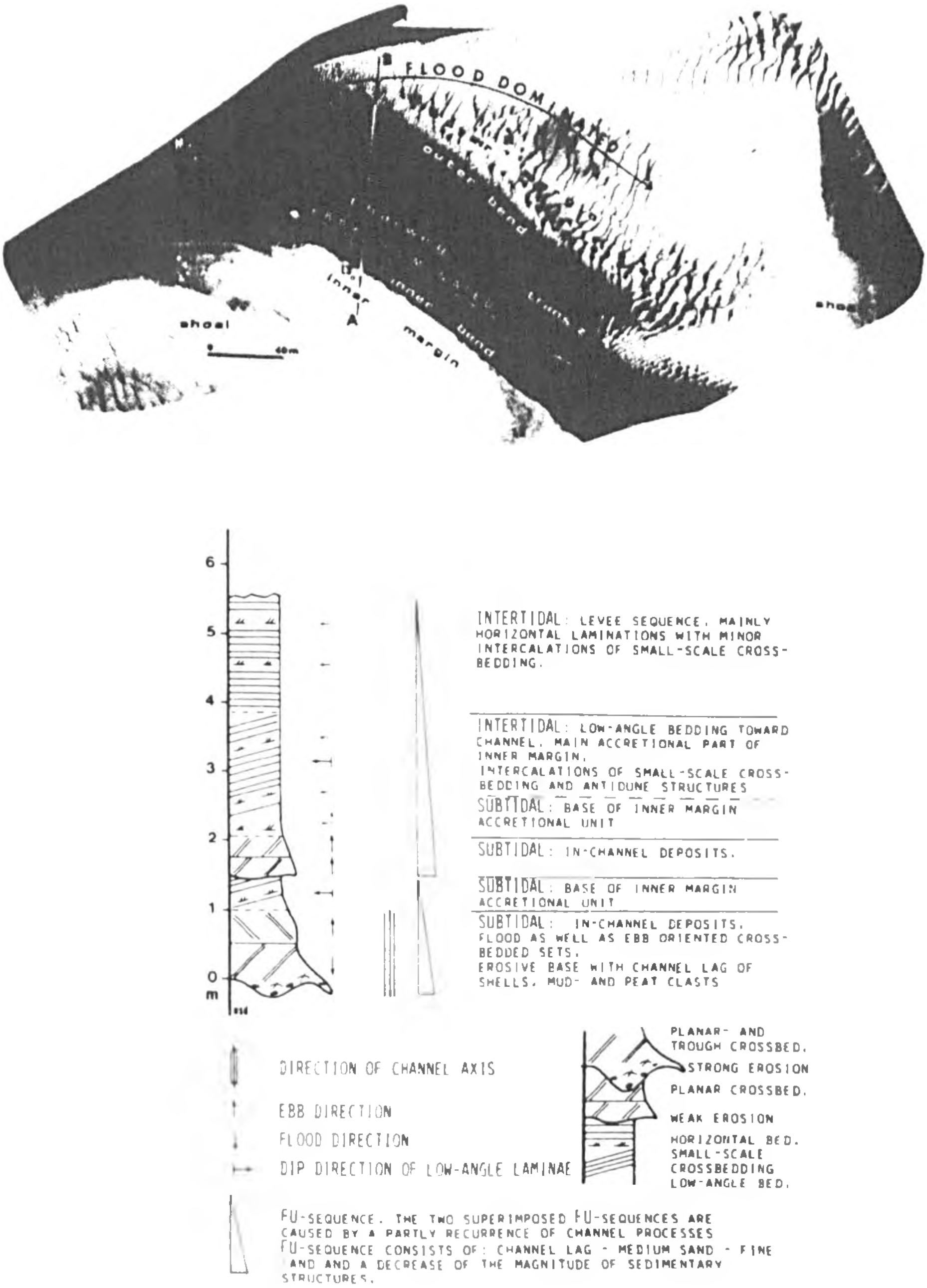
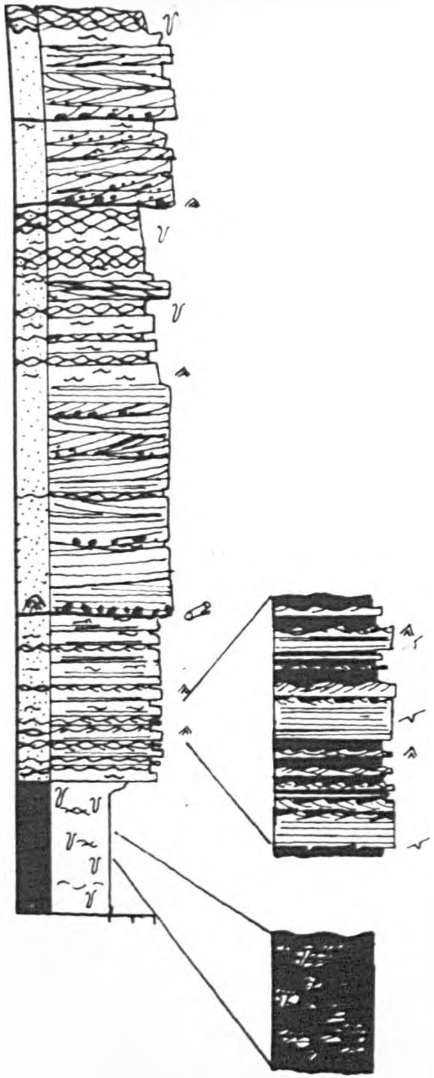


FIG 4.29 Shoal and intershoal deposits of the meso-tidal Scheldt estuary, Dutch coast (i) plan view of shoal and intershoal channel (ii) Reconstucted fining-upward channel - shoal sequence generated by lateral migration of intershoal channel (from Nio *et al.* 1980).

Fig 4.31 (i) Well developed offshore to inshore tidal channel and shoal coarsening-upward sequence. TC:Multistorey tidal channel complex (storeys 1-3) of the channelised sandstone facies Hf:Tidal shoal heterolithic facies Mf:Offshore muddy sandstone facies. Note how the tidal channel passes gradationally out into the shoal facies (arrowed).

FIG 4.31 (ii) Elevated view of multistorey tidal channel complex in Fig 4.3 (i). Vertically and laterally offset tidal channels (TC $^{1-3}$) overlie inshore shoal heterolithic facies (Hf) and offshore facies (Mf). Note how the channels pass out laterally into the shoal facies (arrowed as in Fig 4.3 (i). Interpretative sketch in Fig 4.31 (iv).

FIG 4.31 (i-iv) Offshore to inshore tidal shoal and channel sequence. St. Symphorien section, Vm1 member.



Channelised sandstone facies
Multistorey body comprising channelised fining upward sequences dominated by trough cross-stratification - bi-directional, silt drapes and flow reversal modification common.

Inshore tidal channels, infilled by 3-D sandwaves, tidal current reversals indicated by bidirectionality, and bedform modification. Note opposed palaeoflow dominance in channels indicating flow segregation.

Heterolithic facies
Linsen and wavy ripple laminated bedding, bidirectional, commonly modified and wave ripple sets. Planar laminated, graded sand sheets with climbing ripple sets.

Inshore tidal shoal, weak-moderate tidal currents, periodic storm generates offshore directed graded storm sandstones - ~parallel to ebb tidal currents.

Muddy sandstone facies
Intensively bioturbated muddy sandstone with irregular patches of heterolithic bedding - wave and tidal ripple lamination. Shell beds. Open marine macro-fauna and trace fossil.

Offshore, low wave and tidal energy, above storm wave base.

FIG 4.31 (iii) Descriptive log of offshore to inshore tidal channel and shoal coarsening-upward sequence in Fig 4.31(i-ii). Inserts show (a) the preservation of patches of heterolithic bedding within the offshore facies (b) intimate interbedding of graded storm sheet sandstones, and predominantly tidal, heterolithic bedding.

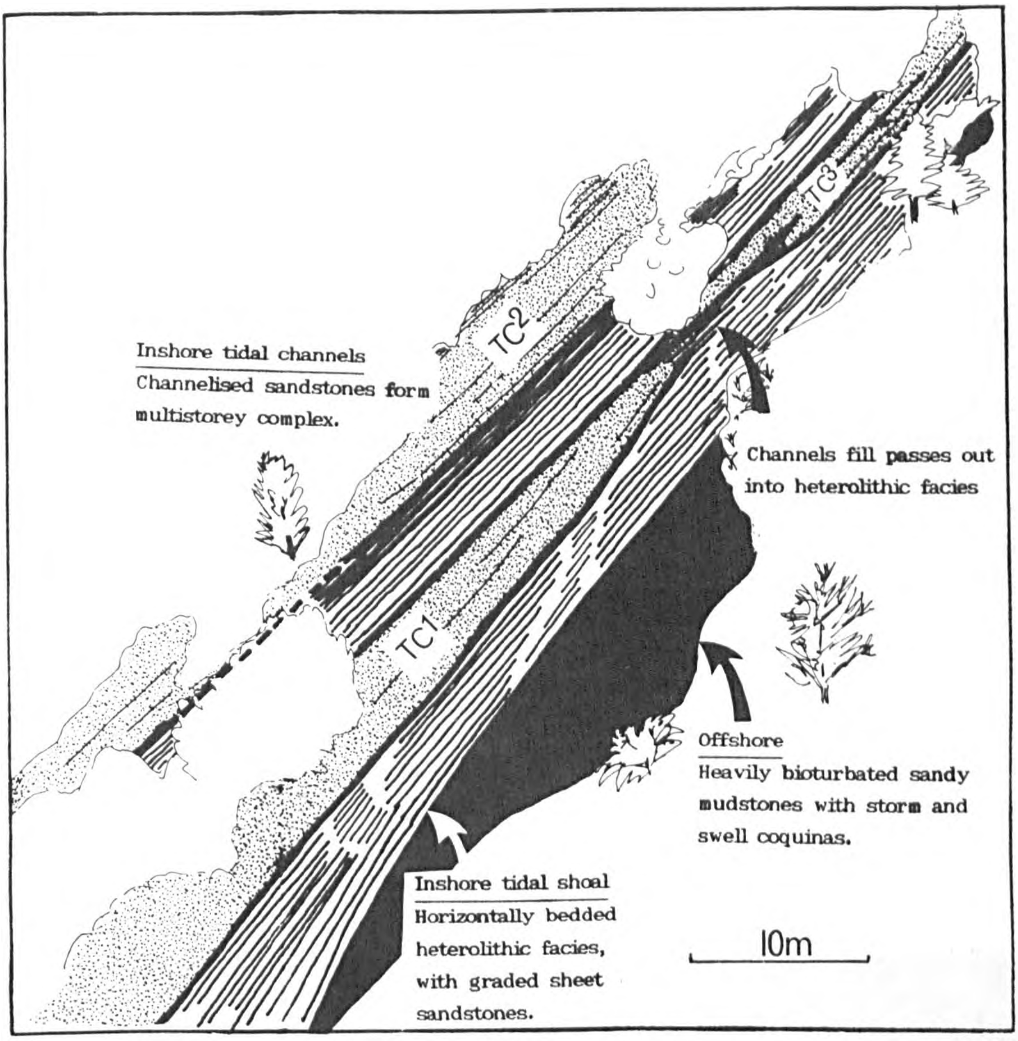


FIG 4.31 (iv) Interpretative sketch of offshore to inshore multistorey channel complex Fig 4.31 (ii).

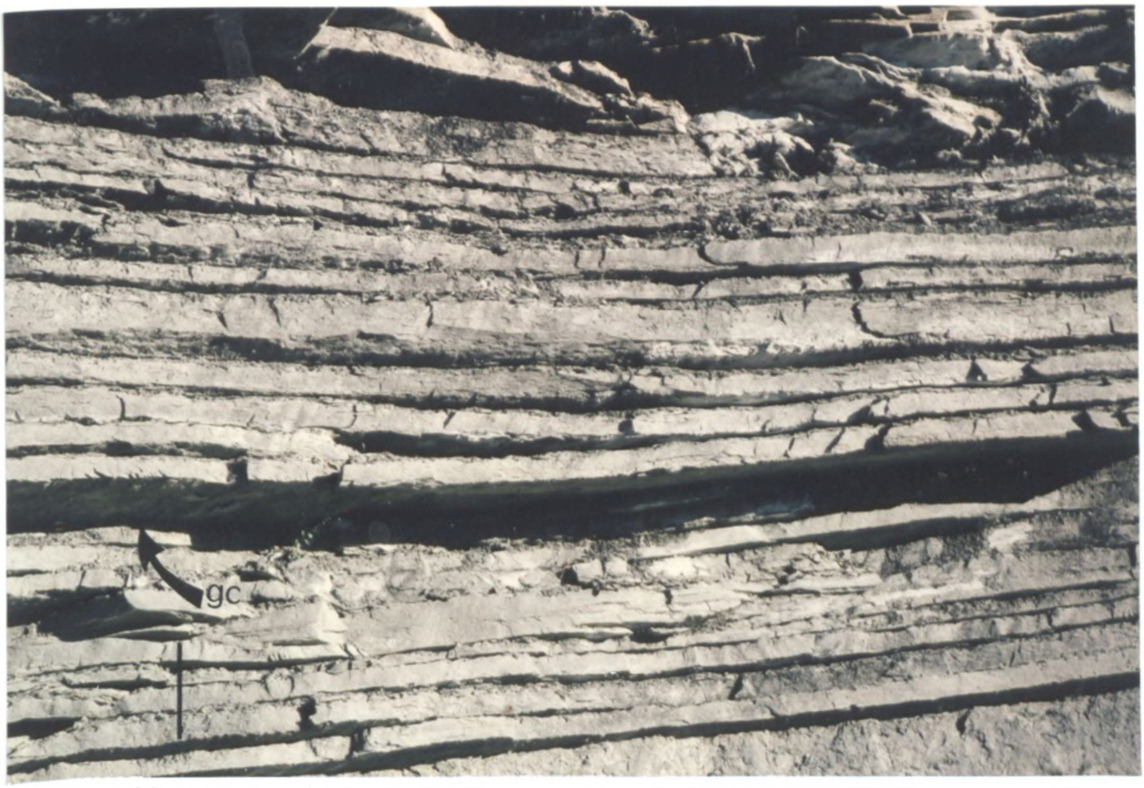


FIG 4.32 Erosive based, graded sandstone beds with groove casts (gc) randomly interbedded within wavy-linsen bedding (recessed due to weathering) of the heterolithic facies. The sandstone beds are interpreted to be the deposits of waning, storm generated currents. They define the base of small scale (cm-dm scale) fining-upward sequences passing from planar laminated sandstone through flaser to linsen bedding which may be attributed to decreasing rates of sand transport over a number of tidal cycles following a storm event. Scale bar 0.5m length.

FIG 4.33 Heavily bioturbated muddy sandstone facies. Note the preservation of a relict planar stratification.

FIG 4.34 Close up of muddy sandstone facies showing the preservation of a set of wave ripple lamination (Wr) within muddy sandstone heavily bioturbated by 'u-shaped' *Diplocraterion burrows* (D). Lens cap 5cm diameter.

FIG 4.35 Composite swell lag coquina within the muddy sansdstone facies. Pecten bivalve shells preferentially orientated convex-upward in their most hydrodynamically stable position. Coquinas interpreted to record concentration of epi-fauna under storm wave swell. Ruler 20cm length.

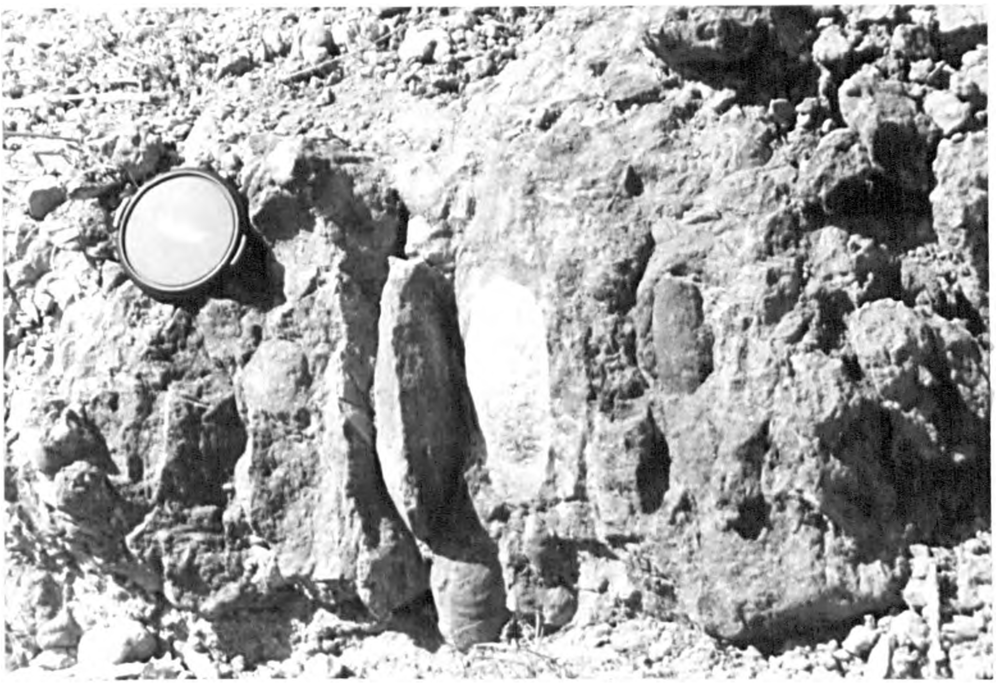
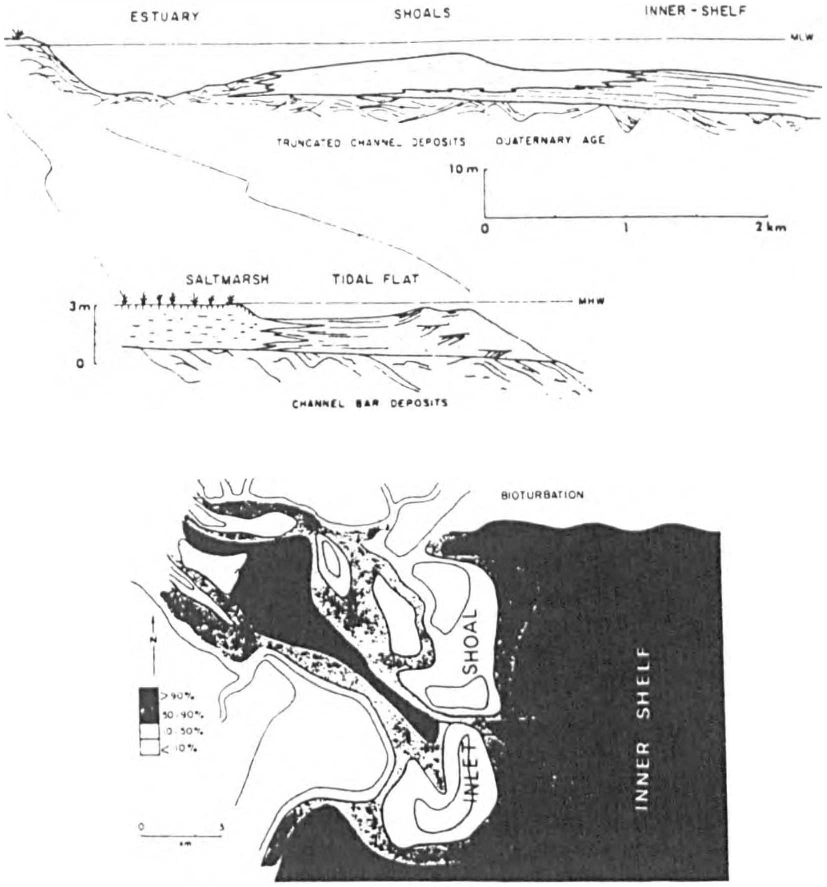
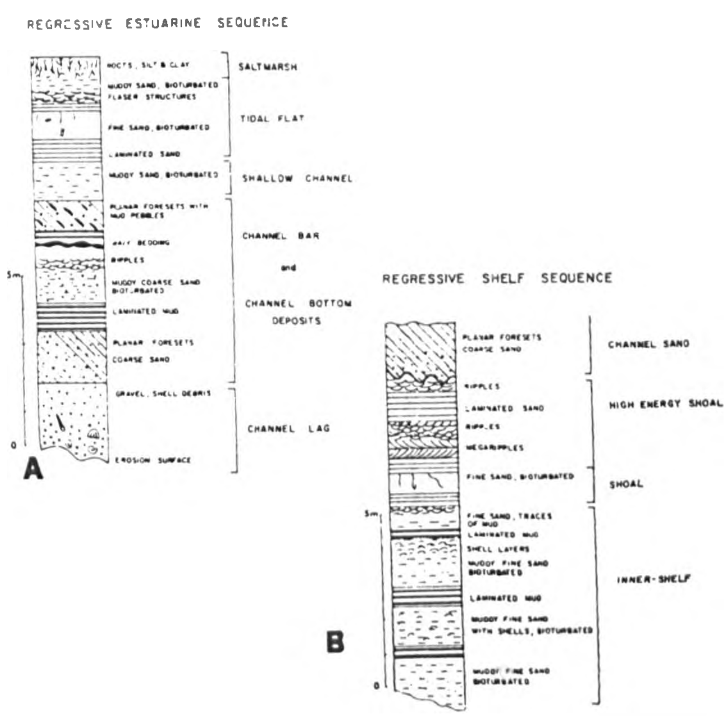


FIG 4.36 *Pholadomya* bivalves in life position within vertical, 'escape' burrows through a graded, storm shell layer in the muddy sandstone facies.



(i) Diagram showing (a) the lateral and vertical relationship of nearshore sandbodies - stippled areas are inlet shoal sandbodies, and (b) the increasing level of bioturbation in passing from inlet shoal to offshore. Note how the inlet shoal passes offshore into inner shelf deposits, which comprise intensively bioturbated muddy sandstones as in the muddy sandstone facies of this study.



(ii) Regressive sequence constructed from the sedimentary facies of Ossabaw Sound. A:Estuarine B:Inner-shelf and inlet shoal. Note that all of the facies contacts are gradational except for the base of the estuarine channel.

FIG 4.37 (i-ii) Inshore estuarine tidal inlet and shoals of Ossabaw Sound, Georgia (from Greer 1975).

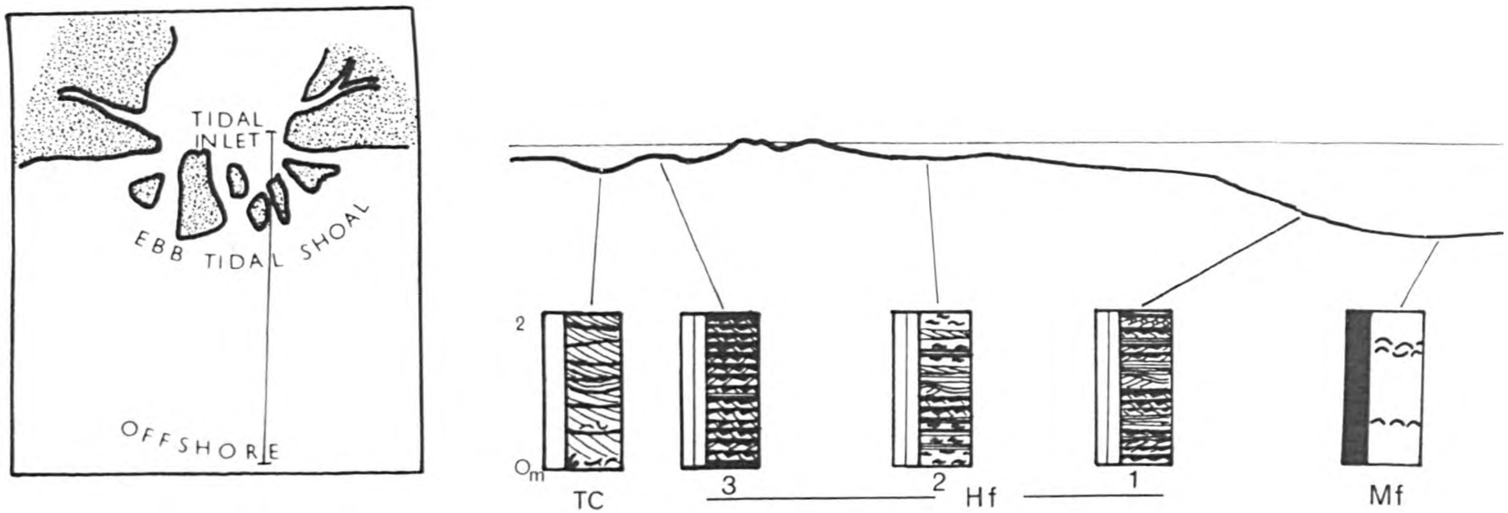
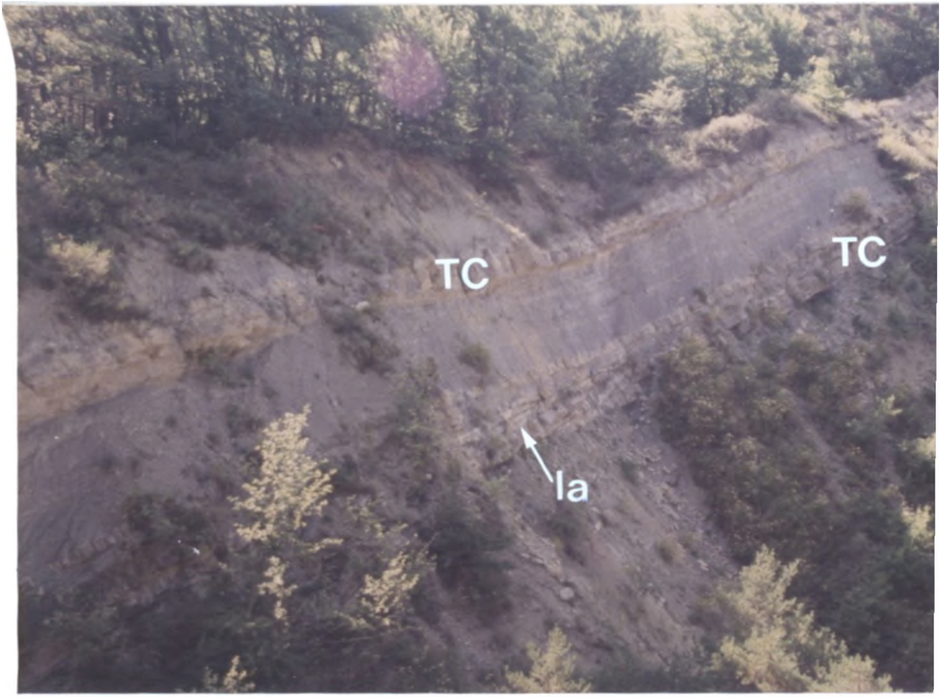
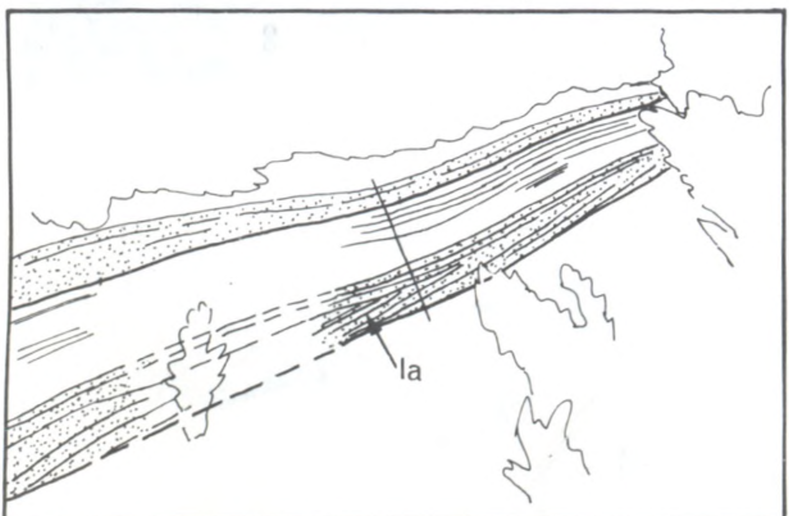


FIG 4.38 Schematic representation of the interpreted depositional settings of the facies of the Offshore-Inshore tidal channel and shoal association. (profile after Greer 1975).
TC: Channelised sandstone facies - trough cross-stratified channel fills.
Hf:Heterolithic facies of tidal shoal. 1 - wavy and linsen bedding with wave ripple sets and graded storm beds in outer shoal.
2 - flaser-linsen bedding with planar lamination and trough-stratification in inner shoal.
3 - Wavy-linsen bedding on margins of intershoal channels, generation of low angle sets of heterolithic lateral accretion bedding.
Mf:Muddy sandstone facies- heavily bioturbated muddy sandstones with shell layers.


TC
TC
Ia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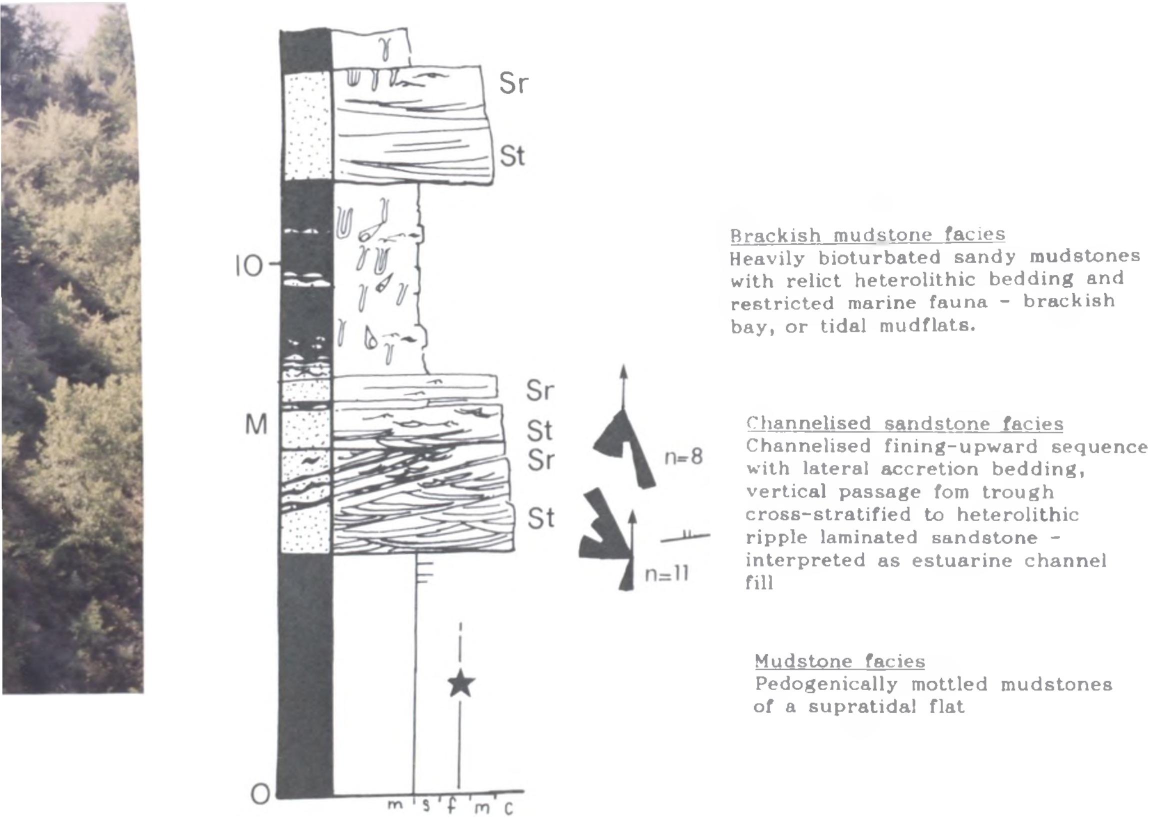

FIG 4.39 Photograph and graphic log of fining upward sequences of the upper estuarine facies association, Vm2 member St.Symphorien section. Tidal channel sandstone bodies (TC) with lateral accretion bedding (la) fine upward into heterolithic tidal flat facies. Note the bidirectional fill to the multistorey channel.


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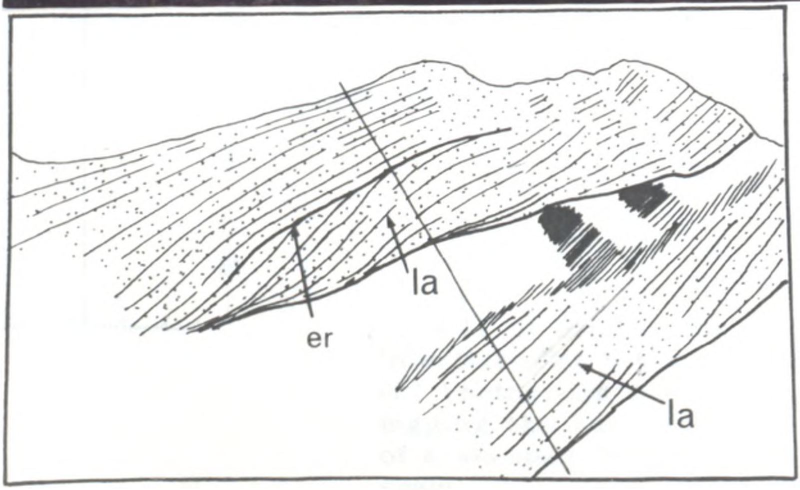
la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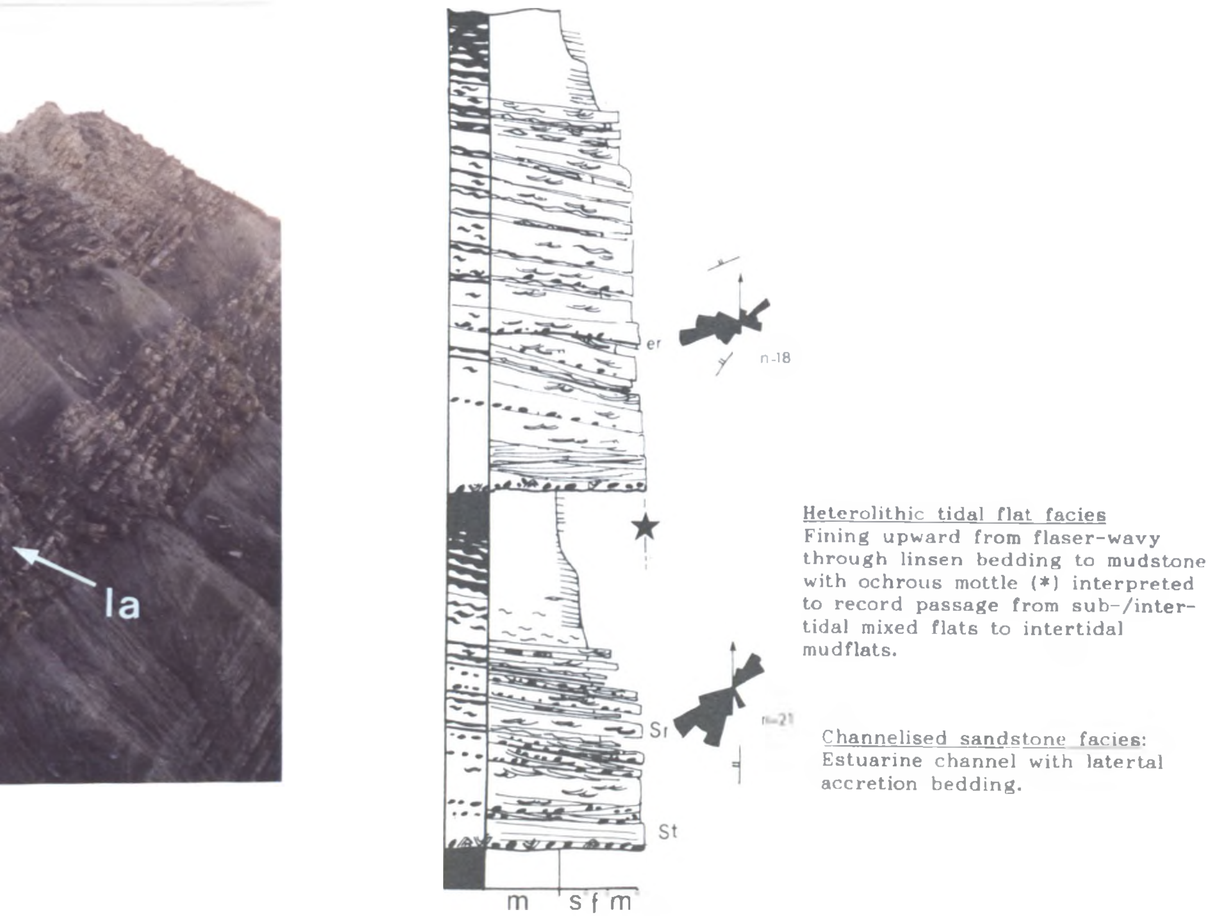
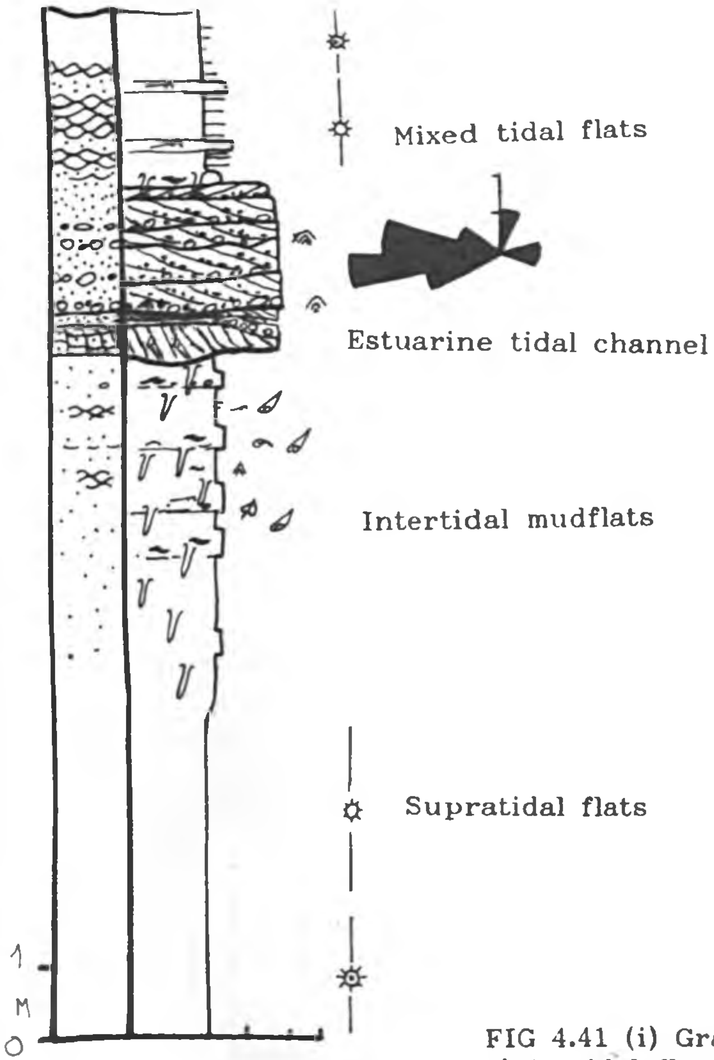
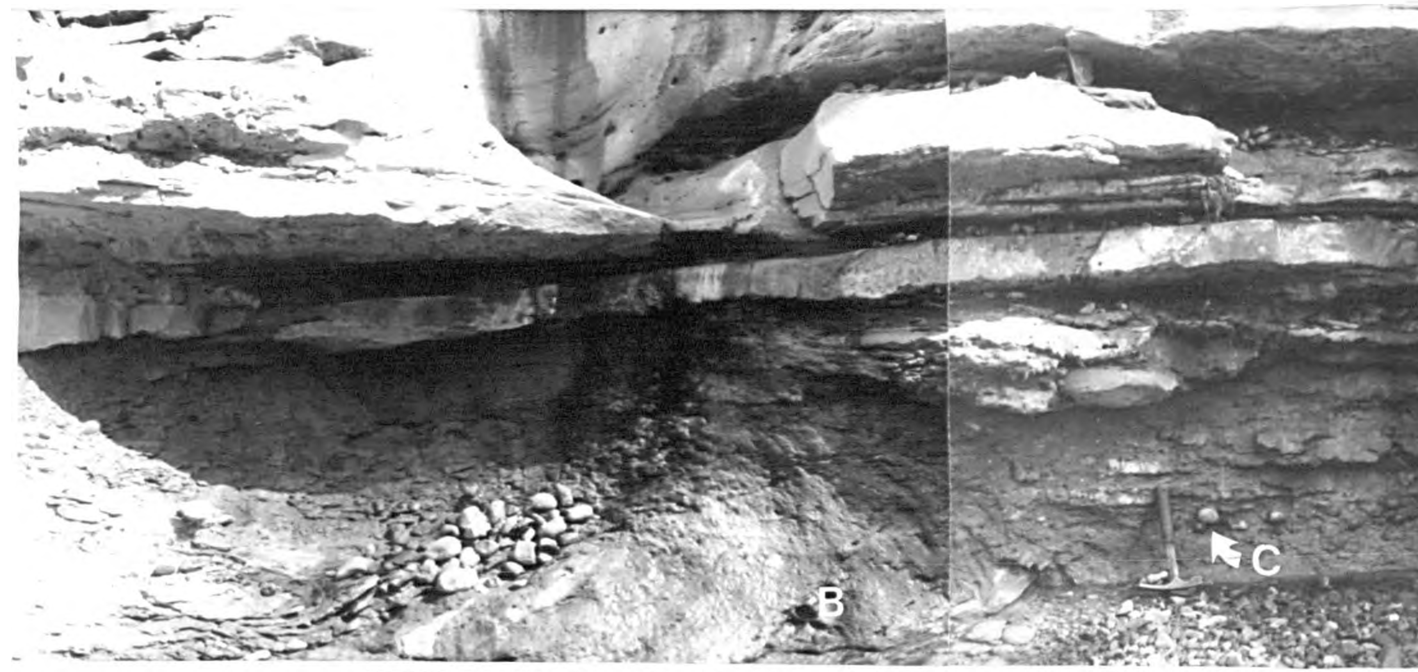


FIG 4.40 Lateral accretion bedding (la) within tidal channel bodies of the channelised sandstone facies, upper estuarine facies association. Channel bodies fine upward into heterolithic tidal flat facies which dip gently into the channel. Note the developement of an erosional surface within the lateral accretion bedding (er) attributed to erosion during exceptional discharge conditions. Auribeau section, Vm2 member.

Maudwech





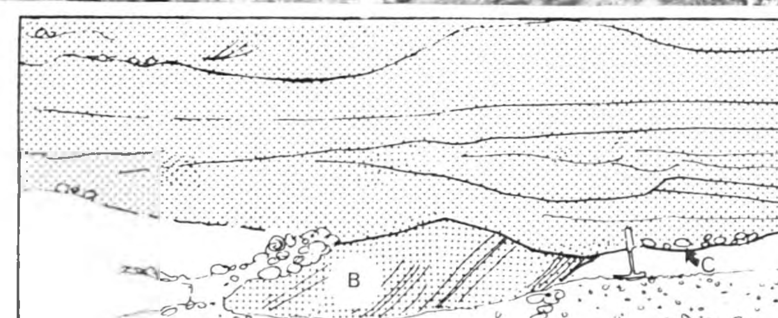


FIG 4.41 (i) Graphic log of fining-upward estuarine channel to intertidal flat sequence. St.Symphorien section, member vm2. (ii) Photogragh of tidal channel base. Note the cobble clasts (c) lagging the basal channel erosion surface, and the developement of a slumped block (b) of heterolithic bedding. Hammer 35 cm for scale.

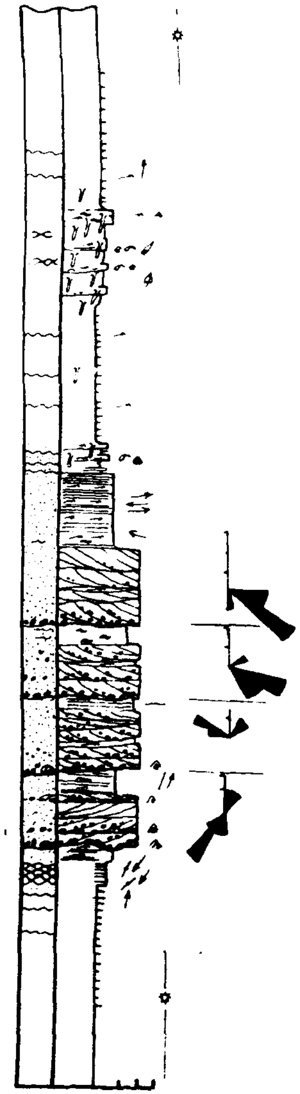
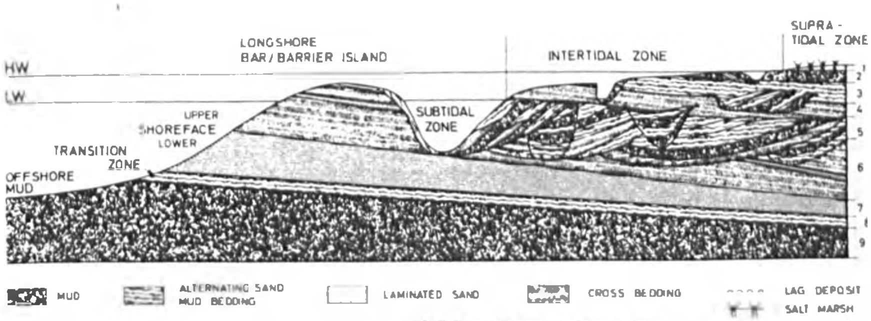


FIG 4.42 Fining-upward sequence of the Upper Estuarine facies association. St. Symphorien section, member Vm1.

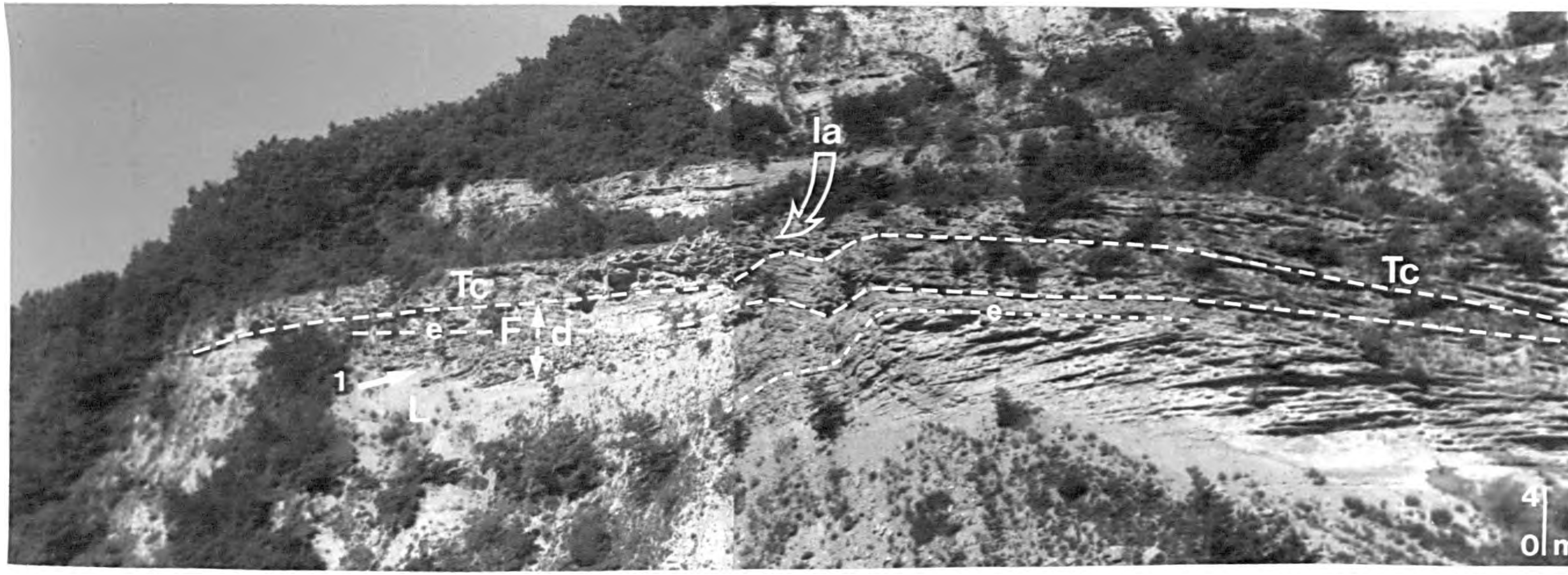
FIG 4.42 (i) View of planar laminated sandstones which form the upper part of an upper estuarine, tidal channel fill fining-upward sequence.

FIG 4.43 Erosive margin (arrowed) of small tidal channel/gully with heterolithic ripple laminated fill within heterolithic facies of the upper estuarine facies association. Bedding is overturned and dips stepply toward the right. Trowel (t) is 20cm length for scale.



1 *Salt marsh (supratidal zone).* Very fine sand and mud, interbedded shell layers, plant roots, irregular wavy bedding. 2 *Mud flat (intertidal zone).* Mud, occasional very fine sand layers, lenticular bedding with flat lenses, strong bioturbation. 3 *Mixed flat (intertidal zone).* Sandy mud, thinly interlayered sand mud bedding, lenticular bedding, flaser bedding, shell layers, bioturbation strong to weak. 4 *Sand flat (intertidal zone).* Very fine sand, small-ripple bedding sometimes of herringbone structure, flaser bedding, laminated sand, occasional strong bioturbation. 5 *Subtidal zone.* Medium to coarse sand, mud pebbles, megaripple bedding, small-ripple bedding, laminated sand, weak bioturbation. 6 *Upper shoreface.* Beach bar- and ripple cross-bedding, laminated sand, weak bioturbation. 7 Lower shoreface. Laminated sand, bioturbation stronger than in the upper shoreface. 8 *Transition zone.* Alternating sand mud bedding, i.e., flaser and lenticular bedding, thinly and thickly interlayered sand mud bedding, moderate bioturbation. 9 *Shelf mud.* Mud with storm silt layers, moderate bioturbation.

FIG 4.44 Schematic cross-section across tidal flats of the North Sea in a hypothetical prograding shoreline showing the vertical organistion of sub-, inter-, tidal flat facies (from Reineck & Singh 1980).



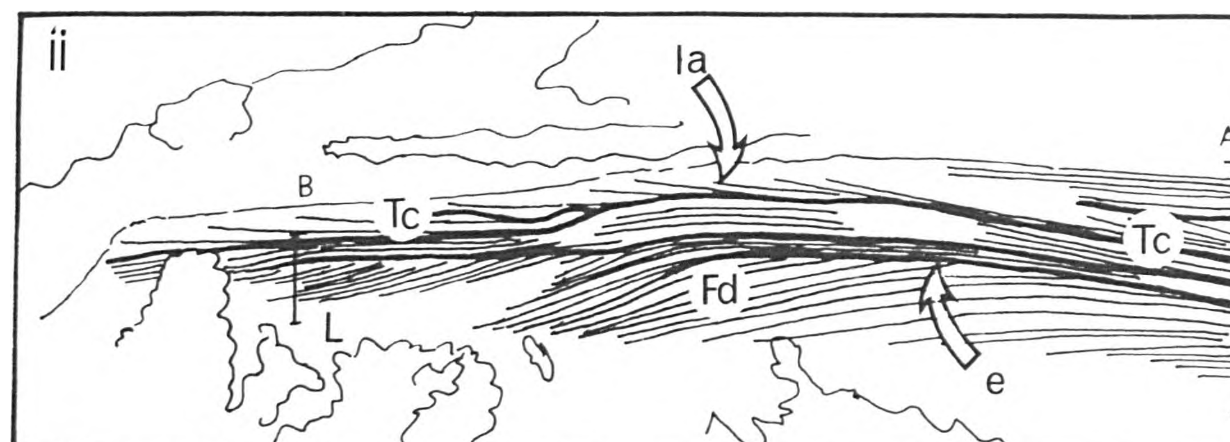
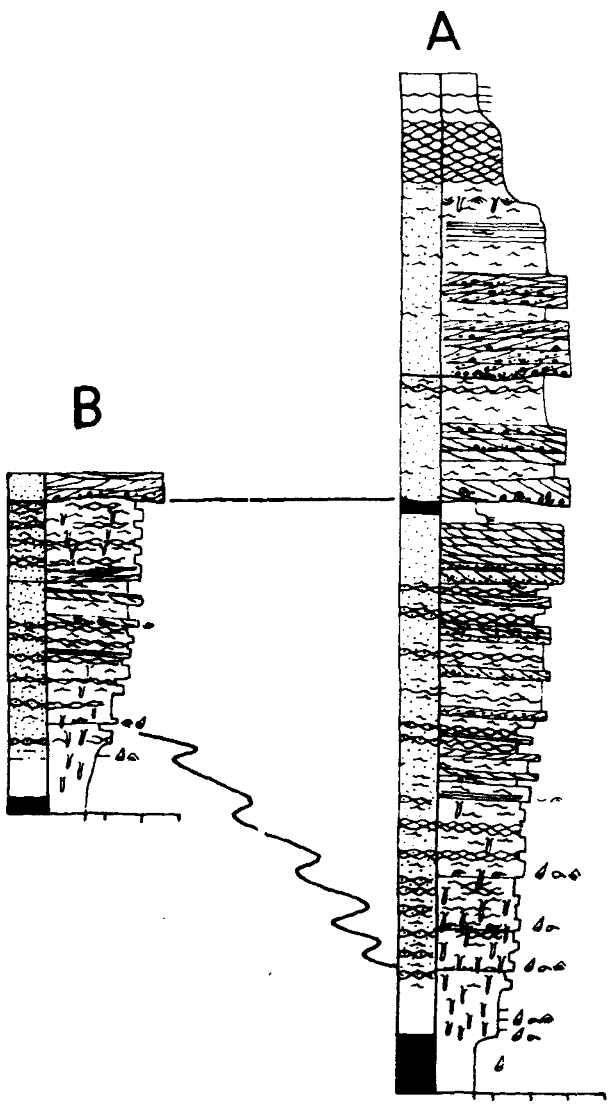


FIG 4.45 (i-iii) Plate (i) and interpretative sketch (ii) of flood tidal delta heterolithic facies (Fd) and associated flood tidal channels (Tc) (channelised sandstone facies) gradationally overlying lagoonal bay brackish mudstone facies (L) in a coarsening upward sequence. Internally the flood delta facies comprises large scale, inclined stratification, arranged into sigmoidal units by major erosion surfaces (e). Note the progressive vertical climb of the delta base as it migrates northward (landward). A and B mark position of graphic logs of Fig 4.45 (iii). -> marks position of Fig 4.50.

FIG. 4.45 (iii) Graphic logs through lagoonal bay - flood delta facies sequence. Member Vm2, St. Symphorien.



Channelised fining-upward sequences of trough cross-stratification and heterolithic ripple bedding. Bidirectional, flood dominated, mud drapes, erosional pause planes common	Migration of 3-D tidal megaripples in channel, channel abandoned/migrates.	Inlet mouth, "feeder channels".
Inclined sets of trough/planar cross-stratified sandstone, heterolithic bedding, bidirectional flood dominated	Migration of 2-D/3-D tidal megaripples alternates with ripple developement	Proximal flood-delta
Inclined sets of linsen-flaser bedding, bidirectional lamination flood dominated. Graded sandstones with shell lags.	Low energy tidal currents. Periodic waning flow event during storm washover, or off flood delta.	Distal flood-delta (bottomsets)
Brackish mudstone Heavily bioturbated mudstones with brackish - open marine fauna, thin graded shell layers.	Negligible tidal and wave currents, periodic introductioin of shell layers from flood-delta/washover.	Lagoonal Bay proximal to flood delta.

(i)

(ii)

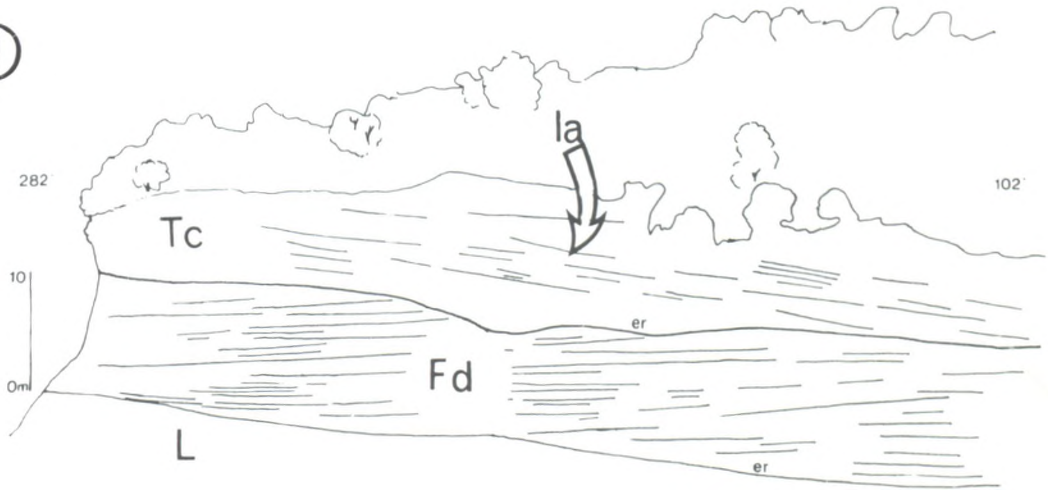
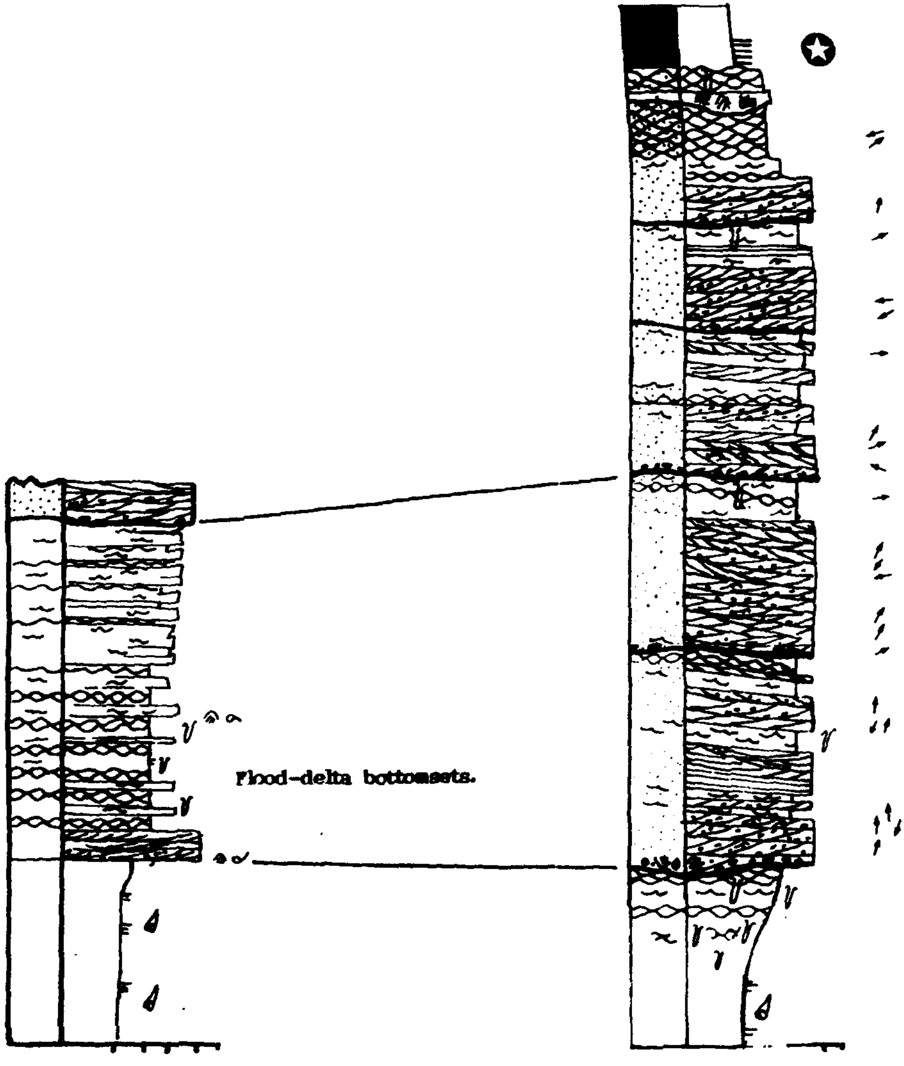


FIG. 4.46 Plate (i), interpretative sketch (ii) and graphic logs (iii) of a lagoonal bay (L) to flood delta (Fd) facies sequence. Member Vm2. St. Symphorien. Note (a) the erosive base to the flood tidal channel (Tc) and developement of lateral accretion bedding (la) (b) cross-stratified sandstone of flood tidal channel (log B) passes northwards (toward left) into inclined, heterolithic bedding of flood delta (log A).


Flood-delta bottomsets.

Pedogenically mottled linsen bedding and mudstone.

Supratidal flats.

Bidirectional linsen bedding, small channels filled with oysters.

Mixed intertidal flats dissected by tidal gully.

Repeated channelised fining-upward sequences with lateral accretion bedding above lagged erosion surfaces. Cross-stratified sandstone passes laterally into flaser-linsen bedding. Flood dominated palaeoflow, mud drapes and erosional pause planes common.

Vertically stacked, flood tidal "feeder" channels, laterally migrating.

Trough cross-stratified sets overlie channel base.

Mudstone with brackish fauna

Lagoonal Bay

FIG 4.46 (iii) Lagoonal bay to flood tidal coarsening-upward sequence, overlain by fining-upward passage into tidal flat facies of the Upper Estuarine facies association. Multistorey flood tidal channel body erosionally overlies lagoonal bay mudstones. Note how channel fill passes out laterally (landward) into heterolithic flood delta deposits.

FIG 4. 47 Lagoonal bay mudstone with monospecific, turretelid epifauna (arrowed). Lens cap 5 cm.

FIG 4.48 Heavily bioturbated, heterolthic bedding at the gradation between the flood delta and lagoonal bay facies. The coarser sandstone beds are the bottomsets to the flood delta, and are seperated by lower energy, lagoonal bay deposits. Lens cap 5cm.

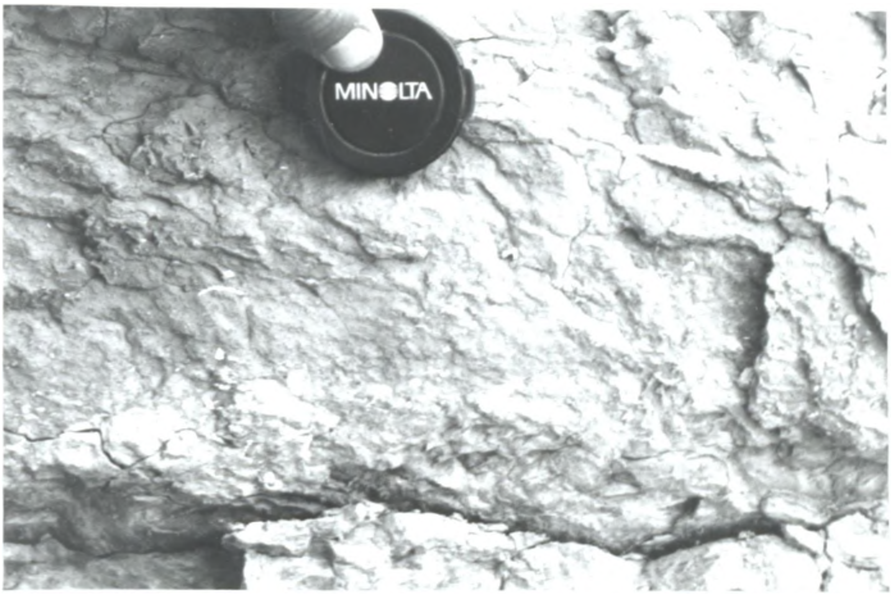


FIG 4.49 Graded, 'washover' sandstone with mudstone rip-up clasts and mixed marine bioclasts. Note epifaunal turretelid shells in background lagoonal mudstone.

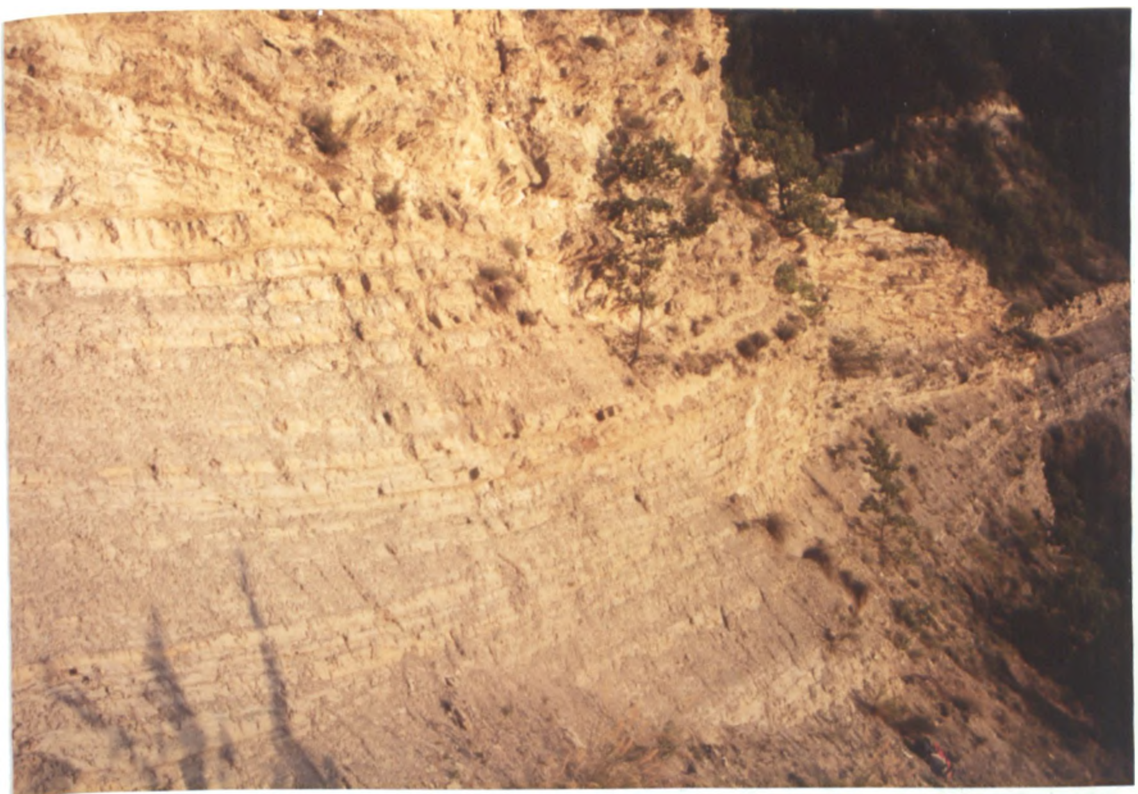


FIG 4. 50 Close up of heterolithic facies which comprises inclined, tabular beds of current ripple laminated sandstone seperated by heterolithic intervals.

FIG 4.51 (i) Spring-Neap-Spring (S-N-S) tidal bundle sequence within tangential, trough cross-stratified set of the channelised sandstone facies. Ruler 20cm length.

FIG 4.51 (ii) Close up of Neap phase of tidal bundle sequence showing thin bundles with closely spaced siltstone drapes.

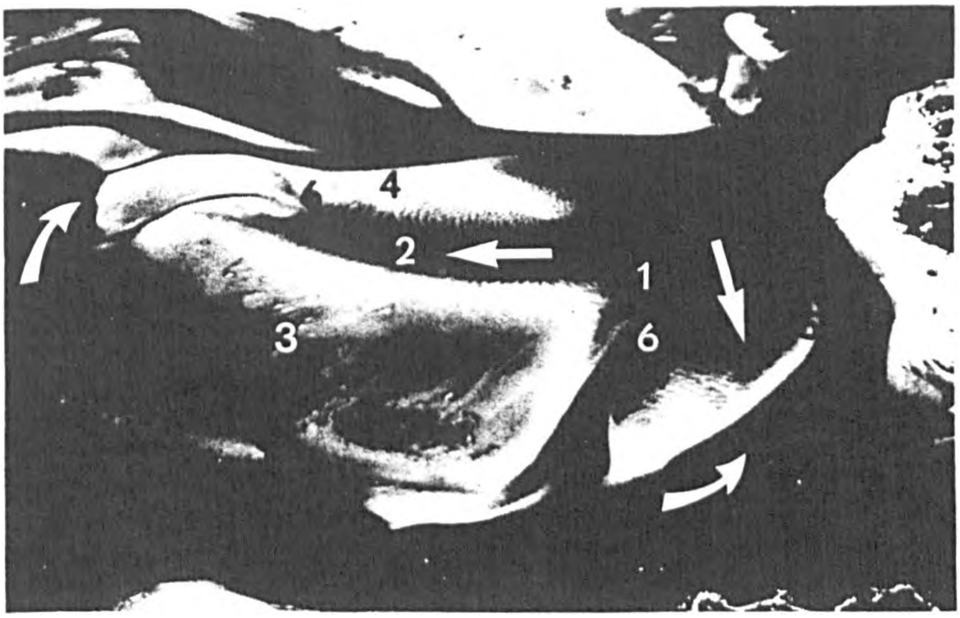
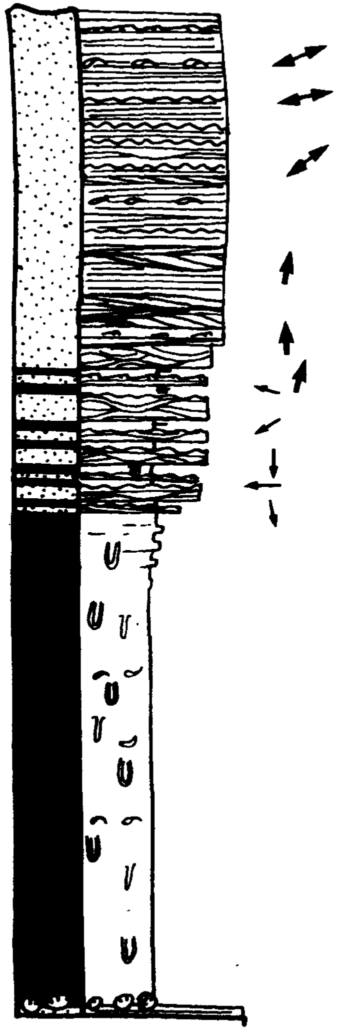


FIG 4.52 Flood tidal delta morphology. Note in particular the position of the flood channels overlying the flood delta.
1 = flood ramp, 2 = flood channel, 3 = clam flat,
4 = ebb shield, 5 = ebb sppit, 6 = spillover lobe.
(from Hayes 1980).



DESCRIPTION

INTERPRETATION

Planar Sandstone Facies
Dominated by planar laminated sandstone, wave ripple lamination, amalgamated hummocky cross-stratification, "gutter-cast" scours, rip-channels. Onshore directed trough-stratification. Convex upward shell layers

Shoreface - Foreshore:
Shoaling wave action above fair-weather wave base. Periodic storm scouring.

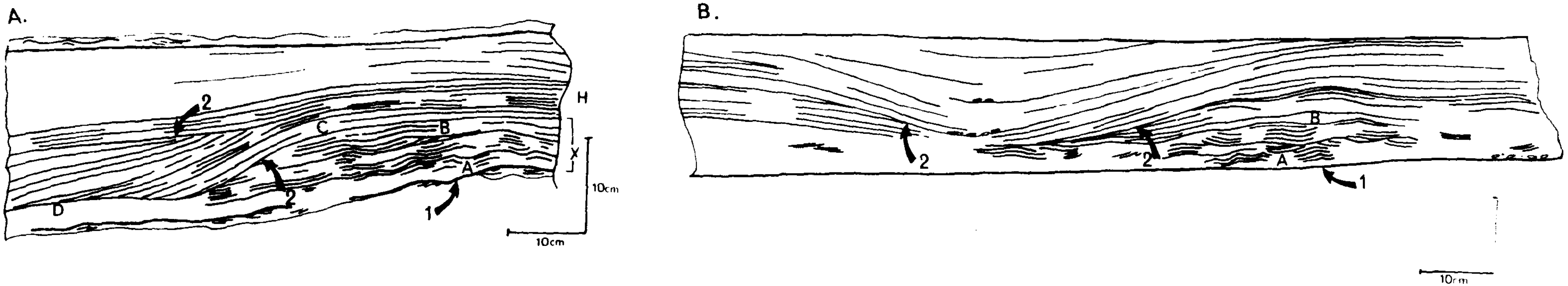
Sandstone Sheet Facies
Graded sandstones with hummocky cross-stratification or planar stratification -> ripple drift lamination. Ripple reworked tops. Heterolithic wave ripple laminated intervals.

Transition Zone: Alternation of storm and fairweather processes. Offshore directed, unidirectional and combined-flow storm currents.

Muddy Sandstone Facies
Heavily bioturbated muddy sandstone with relict patches of linsen bedding with wave and current ripple lamination. Open marine fauna, graded and convex-upward shell layers.

Offshore: Above storm wave-base. Storm wave swell and current shell layers.

FIG 4.53 Offshore to foreshore facies sequence.



Variation from the 'typical' H.C.S bed is shown in these two sections (orientated at 80° to each other) of a hummocky cross-stratified bed
The base to the bed (1 – 1st order erosion surface) is overlain by a symmetrical wave ripple laminated zone (X). The wave ripple sets show a vertical transition from small trochoidal, vortex ripples (A) to larger sinusoidal post-vortex ripples (B). In section B wave ripples change from vortex ripples of n=0.75cm, w=7.0cm (n/w=0.107) to post-vortex ripples of n=0.8, w=13.5 (n/w=0.059). Second erosion surfaces (2) cut down into the wave ripple laminated zone to produce a hummocky surface typically draped by concordant laminae (C), and to lesser degree by downlapping laminae (D). Laminae thicken into the hummock depressions progressively subdueing their topography toward the top of the bed.
Post-vortex (rolling grain ripples) n/w < 0.06. Vortex ripples 0.12 < n/w > 0.20. n = amplitude w = wavelength.

FIG 4.54 Interpretative sketch of hummocky cross-stratified set with wave ripple laminated base within the planar sandstone facies, Esclangon section, member Bm2.

FIG 4.55 Planar laminated sandstone facies. Planar laminated sandstone beds with tops reworked by straight crested, wave ripple sets. P.c.l in sandstone trends parallel to pencil line (circled). Pencil line 10cm length.

FIG 4.56 Sets of planar laminated sandstone with tops reworked by symmetrical wave ripple sets with bifurcating crest lines are traceable for tens of metres across the exposure. Note the paired *Diplocraterion* burrows (circled) emergent on the bedding plane. Lens cap 5cm.

FIG 4.57 Thin concentration of bivalve shells (casts) on bedding-plane surface within planar laminated sandstone facies. Note the predominantly convex upward form of the shells which is their most hydro-dynamically stable position. Lens cap 5cm.

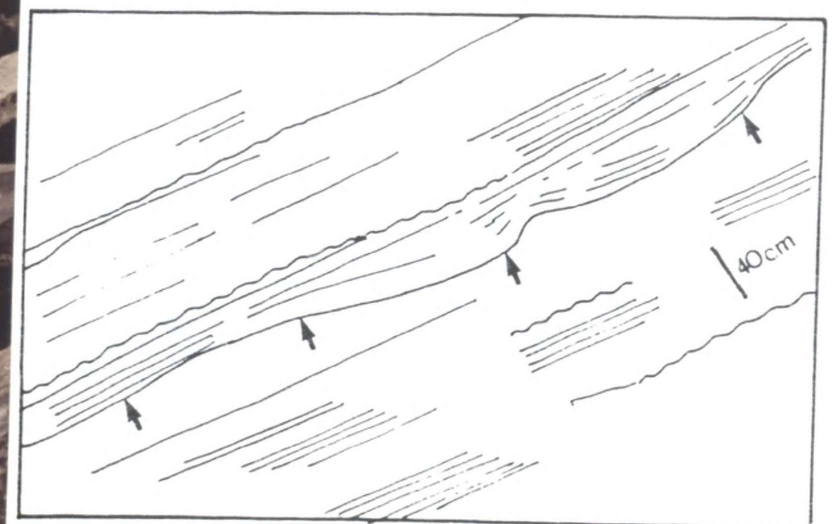


FIG 4.58 Erosional, swaly scour within predominantly planar laminated sandstones of the foreshore facies. Note the sharp, swaly erosive base (arrowed) to the prominent sandstone bed in the centre of the photograph.

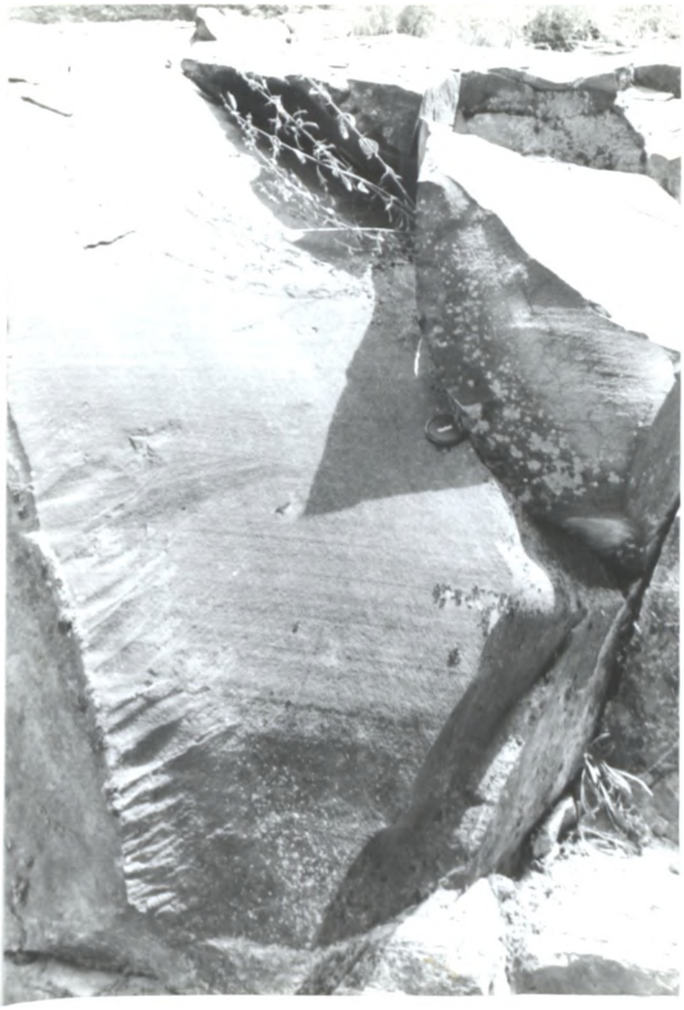


FIG. 4.59 Detailed view of internal stratification within amalgamated hummocky cross-stratification (swaly stratification), comprising low angle scours (1) draped by parallel laminae of well sorted sandstone. Note the light (mica-poor) and dark (mica-rich) mineral layering which defines the laminae. Lens cap 5cm.

FIG 4.60 Birds footprint ornamenting a wave rippled bedding surface within the planar sandstone facies.

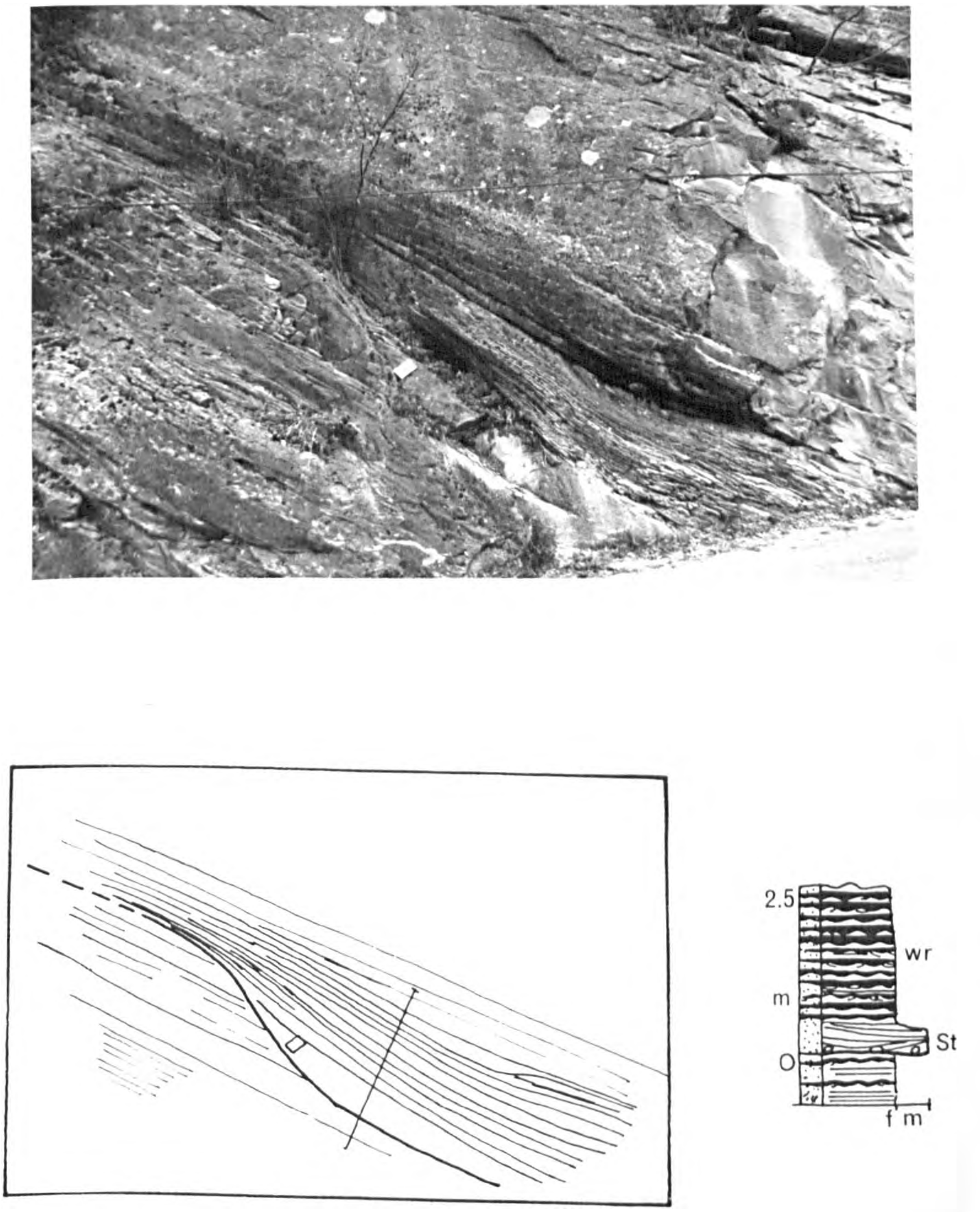
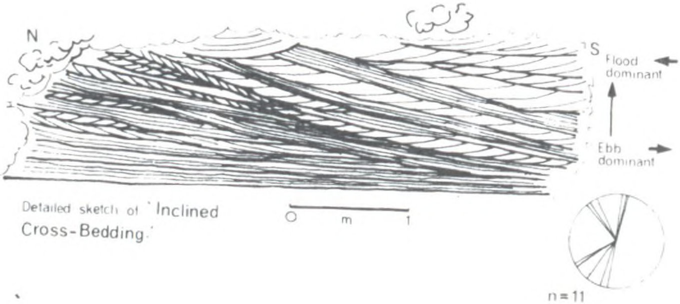
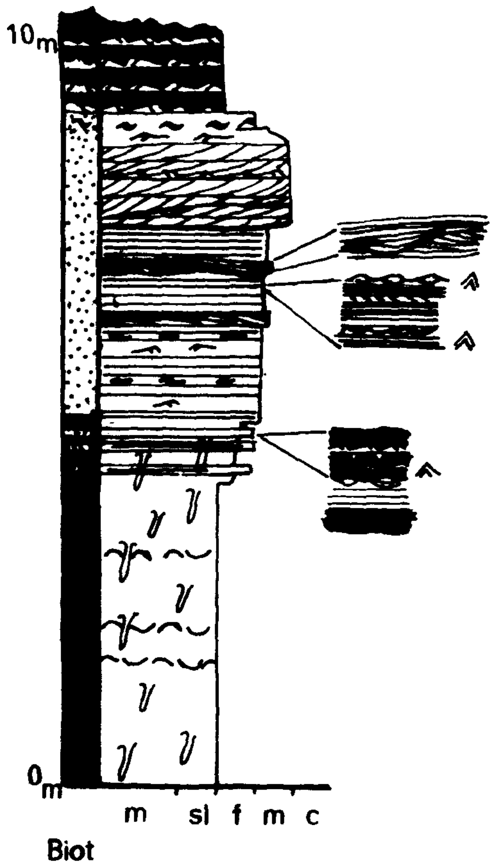


FIG 4.61 Rip-current channel with simple fining-upward fill within planar laminated sandstones of the foreshore facies. St - Cross stratified sandstone. Note how the upper part of the channel is progressively filled by thinly bedded sets of wave ripple lamination (Wr). Notebook 30cm length.

FIG 4.62 (i-ii) Photograph and interpretative sketch (i) and graphic log (ii) log of offshore to inshore tidal channel facies sequence, Sourribes section, Bm2 member.



Detailed sketch of 'Inclined Cross-Bedding.'



Channelised sandstone facies
Channelised fining upward sequence of bidirectional trough cross-stratified sandstone. Generation of inclined cross-bedding.

Inshore tidal channel.

Planar sandstone facies
Planar laminated sandstone with dm-scale beds of hummocky cross-stratification. Wave ripple lamination commonly reworks tops of beds.

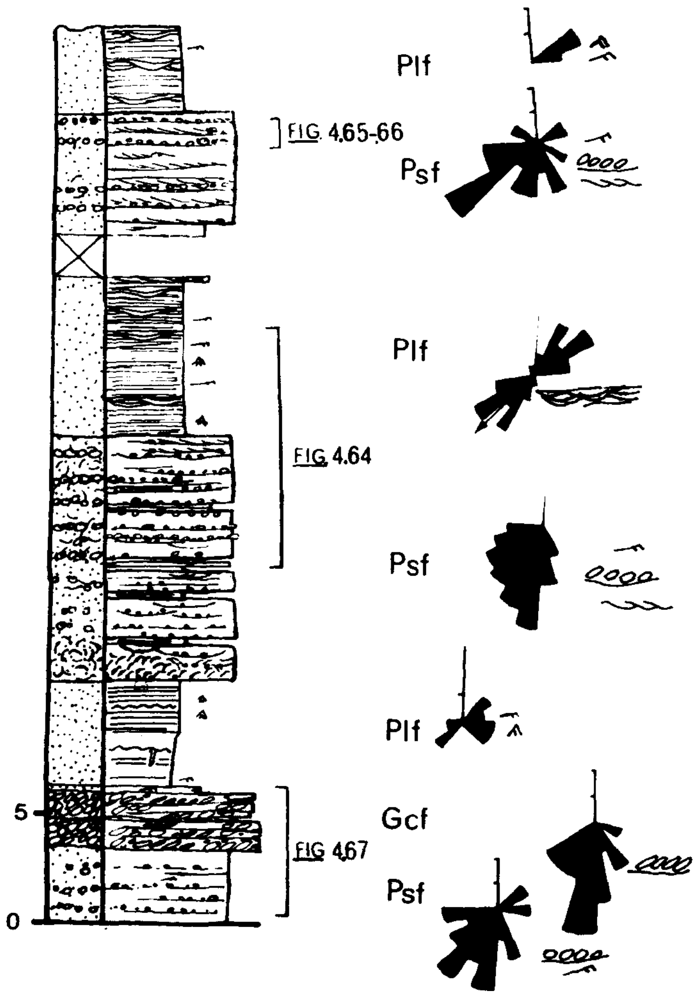
Sheet sandstone facies
Alternation of dm-scale graded sandstone beds and muddy sandstone or wave ripple laminated heterolithic bedding. Sandstone shows planar/low angle lamination, and current ripple lamination.

Transition zone, alternation of storm and fairweather wave deposition.

Muddy sandstone facies
Intensively bioturbated muddy sandstone with open marine bivalve and echinoid shell layers.

Offshore above storm wave base.

FIG 4.62 (ii) Interpretative log of offshore to inshore tidal channel sequence illustrated in Fig 4.62 (i)



Plf
FIG. 4.65-66
Psf
Plf
FIG. 4.64
Psf
Plf
5
Gcf
FIG. 4.67
Psf
0

DESCRIPTION	INTERPRETATION
Planar sandstone facies Plf Metre scale units of well sorted fine grained sandstones dominantly planar laminated with swaly 'gutter casts', and graded beds with hummocky cross-stratification, climbing ripple lamination, and wave ripple lamination. Assymetrical wave ripple lamination directed toward the E-NE; h.c.s and climbing ripple sets directed toward the W-S.	Foreshore-Upper Shoreface.
Pebbly sandstone facies Psf Metre scale units of coarse pebbly sandstone. Dm-scale beds of graded, cobble-pebble conglomerate interbedded with trough cross-stratified pebbly sandstone. Conglomerate imbrication and stratification gives SW palaeoflow. Within facies thin beds (cm-dm scale) of planar sandstone facies, commonly overlying planar erosion surface with cobble lag.	Distributary mouth bar Flashy fluvial discharge into shallow marine environment, bedload sheetflows and 3-D dunes. Periodic storm reworking
Channelised gravel facies Gcf Channelised unit of imbricate cobble conglomerate.	Fluvio-distributary channel with longitudinal conglomerate bars.

FIG 4.63 Shoreface-Foreshore - Fluvio-Distributary facies association. Esclangon section, Bm1 member

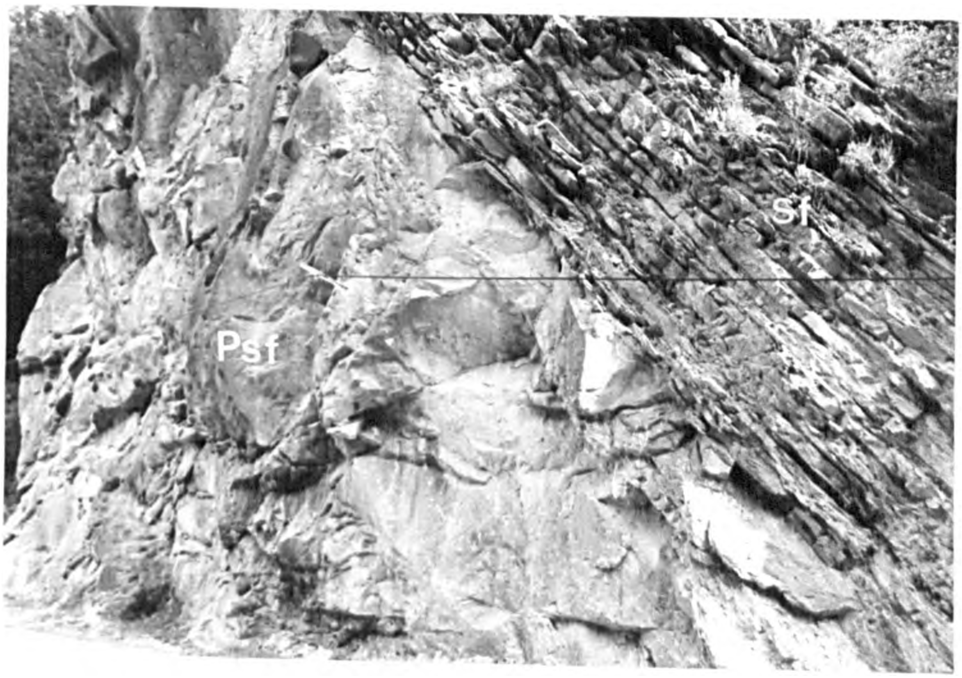
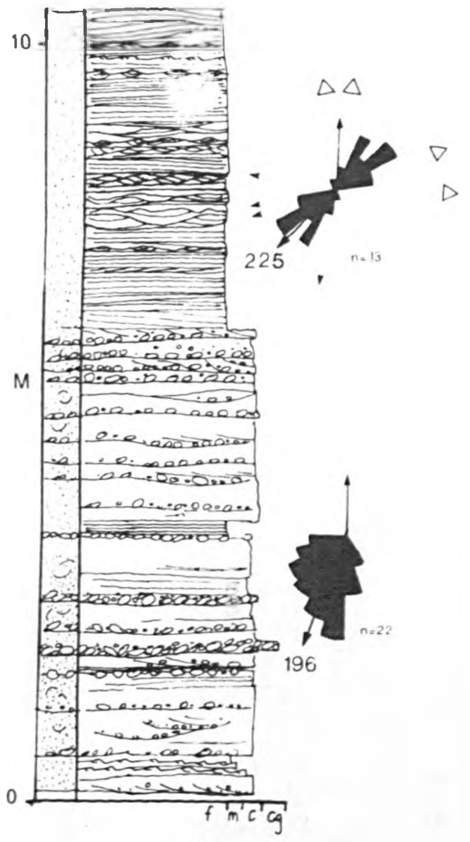


FIG 4.64 (i-ii) Photograph (i) and interpretative graphic log (ii). Coarse sandstones and graded conglomerates of the pebbly sandstone facies (Psf) abruptly overlain by an association of well sorted, fine sandstones showing planar lamination, graded sandstone beds with h.c.s, and thin wave ripple lamination of the planar sandstone facies (Plf)



Planar sandstone facies

Fine grained sandstones of the planar sandstone facies sharply overlie the Psf facies. Planar to low angle laminated sandstone dominant with intervals of amalgamated h.c.s. Also graded beds with climbing ripple lamination. Assymetrical wave ripples reworking tops of beds are directed towards the NE.

Shoreface-foreshore.

Planar and wave ripple lamination developed under shoaling fair weather wave action, periodic storm action scours substrate generating h.c.s (combine flows) and 'gutter cast' scours. H.C.S and climbing ripple sets directed offshore ~parallel to wave approach.

Graded conglomerate facies

Graded, cobble-pebble grade conglomerate (Gg) with cross-stratified (St) or ripple laminated (Sr) sandstone top. Interbedded trough cross-stratified pebbly sandstone. Open-restricted marine bivalve shells. Conglomerate beds palaeo-flow offshore, sandstone stratification and lamination rarely on-shore directed

Distributary mouth-bar: Gravel beds deposited from fluvial generated, turbulent flows - bedload traction deposits. Marine bioclasts erosionally incorporated. Low discharge sandstone deposited by 3-D dunes. Rare bidirectionality - tidal (?).

Interbedded, on dm-m scale are well-sorted planar-, swaly- laminated sandstones with thin conglomerate/bivalve shell lags.

Storm wave reworking, waves winnow substrate concentrating conglomerate and shell material; as storm wanes fine sand deposited from suspension under wave/combined flow currents.

FIG 4.64(ii) Interpretative log of Fig 4.64(i).

FIG 4.65 Detailed view of pebbly sandstone facies comprising stacked, dm scale graded, cobble-granule conglomerate beds - each white arrows marks a fining upward bed. Clast imbrication and associated cross-stratified sandstone gives flow direction (black arrows). Lens cap 5cm

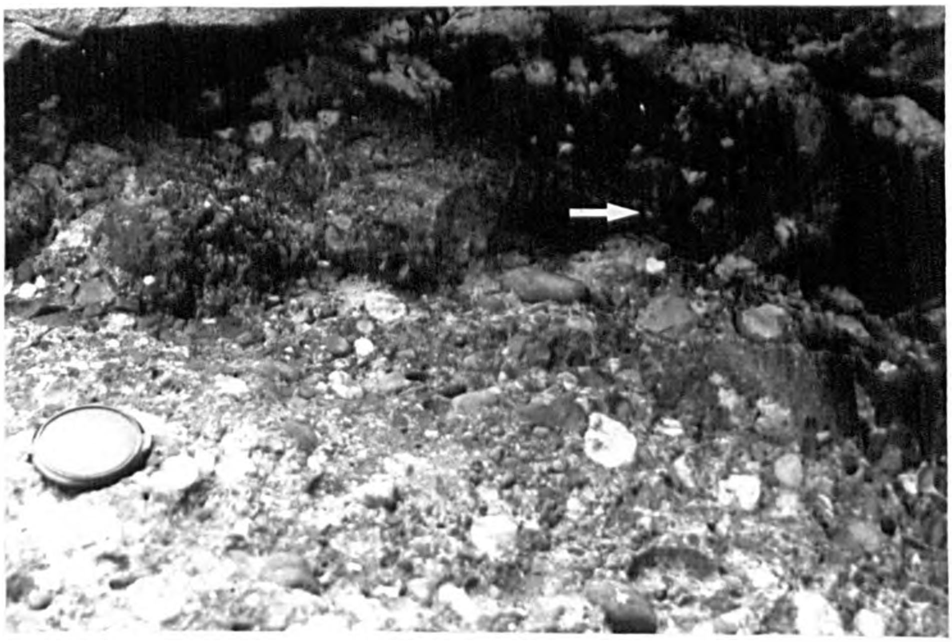
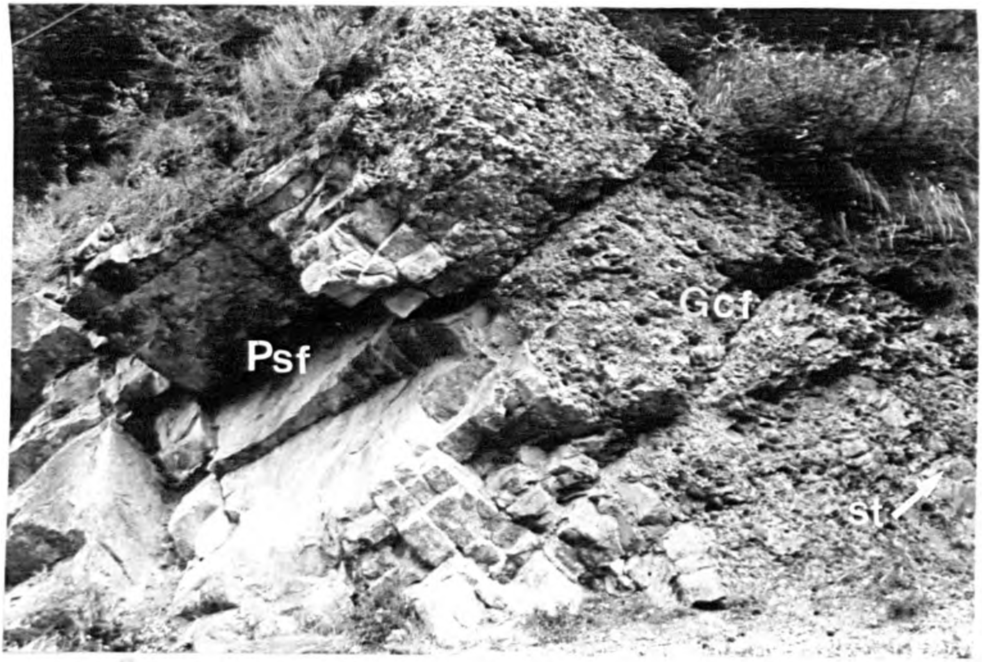
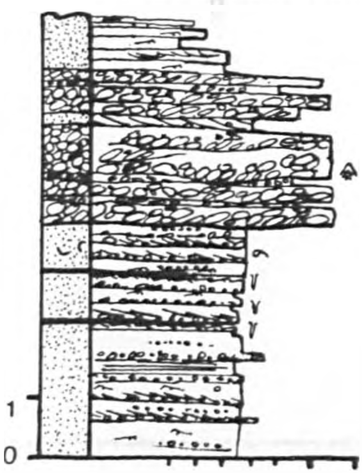


FIG 4.66 Close-up of graded conglomerate bed showing developement of cross-stratification in upper part. Lens cap 5cm.





Channelised conglomerate facies

Channelised fining-upward sequence of imbricated cobble-pebble conglomerate (Gm). Beds of dm-scale, weakly graded. Interbedded trough-cross stratified (St) coarse sandstone. Oyster shell bioclasts.

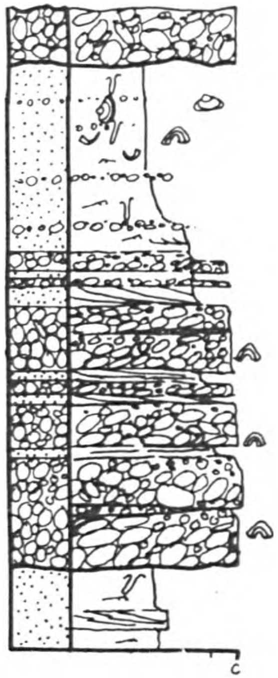
Pebbly sandstone facies

Dm scale, graded beds of pebble conglomerate with cross-stratified or ripple laminated top. Planar laminated, well sorted sandstone beds subordinately developed.

FIG 4.67 Coarsening upward sequence with coarse sandstones and graded conglomerates of the pebbly sandstone facies (Psf) erosionally overlain by a channelised unit of imbricate conglomerate of the channelised conglomerate facies (Gcf). Note the developement of erosive lenses of trough cross-stratified sandstone (st) between the conglomerate beds towards the top of the channel.

Distributary channel filled by longitudinal gravel bar deposits migrated during high discharge; 3-D dunes and ripples developed during low discharge.

Graded bedload sheetflows interpreted to be deposited from fluvial currents issueing onto distributary mouth bar. Flashy discharge Planar laminated sandstone developed during wave reworking.



DESCRIPTION

INTERPRETATION

Brackish muddy sandstone facies
Heavily bioturbated muddy sandstone with ripple lamination and thin, graded pebble-granule conglomerates.

Low energy, brackish marine environment. marginal to channels

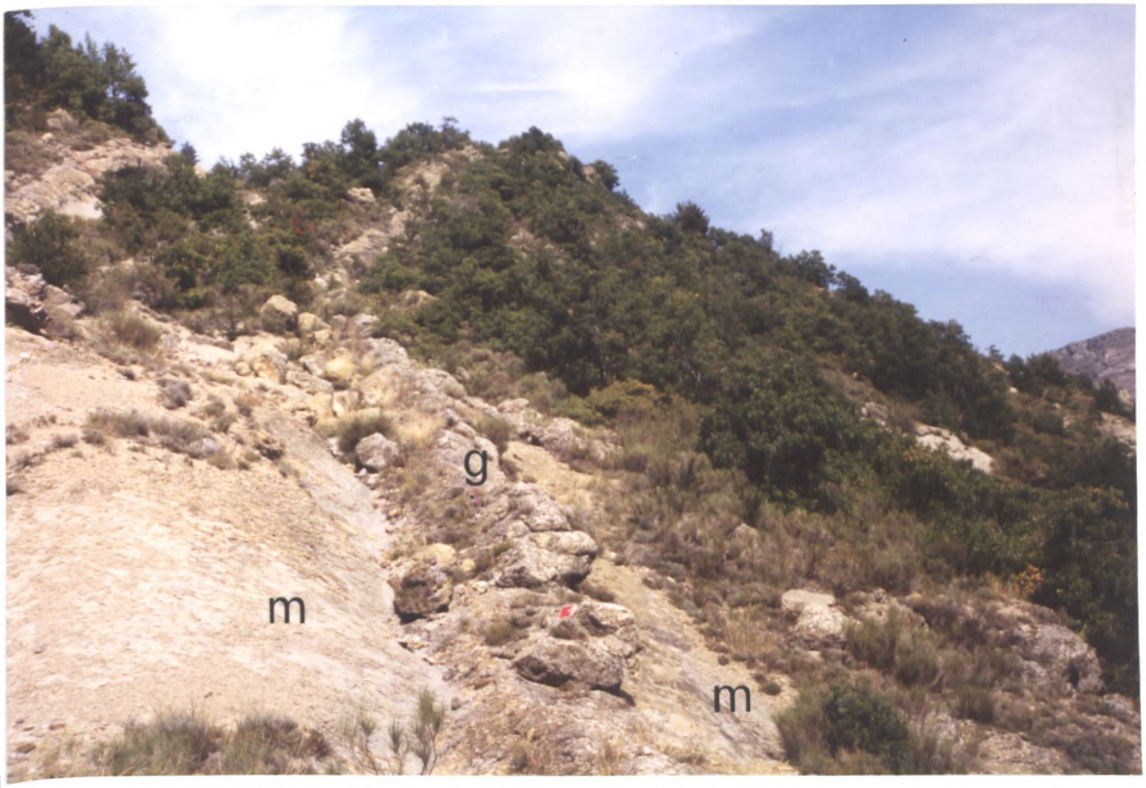
Channelised conglomerate facies
Erosive based, fining-upward unit. Internally tabular beds of cobble grade, clast supported, imbricate conglomerate (Gm). Thinly interbedded pebble conglomerates, and coarse sandstone, trough stratified (St), or, planar laminated (Sh). Oyster shells in conglomerate.

Shallow, fluvio-distributary channel fill. High flow stage longitudinal bar conglomerates. Low flow stage reworking of bars gives pebble conglomerates, and 3-D dune deposits. Oysters erosionally incorporated.

FIG 4.68 Characteristic fining-upward sequence from channelised conglomerate, to, brackish bay sandstones, within the Fluvio-distributary facies association.

FIG 4.69 Tabular conglomerate units of the channelised gravel facies (g) interbedded within deposits of the muddy sandstone facies (recessed) extend for about a hundred metres across the outcrop.

i



ii

FIG 4.70 (i-ii) Detailed views of a channelised conglomerate unit (g) of the fluvio-distributary facies interbedded within brackish muddy sandstones (m) (see Fig 4.71 for position and context). Note how laterally, the conglomerate progressively thins as it passes out into the muddy sandstone (ii).

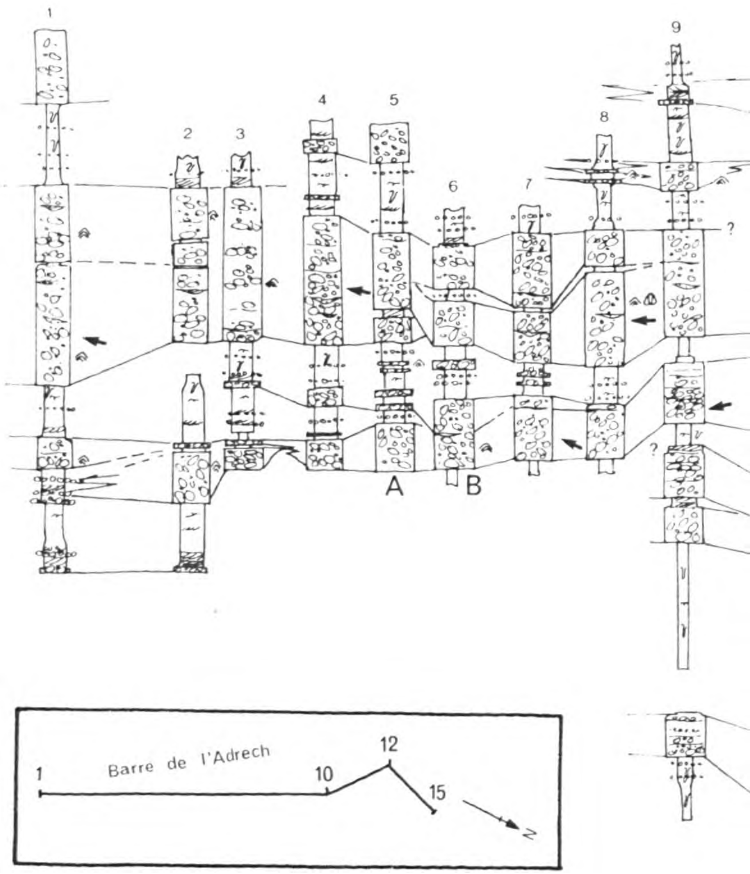
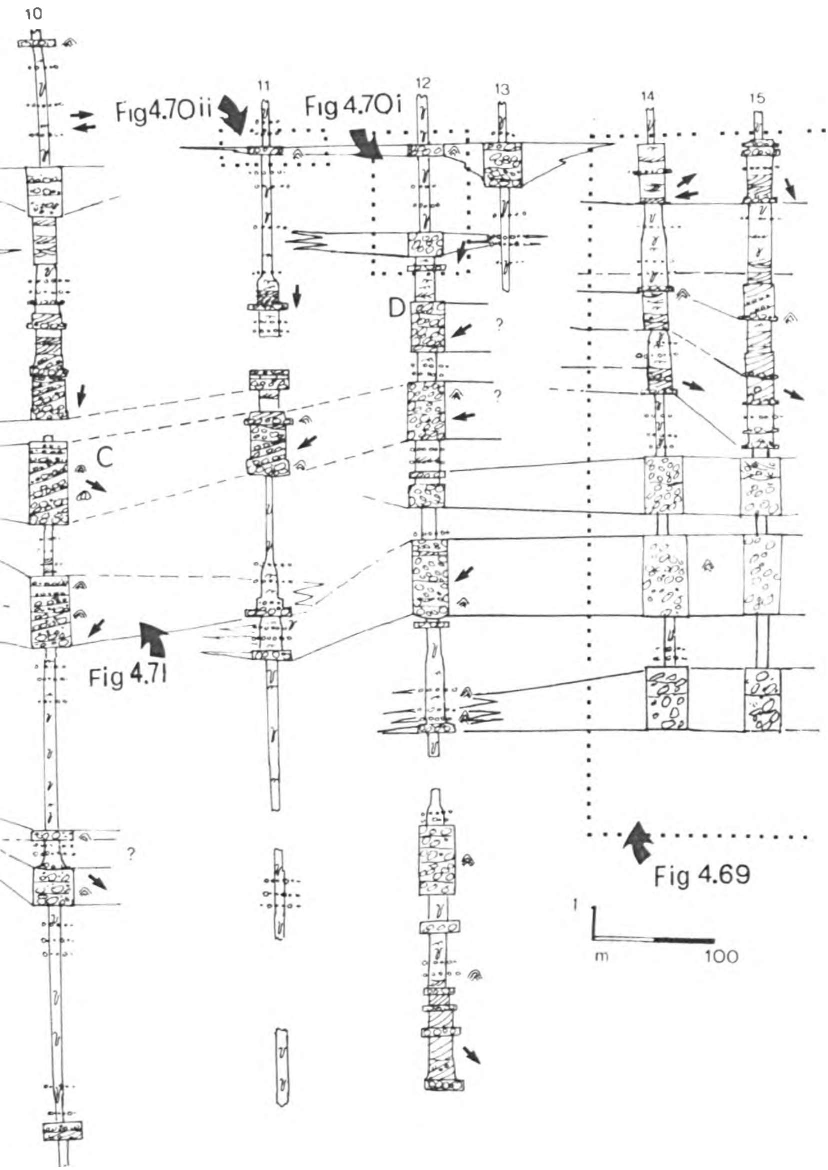


FIG 4.71 Fence diagram through the fluvio- distributary facies associaton at Barre de l'Adrech, Esclangon.
Note (i) the large scale coarsening-upward sequence of the association (mega-sequence), best displayed in section 10. (ii) repeated m-scale fining upward sequences from channelised conglomerate to brackish bay mudstone (iii) lateral passage of conglomerate filled channels into bioturbated, brackish, muddy sandstones. (3) tabular form of conglomerate units is partially reflection of sub-parallel to flow, exposure.
Note also scale exaggeration.



10
11
12
13
14
15
Fig 4.70 ii
Fig 4.70 i
D
C
?
Fig 4.71
Fig 4.69
1
m
100

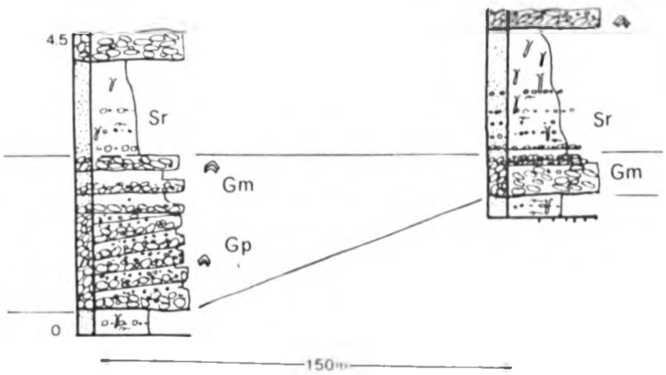


FIG 4.72 Graphic log detailing the lateral thickness and grain size variation, and shallow erosional relief on a unit of the channelised conglomerate facies (see Fig 4.71 for position)

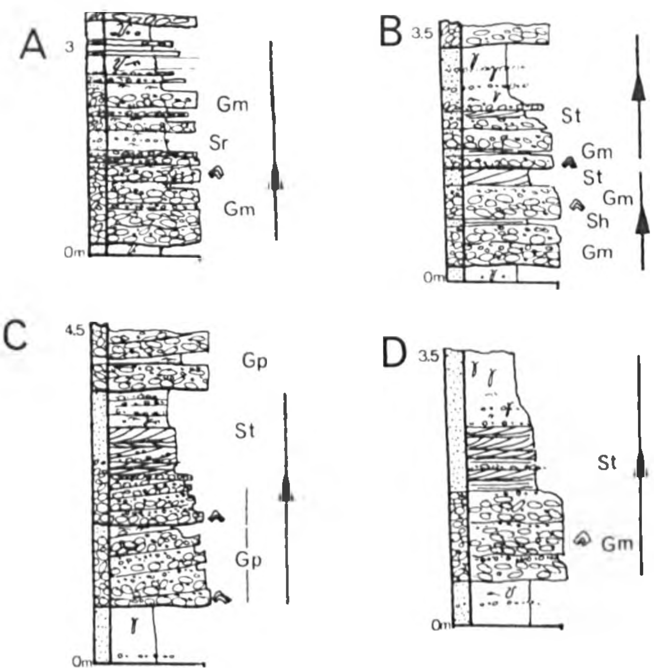
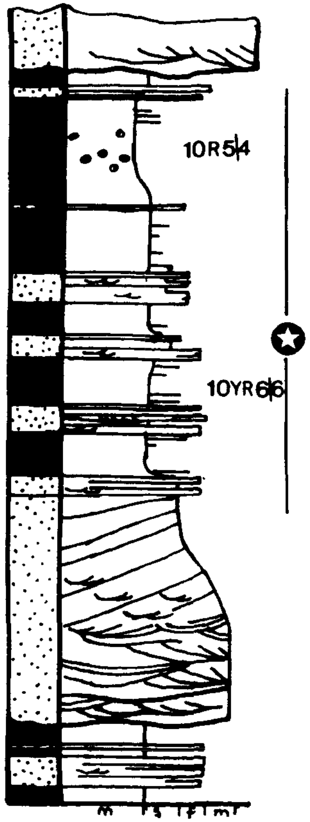


FIG 4.73 Graphic logs detailing the variation in vertical fining-upward sequences within channelised conglomerate units of the Fluvio-distributary Facies Association. See Fig 4.70 for position. Gm: Horizontally stratified conglomerate, commonly imbricate. Gp: Planar stratified conglomerate. St:Trough cross-stratified sandstone Sg: thin graded pebbly sandstone beds. Sr:Current ripple lamination

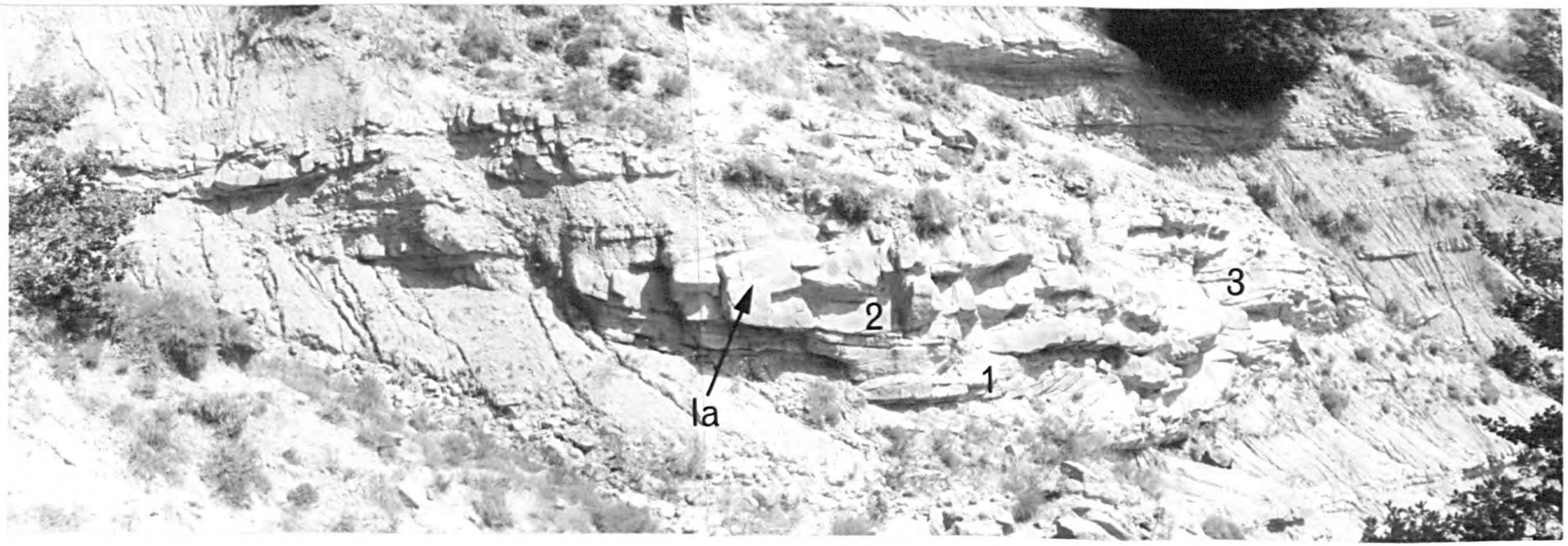
FIG 4.74 Close up view of imbricate conglomerate of channelised gravel facies. Lens cap 5cm

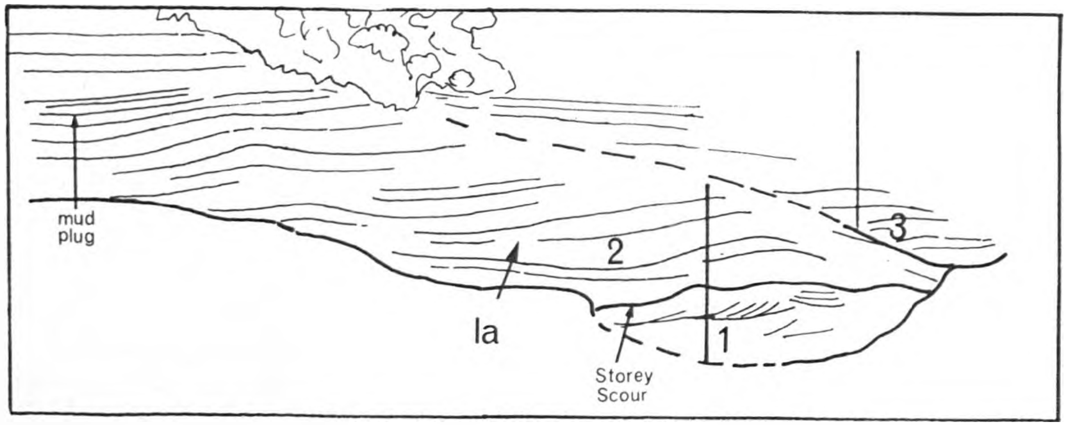
FIG 4.75 Whole, articulated oyster (6 cm length) within imbricate conglomerate of the channelised gravel facies.



DESCRIPTION	INTERPRETATION
Gley mottled mudstone, carbonate glaebules.	Higher, better drained area of floodplain. Low sedimentation rate allows calcrete formation
<u>Sheet sandstone and mudstone facies</u> Graded sandstone units with small ribbon sandstone channels. Pseudo-gley mottled mudstone	Overbank deposits in inter-channel area with impeded drainage.
<u>Channelised fluvial sandstone facies</u> Multistorey channel fill with lateral accretion bedding. Unidirectional stratification.	Mixed bedload channel, sinous.

FIG 4.76 Fining-upward sequence of the Alluvial Plain facies association.





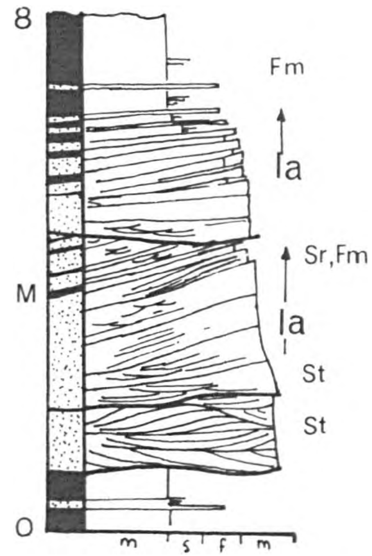
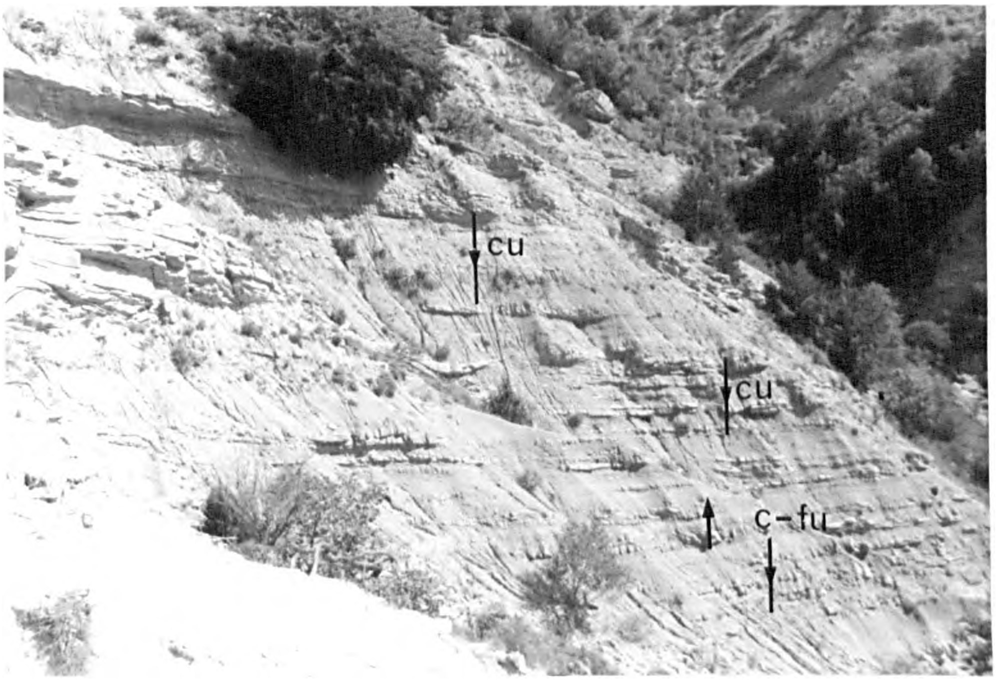


FIG 4.77 Multi-storey ribbon channel (type 2c) of the channelised sandstone facies, alluvial plain facies asociation. Note the developement of three channel storeys (1-3) with lateral accretion bedding (la), and the fine-grained deposits which 'plug' the channel and record its progressive abandonement.



4.78 Tabular sandstone and mudstone facies developed between the channelised sandstone bodies. The sandstones form m-scale assymetrical coarsening-upward packages (cu), or symmetrical packages (c-fu) with a basal coarsening upward section succeeded by a progressive fining-upward into mudstone. Note small ribbon channel sandstone (arrowed) shown in detail in Fig 4.79. Section is 40m thick.

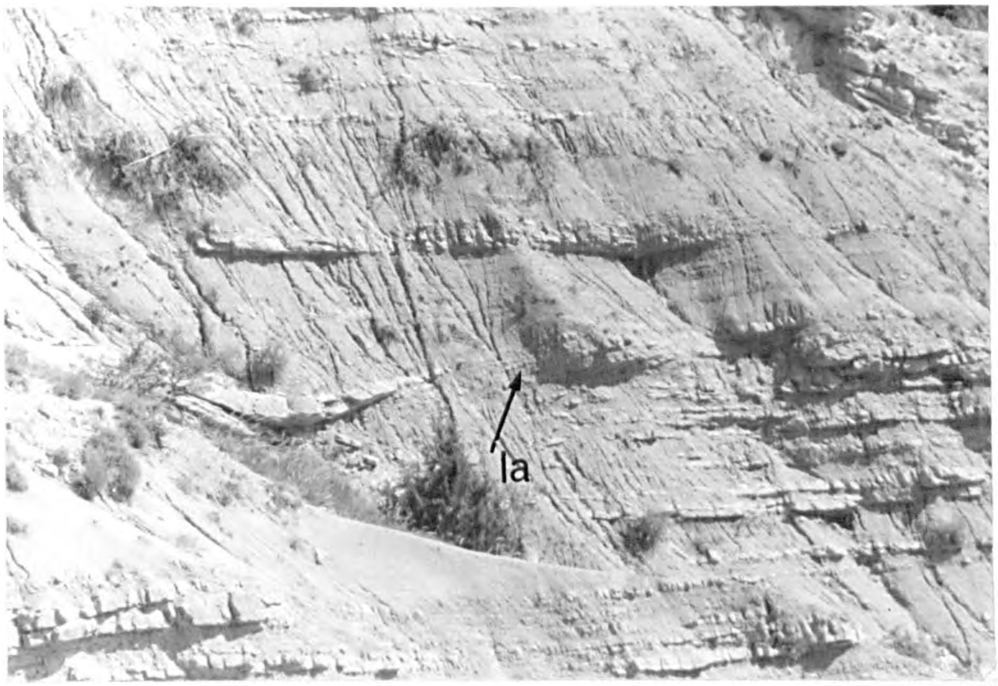
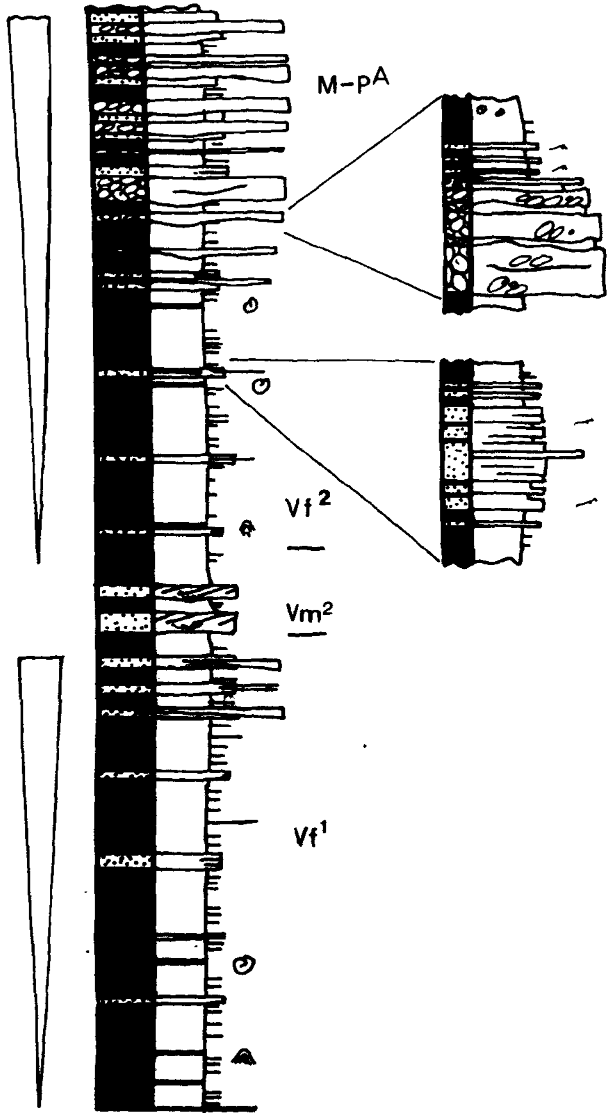


FIG 4.79 Close up view of small ribbon channel developed in tabular sandstone and mudstone facies. Note the heterolithic lateral accretion bedding (la) and mudstone plug.



Channelised gravel facies
Single-/multi- storey ribbon gravel bodies. Imbricate conglomerate fill. Channels up to 3m thick.

Alluvial fan- Mid-fan. Braided distributary channels, longitudinal bar deposits.

Sheet sandstone and mudstone facies
Alternation of heterolithic units of graded sheet sandstones with small ribbon conglomerate scours, and of heterolithic mudstones. Mudstones may be organic rich with fresh-water gastropods/oysters. Small scale soft-sediment deformation structures common.

Terminal sandstone lobes, ephemeral feeder channels, interchannel mudstones -lacustrine, lagoonal at

FIG 4.80 Large scale coarsening-upward sequences of the Alluvial fan facies association (members Vf1 - Vf2/M-P^A) form a coarsening-upward progradational mega-sequence interrupted by a marine unit (member Vm2) comprising upper estuarine channel and tidal flat facies (see Chapter 5).
The inserts show details of the ribbon channel conglomerate and sheet sandstone-mudstone facies.

FIG 4.81 Sheet sandstone and mudstone facies; repeated coarsening upward sequences passing from a mudstone dominated heterolithic interval into a tabular unit of graded sheet sandstones and thinly interbedded mudstone. Note the small, single storey conglomerate ribbon channels (arrowed). Channel in foreground is 60cm thick.

FIG 4.82 Close up of sandstone sheets showing their erosive based nature and well developed grading.

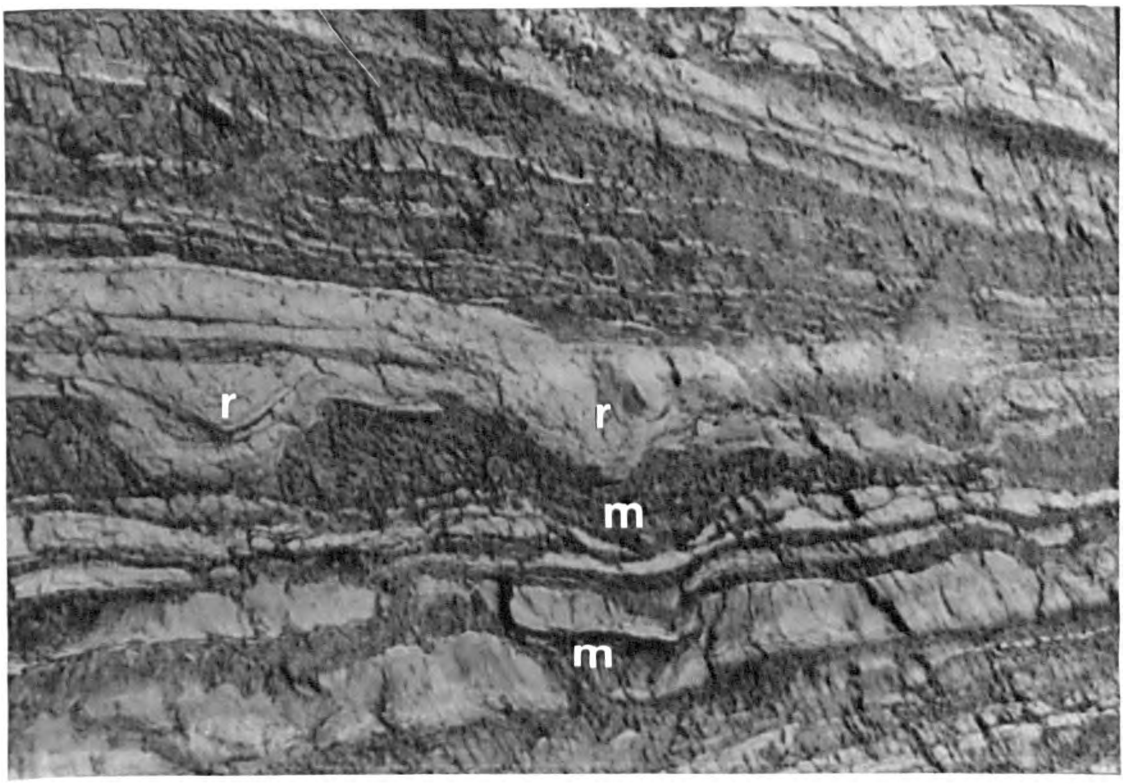


FIG 4.83 Sandstone sheets show the developement of small (cm-dm scale) ribbon shaped scours (r) with a simple fining upward sandstone fill. Note the equally common developement of mud-filled scours (m) and the sharp erosive base (arrowed) to these mudstone layers outside the ribbon scour.

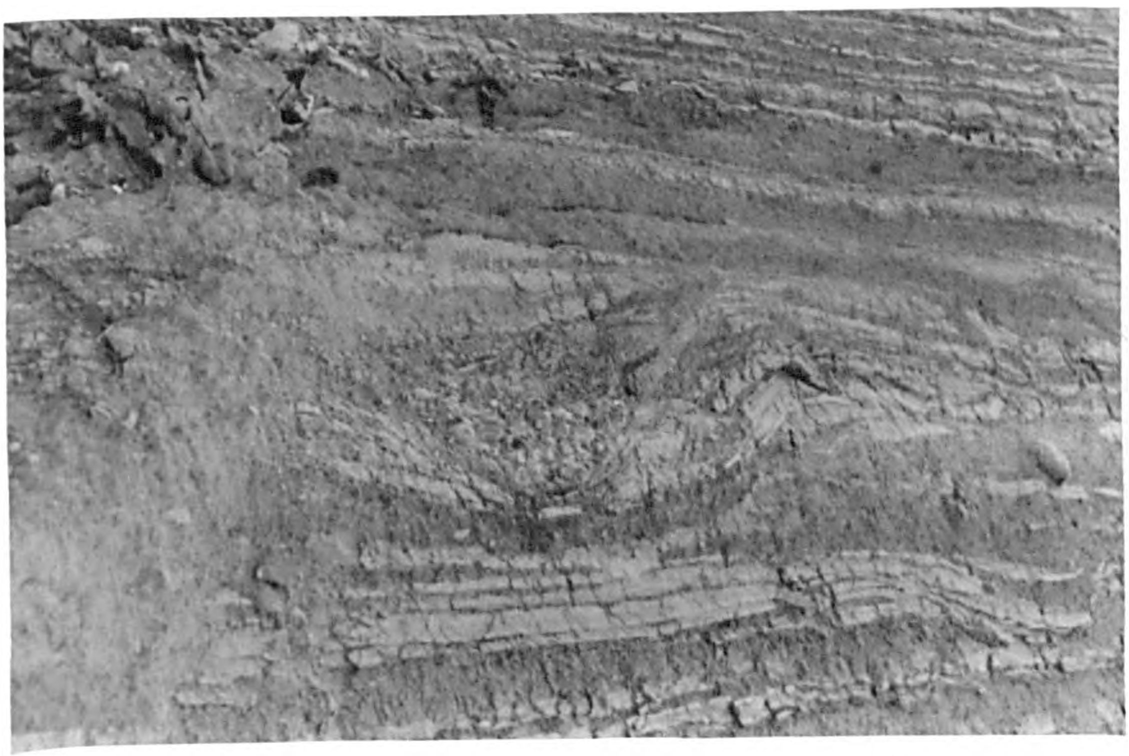


FIG 4.84 Small conglomerate filled ribbon scour within a unit of sheet sandstones.

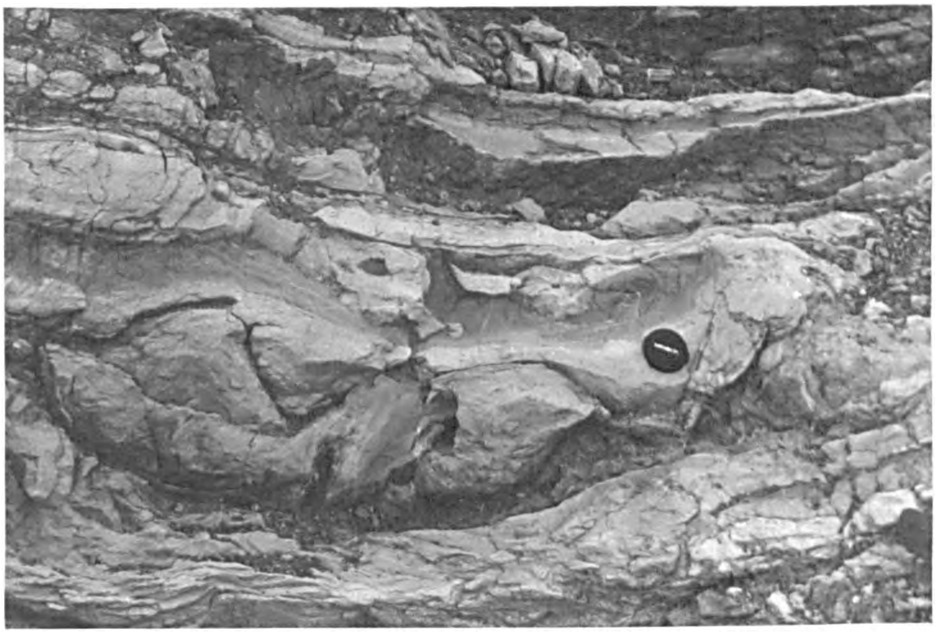


FIG 4.85 Syn-sedimentary soft-sediment deformation within the sheet sandstone and mudstone facies. Sandstone pillow enclosed within mudstone in centre of photograpgh is overlain by sandstone bed with convolute lamination. Lens cap 5cm diameter.

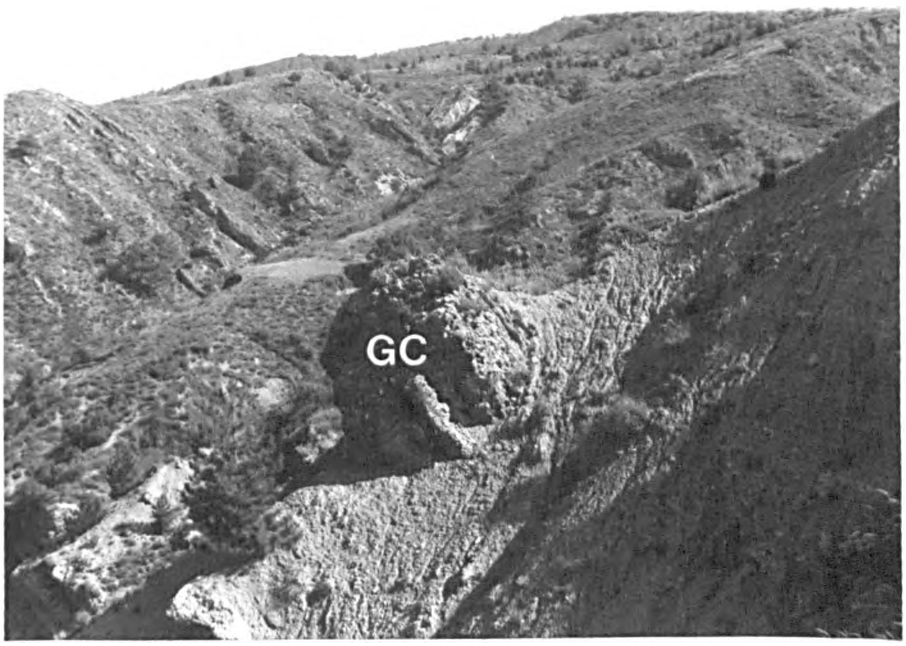
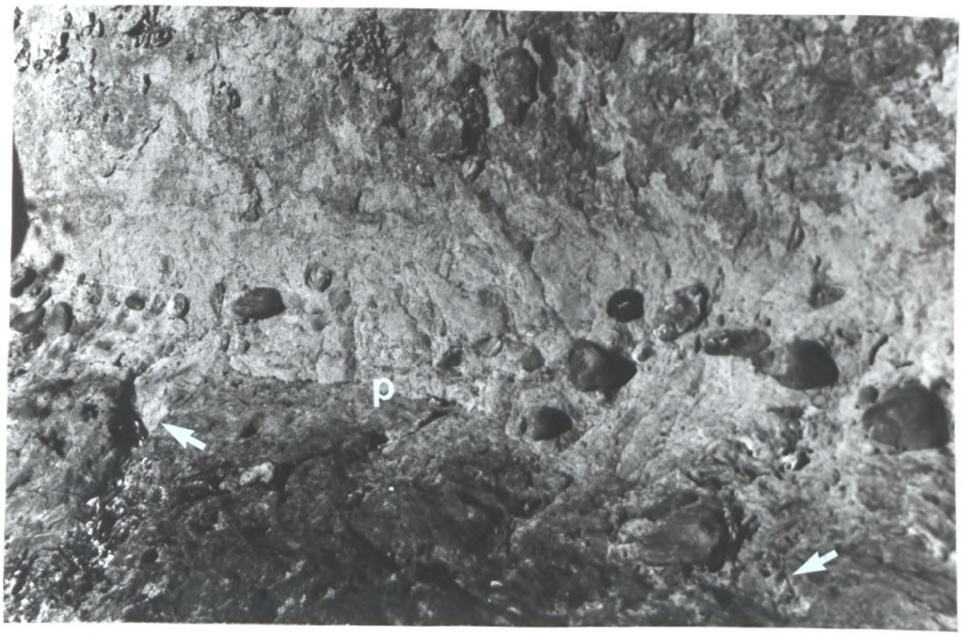


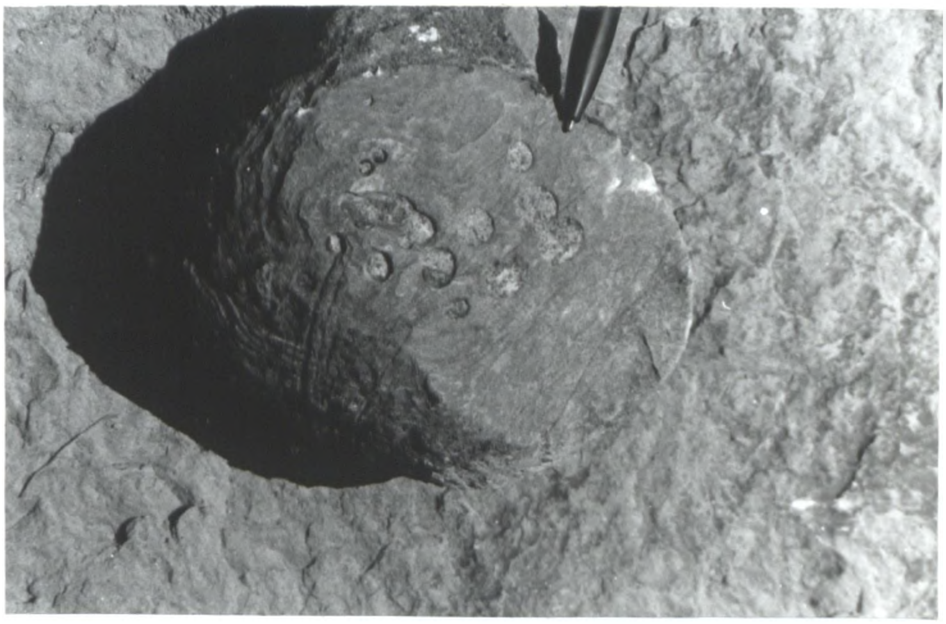
FIG 4.86 Conglomerate ribbon channel body in foreground (Gc) interbedded with mudstone and sheet sandstone facies. Channel is 2.4 m thick.

FIG 4.87 Shoreface conglomerate facies; concentration of cobble-pebble grade conglomerate clasts above a planar erosion surface (p). Note that the clasts are well rounded and sorted and are enclosed within fine grained deposits of the muddy sandstone facies. Note also local developement of irregular erosional relief (arrowed) filled by overlying sandstone. Esclangon section, conglomerate shoreface lag at base of Vm1 member.

FIG 4.88 Pholad borings in limestone cobble of the shoreface conglomerate bed detailed in Fig 4.87.

FIG 4.89 Intensively bored (pholad borings) limestone boulder from the shoreface conglomerate horizon at the base of the Bm1 member, Sourribes.


p




Vivitar

FIG 4.90 Tabular unit of heavily bioturbated glauconitic sandstone within offshore muddy sandstone facies. Rucksack for scale.

FIG 4.91 Close-up view of the basal part of a glauconitic sandstone unit showing its homogenous form and lack of physical sedimentary structures due to intensive bioturbation. Lens cap 5cm diameter.

CHAPTER 5

THE MARINE MOLASSE FORMATION - VERTICAL AND LATERAL DISTRIBUTION OF FACIES ASSOCIATIONS.

In this chapter the vertical and lateral distribution of the facies associations of the Marine Molasse formation (as detailed in Chapter 4) are discussed. The associations are interpreted to form distinct members (Bm1 - Vf2) (Fig 1.11) which define four major transgressive-progradational cycles of 100-300m thickness in the Digne - Valensole basin. These cycles are superimposed upon a general transgressive-progradational Miocene cycle. Transgression was initiated in the Burdigalian with a regional 'eustatic' (see 5.5 for discussion) sea level rise which resulted in the establishment of shallow marine seas in the Durancian basins. Progradation was dominant during the late Burd (galian and Vindobonian periods and culminated in the Messinian with the establishment of an alpine fluvial system (the Valensole Formation) across the Durancian basin.

The description and interpretation of the members is preceeded by a brief review of transgressional and progradational mechanisms.

5.2 Transgressional and Progradational Sequences - Controlling Factors in their Development.

This section briefly addresses the problem of the variable factors involved in the generation of transgressive and progradational sequences, and serves as an introduction to the sequences discussed in the succeeding section.

5.2.1 Transgressions.

Two contrasting mechanisms of transgression have been suggested for mixed tide-wave, and wave dominated, shorelines.

The most widely accepted is **shoreface retreat** in which, as sea level rises, the barrier migrates continuously landward by shoreface erosion and / or tidal inlet erosion (Fig. 5.1) (Schwartz 1977, Swift 1968, Swift *et al.*, 1971).

The alternative mechanism is that of **in-place drowning**

(Fig. 5.1) in which the barrier remains in place until the breaker zone reaches the top of the barrier. At this point the breaker zone jumps landward, drowning the barrier.

These two different styles of transgression may be related to the rates of relative sea-level rise and sediment supply. Rapid rates of submergence favour rapid retreat of the shoreface by in-place drowning, generating thick back-barrier sequences (Kraft 1971, Swift 1975, Rampino & Sanders 1981, & 1982). By contrast, shoreface erosional retreat results from slow rates of submergence, and low sediment supply, and is characterised by thin transgressive horizons (Swift 1975, Rampino & Sanders 1982). Where sediment supply fluctuates, **discontinuous shoreline retreat** occurs, with shoreface erosion punctuated by periods of barrier up-building or progradation.

These contrasting mechanisms of transgression generate characteristic sequences, which may be different for wave dominated, and mixed, tide-wave dominated shorelines (Rampino & Sanders 1982). These sequences are detailed in Fig. 5.1.

5.2.2 Progradations.

Progradation occurs where the relative rate of sediment supply is greater than that of base level rise. The maintenance of relatively high rates of sediment supply and low rates of submergence results in progradation and the generation of **continuous progradational sequences.** In contrast, **discontinous progradational sequences** are generated where fluctuations in sediment supply or submergence lead to step-wise progradation.

Models of progradational sequences are varied and dependent upon the shoreline system involved (see Elliott 1986 for review).

5.3 Spatial Variation of Facies Sequences.

5.3.1 Introduction.

Correlation of the Marine Molasse successions in the study area is difficult. Exposure is excellent in a number of key localities (Fig. 4.4), generally along river valleys and road sections, but is poor elsewhere, and the overthrusting of sections of the basin between key localities did not allow physical correlation. The deposits of the study area outcrop in two geographically distinct regions and are interepreted to represent the remnants of the fills of two discrete sub-basins, the Digne-Valensole and Jabron basins (Fig. 4.4).

The bio-stratigraphy of the formation, is detailed in Fig. 1.11, and takes the form of a simple division into a Burdigalian (21-15.5ma), and Vindobonian (15-6.5ma) succession.

Stratigraphic control is extended in the Digne-Valensole basin by the recognition of basin-wide, transgressive events which define three basin-wide transgressive - progradational cycles, Bm1, Bm2 and Vm1-Vf2., which are of 100-300m thickness.

A fourth transgressive - progradational cycle is restricted in its development to the north of the basin .

The base of each cycle is typically marked by a thin, dm-m scale, transgressive facies or facies sequence, but transgressional mega-sequences of up to 100m thickness are developed at the base of the member. The upper progradational part of the member is typically much thicker and comprises megasequences of the order of 100- ~300m thickness, which internally comprise smaller scale (tens of metres) facies sequences.

There are drawbacks to this lithostratigraphic method of correlation, notably the problem of diachronism, but given the limited biostratigraphy of the formation this is not definable.

To avoid repetition, the description and interpretation of the logged sections will be considered in a stratigraphic context only, and will take the form of an analysis of the vertical organisation of facies associations within members.

The members will be considered in terms of:

(1) the Digne-Valensole basin.

(2) the Jabron basin.

5.3.2 The Digne-Valensole basin.

The Marine Molasse is well exposed in a series of localities (Fig 4.4) along the western and northern margins of the Digne-Valensole basin (Fig. 4.4). Three major transgressional-progradational cycles represented by the Bm1, Bm2, and Vm1/2-Vf2 members are recognised across the basin, with a fourth cycle, developed in the north of the basin, defined by the localised progradation of fluvial systems (member Vf1) in response to the uplift of the basin margin during the Vindobonian (Fig. 5.2).

5.3.2.1 **Member Bm1.**

This basal member of the formation (Fig 5.2 - 5.3) lies conformably, or locally disconformably, upon upper Oligocene/ Aquitanian continental deposits of the Molasse Rouge formation.

The base of the member is defined by an thin transgressive facies or by the gradational passage from distal alluvial fan to nearshore marine deposits. The top of the member is marked by a basin-wide transgressive facies. The member has a variable thickness, ranging from 120m at Sourribes to 200m at Esclangon as detailed on Fig. 5.2 .

The member can be considered in terms of two sub-members. The lower sub-member (Bm1a), typically displays a vertical change from inner estuarine to offshore facies and is interpreted to reflect a transgressive phase of 40-140m thickness. The upper sub-member (Bm1b) marks a return of inshore conditions and is interpreted as a progradational succession of 40-80m thickness.

Bm1a - Transgressive sub-member

This transgressive sub-member is well exposed at **St.Symphorien** along the Ravine de Symphorien and at Mens (Fig 4.40) where it comprises a 120m thick, transgressional mega-sequence (Fig 5.3). The basal 20m, comprises upper estuarine, tidal channel and flat facies sequences which gradationally overlie distal fan facies associations of the Vancon fan system. These estuarine facies pass gradationally up into a 50m thick succession of the inshore tidal channel and shoal facies association (Fig. 4.15). The transgressive trend is continued in the upper part of the sub-member with the passage, across a transgressive shoreface lag facies of bored cobbles into a 40m thick succession of muddy sandstone offshore facies, which is traceable laterally for over 2km.

At **Auribeau** a 140m thick transgressional mega-sequence is developed. The basal 40m of the sub-member is dominanted by planar and cross-laminated sandstones of the shoreface- foreshore facies. The succession fines upwards initially into heterolithic, and channelised sandstone facies of the inshore tidal shoal facies association and subsequently into a thick succession of muddy sandstones of the offshore facies.

At **Esclangon** the sub-member comprises a 100m thick transgressional mega-sequence which displays the progressive, passage from an upper estuarine tidal facies association, into wave dominated, foreshore - shoreface facies. At the base of the sub-member, repeated fining-upward, tidal channel to freshwater marsh facies sequences of the upper estuarine facies association are developed. These are gradationally succeeded by shoreface and fluvio-distributary facies association deposits (Fig. 4.63). The conglomerates of the fluvio-distributary facies are a polymict assemblage dominated by chert, quartz and micritic Mesozoic limestones and including granite (two types) and serpentinised gabbro within a quartz-rich sandstone.

The top of the mega-sequence is marked by the abrupt passage from foreshore sandstones, to offshore mudstones across a shoreface lag horizon.

At **Sourribes**, the sub-member is 90m thick and commences with a thin, transgressive shorelag facies of bored limestone boulders. This facies is overlain by a succession of muddy sandstones of the offshore facies.

Bm1b - Progradational Sub-member.

The succeeding Bm1b sub-member has a gradational base and is interpreted to record a phase of progradation across the basin (Fig 5.2 -5.3).

At **St.Symphorien** the sub-member is 60m thick and takes the form of a progradational succession which initially coarsens-upward from offshore muddy sandstones into inshore tidal shoal facies, and subsequently fines-upward into a thick unit (45m) of estuarine mudflat, or brackish bay facies.

To the east at **Auribeau**, the 65m thick succession shows a fining upward from offshore muddy sandstones to muddy siltstones, with a brackish marine fauna, including monospecific

Turretella shell layers, and small cobble lagged, tidal gully fills. This upper level is interpreted to reflect the progradation of a brackish bay, or estuarine mudflat environment.

Progradation at **Esclangon** is marked by a coarsening-upward, offshore to shoreface-foreshore facies sequence of 10m thickness, above which a 90m thick aggradational succession of massive sandstones of the shoreface-foreshore and distributary mouth bar facies association (pebbly sandstone facies) is developed (Fig 5.16).

At **Sourribes** a coarsening-upward offshore to inshore tidal shoal and channel facies sequence is initially developed. This fines into a poorly exposed heterolithic tidal facies succession. An abundance of woody and plant material in channelised sandstone facies within this succession supports an upper estuarine interpretation. The coarsening upward into these facies from the offshore facies at the top of the underlying sub-member may be interpreted in terms of shoreline progradation.

Interpretive synthesis.

Sub-member Bm1a.

The transgressive shoreface lag facies at the base of the member in the south-west of the basin at (Sourribes) is correlatable to the south, with a coarse lag unit in the Mirabeau borehole (Fig 4.4), and in the Forcalquier basin (m1bs unit of Gigot *et al.*, 1978) (see Fig. 1.9 ii). The developement of this facies is interpreted to indicate that during the Burdigalian, a shoreline transgressed northward, through the Forcalquier basin and into the Digne-Valensole basin as far north as Sourribes, by the process of erosional shoreface retreat. An elongate, NE-SW trending gulf, the Digne-Valensole gulf, was established which passed south-westward into the Apt seaway (Fig 5.14). The gulf was fringed by siliciclastic shorelines with a mixed tidal (semi-diurnal) and wave influenced regime (see Fig 5.16)

In the north of the basin, transgression was accompanied by higher rates of sediment supply and relative subsidence resulting in the developement of thick transgressive mega-sequences at

St.Symphorien, Auribeau and Esclangon.

At **St.Symphorien** the fining-upward mega-sequence from inner estuarine to offshore facies associations is interpreted to reflect the establishment of an estuarine, and subsequently inshore tidal shoreline system. The development of a shoreface lag within the transgressive mega-sequence separating the inshore from offshore facies suggests that the transgression involved a phase of erosional shoreface retreat.

Palaeocurrents from inshore tidal channel sequences are bimodal, with a NNW-S trend and with the former being dominant. The inferred direction of bedform migration within these channel facies is probably shore normal in an analogous manner to the tidal channel systems of the Dutch-German and Nigerian coasts (Oomkens & Terwindt 1960, Oomkens 1974, and Nummedal 1981) inferring an ENE-WSW trending shoreline (Fig 5.14). Wave ripple lamination from the offshore and inshore shoal facies indicate a mean wave oscillation direction of 135 - 315° oblique to the shoreline , with groove casts and ripple lamination in graded storm sandstones indicating southerly flow directions (175°N).

The absence of beach facies within the sub-member may indicate that an open estuary was developed, or, that the shoreline was subjected to erosional reworking by sub-tidal channels (Kumar & Saunders 1974).

At **Auribeau** a wave dominated shoreline (Fig 5.14) was succeeded by an inshore tidal shoreline system. Wave ripple lamination from the beach facies indicates a mean wave oscillation direction of 161-341°, with NNE directed sets interpreted as being landward directed.

To the north-east, at **Esclangon** a wave-dominated, barrier (?) shoreline apparently fronted a wave protected upper estuarine or lagoonal system (Fig 5.14 & 5.16). Fluvio-distributary channels issued directly onto the shoreface as distributary mouth-bars subjected to storm-wave reworking. The paleocurrents of imbricated conglomerate and cross-stratified sandstones of these distributary facies display a unimodal 221°N palaeoflow trend which clearly indicates that offshore conditions lay to the south, and suggest that the shoreline had a general E-W trend. Wave ripple lamination in the beach facies indicates that waves had a mean oscillation direction of 182°, with asymetrical sets predominantly directed toward the NE (landward). Groove casts on h.c.s beds, and gutter cast scours of the beach facies have a mean SW orientation.

High rates of subsidence during this transgressive phase are obvious from the 100m thickness of the transgressive mega-sequence.

At **Sourribes,** the transgression is recorded by a thin transgressive shoreface lag horizon indicating that transgression involved a slow rate of relative sea level rise and sediment supply.

Sub-member Bm1b.

This sub-member records a period of progradation of the shorelines of thr Digne-Valensole basin, with inner estuarine or lagoonal conditions established in the north and west of the basin (Sourribes, St. Symphorien, Auribeau) and a wave dominated shoreline in the north-east (Esclangon).

At **Esclangon,** palaeocurrents from cross-stratification within sandstone dominated distributary channel facies (Fig 5.3i) show a dominant SW trend (215°N), with a subordinate NE mode attributed to either wave or tidal currents. Palaeocurrent data from groove casts on hummocky cross-stratified beds of the transition zone facies, and gutter cast scours of the shoreface display a range of SW-SSE orientations with a mean SSE trend, whilst associated wave ripple lamination indicates a mean 010-190° wave oscillation direction. As in the Bm1a sub-member this data indicates that an E-W shoreline was developed at Esclangon, which passed offshore southward.

5.3.2.2 Member Bm2

This upper Burdigalian member varies in thickness from 100m at Esclangon, to 180m at Sourribes (Fig. 5.4) and forms an asymetrical, transgressive-progradational/aggradational cycle. The base of the member is marked by a transgressive shoreface lag facies in two of the areas, and by a marked facies change in the other areas. The top of the member is defined by a basin-wide transgressive facies.

The vertical arrangement of facies sequences within the member across the basin are interpreted to reflect a period of transgression which was succeeded by a period of shoreline aggradation and progradation .

At **St.Symphorien** the member is 160m thick and comprises an aggradational succesion of repeated inshore tidal channel (25%), and tidal shoal facies (25%). The base of the member is marked by a prominent erosion surface lagged with bored cobbles and pebbles, which is directly overlain by cross-stratified sandstones of the channelised tidal sandstone facies. This surface is discontinuosly exposed across 2kms within the St.Symphorien area. A 45m thick fining-upward sequence passing from inshore tidal channel, into shoal facies is truncated by a second prominent, and laterally extensive, bored cobble lagged erosion surface.

At **Auribeau** the base of the member is defined by a 7m thick unit of channelised tidal sandstone facies which erosively overlies upper estuarine (lagoonal) mudflat facies at the top of of the Bm1b sub-member.

Above this unit the succession is dominated by wavy-flaser heterolithic tidal facies (80%), and tidal, channelised sandstone facies (20%) which show the principal features of the inshore tidal channel and shoal facies association. However, the presence within the heterolithic facies of isolated examples of small tidal gully fills are characteristic of the upper estuarine facies association, suggesting that an inshore setting transitional between these systems was established.

In the north-east of the basin at **Esclangon**, the base of the member is marked by a cobble lagged shoreface erosion surface above which a 20m thick succesion of the transgressive glauconitic sandstone facies is developed. This distinctive transgressive unit can be traced for some 2km around the 'Villodrome' syncline (Fig. 5.8).

The upper 60m of the member, comprises a large scale offshore to estuarine tidal channel coarsening-upward (mega-) sequence. The sequence is dominated by muddy sandstones of the offshore facies within which repeated dm-m scale units of distal, tidal shoal sandstone are developed. The top of the sequence is marked by the develop ment of a multi-storey channel complex of the inshore tidal channel facies which erosively cuts down into the muddy sandstone facies. This is exposed as a prominent sandstone ridge which is laterally traceable for up to 1.6 km along the Barres de l'Adrech (Fig. 5.8).

The member is some 180m thick at **Sourribes** (Fig 5.4) where the exposed section is dominated by a progradational succession passing from a lower section dominated by offshore-inshore tidal facies sequences into an upper section dominated by upper estuarine facies association sequences. The contact with the underlying member is not exposed, but the development of offshore muddy sandstone facies at the base of the member is interpreted to record an initial transgression.

The lower section of the progradational succession comprises five repeated, (8-15m thick) coarsening-upward, offshore to inshore tidal channel sequences (Fig. 5.7), with two showing the vertical passage from offshore, through shoreface into tidal channel facies, and the remaining three inshore tidal shoal to channel facies sequences. The top of each of these repeated coarsening-upward sequences shows a thin and gradational fining-upward sequence back into the offshore facies, with the exception of the last offshore to tidal channel sequence which fines upward into upper estuarine facies. The upper section of the member comprises repeated fining upward, tidal channel and flat facies sequences of the upper estuarine facies association.

Interpretative synthesis.

The developement of basin wide, transgressive facies sequences at the base of the member records a major phase of transgression in the Digne-Valensole basin. Spatial variations in the facies sequences records the establishment of offshore conditions in the north-east and south-west of the gulf (Esclangon and Sourribes), and inshore tidal channel and shoal system in the north and north-west of the gulf (Auribeau and St.Symphorien) (see Fig 5.16).

At **St.Symphorien** the transgression was apparently achieved by tidal channel erosion during shoreface retreat (see Fig. 5.1) with the cobble lagged erosion surfaces interpreted to be the lags of tidal channels which reworked laterally adjacent barrier shoreline and estuarine/lagoonal facies.

The development of a second transgressive horizon 45m above the base of the member probably reflects that shoreline retreat was discontinuous and punctuated by a phase of progradation or aggradation. The tidal shoreline system subsequently stabilised and aggraded developing a thick succession of inshore tidal facies associations. Palaeocurrents from tidal channels (Fig. 5.4 (i) have a bipolar NNW-SSE distribution with a well developed NNW (landward) directed assymetry indicating that the shoreline maintained the WNW-ESE trend of the preceeding member. Wave ripple lamination gives a mean NW-SE wave oscillation direction, with limited storm sand beds displaying a mean SE trend (154^{0}N).

At **Auribeau** the transgression is interpreted to have resulted in the landward displacement of an upper estuarine environment and the establishment of a more seaward, inshore tidal channel and shoal system. The majority of the member records the aggradation of this tidal system indicating that a balance was achieved between the rates of relative subsidence and sediment supply.

In the west, at **Sourribes**, the basal part of the member records aggradation of a mixed wave and tide influenced, shoreline. The cyclic offshore-inshore sequences may be attributed to short-lived, progradational-transgressional shoreline cycles or, in the case of the tidal shoreline sequences to the lateral (alongshore) migration of tidal inlet and estuarine channel systems

Palaeocurrents from cross-stratification of the inshore tidal channel facies show a general bimodal NW-SE trend, with channels showing an equal predominance of northward and southward dominated fills, and suggesting that the shoreline had a general

NE-SW trend. Wave ripples of the tidal shoal facies, show a mean wave oscillation direction of 140-320°. Palaeocurrent data from groove casts and ripple lamination of storm sand sheets of the shoreface and tidal shoal facies shows a mean SSE (164°N) trend suggesting that storm currents were directed obliquely offshore toward the south.

The upper part of the member records continued progradation with the establishment of an upper estuarine environment. Palaeocurrents measured from cross-stratification within these channel sequences display a bipolar S (173°N) to NNW (349S0oN) trend, with a well developed southward directed dominance.

At **Esclangon** the development of the glauconitic sandstone transgressive facies at the base of the member indicates that transgression was achieved by erosional shoreface retreat accompanied by sediment starving of the offshore environment. Subsequent shoreline progradation was preceeded by a phase of aggradation generating a thick sequence of offshore facies with the development of tidal sandstone units indicating proximity to a tidal shoreline. Palaeocurrents from cross-stratification within these sandstone units, and from the inshore tidal channel complex at the top of the mega-sequence show a well developed bimodal trend of 032°N - 172°N, with a southward directed (offshore) as ymetry.

5.3.2.3 Vindobonian

Across the basin the Vindobonian records the vertical, progradational change over 180-450m from shallow marine (sub-member Vm1) to continental facies (sub-member Vf1-Vf2b) following a transgressive event (sub-member Vm1a).

Simple progradational successions (mega-sequences) passing from shallow marine to fluvial facies associations are developed in the north-east (Esclangon) and south-west (Sourribes) of the basin. In contrast, in the north and north-west (Auribeau and St.Symphorien) discontinuous progradational successions are developed with a 60-100m thick fluvial unit (Vf1) developed within the Vindobonian marine sequences. This fluvial unit allows the division of the Vindobonian marine molasse in the north of the basin into a lower Vm1 member, and an upper Vm2 member (Fig. 1.11). In the south where the Vf1 member is absent, the total Vindobonian marine sequences are classified as the Vm1-2 member.

5.3.2.4 **Member Vml.**

The base of this predominantly progradational member (Fig 5.2 & 5.4) is marked by the development of a transgressional unit (Vmlb sub-member) of variable thickness (dm - 5m scale).

At **St.Symphorien** the base of the 300 thick member is marked by the gradational passage over (2m) from inshore tidal shoal facies at the top of member Bm2 into offshore muddy sandstones within which a transgressive condensed limestone facies (3m thick) is developed.

This transgressive sequence is overlain by a 50m thick succession, comprising three, vertically stacked, coarsening-upward, offshore to inshore tidal channel and shoal facies sequences (an example is given in Fig. 4.31). The member coarsens upward into a 100m thick aggradational succession of repeated, inshore tidal shoal and channel facies (Fig. 4.9). The upper section of the member (160m thick) comprises repeated, fining-upward sequences, of tidal channel, tidal flat, and fresh water marsh or brackish bay facies, of the upper estuarine facies association.

The base of the member at **Auribeau** is marked by a 5m thin transgressive fining-upward sequence passing from inshore heterolithic tidal shoal facies into offshore muddy sandstone facies across a transgressive shoreface lag.

This transgressive sequence is succeeded by a progradational succession whose base is marked by a 10m thick multi-storey body of channelised sansdtone facies which erosively overlies the offshore facies. The progradational marine succession is 140m thick and dominated by upper estuarine tidal channel and flat facies sequences.

At **Esclangon** the base of the 170m thick member is defined by a transgressive shoreface lag horizon above which a coarsening-upward, offshore to fluvio-distributary mega-sequence

is developed (Figs. 5.2 & 5.8). The mega-sequence shows an initial coarsening-upward from offshore to brackish bay facies, succeeded by a continued coarsening upward into a fluvio-distribuatry facies association (Fig. 5.8) The fluvio-distributary facies association forms a distinctive unit which is traceable for some 2kms around the 'Villodrome' syncline (Fig.4.71) and within which vertically and laterally offset mouth bars and channels are individually traceable for several hundred metres (Fig. 4.71).
The conglomerate of the facies is predominantly sourced from micritic Mesozoic limestones, but with chert, quartz, and granite clasts forming a low proportion of the clasts (<10%).

At **Sourribes** a 100m break in exposure, and the absence of any fluvial facies in the exposed section means that the marine sequences above the Bm2 member cannot be differentiated into an upper and lower member (Fig 5.2). The base of the Vindobonian member (Vm1-2) at **Sourribes** comprises a 15m thick succession of muddy sandstones with an open marine fauna, interpreted as transgressive deposits of the Offshore facies.

The succeeding succession is discontinuosly exposed, with the outcrop comprising prominent sandstone ridges of 3-6m thickness separated by similar scale poorly exposed fine grained intervals. The ridges comprise sandstone bodies of the upper estuarine channelised sandstone facies seperated by brackish marine muddy sandstones of the mudflat facies. Channels, and their internal, cross-stratification show the development of thick cobble-pebble basal lags (Jurassic-Cretaceous limestone), as well as pebbly foresets indicating proximity to a fluvial source.

A ~ 100m break in the exposure at this level separates this section from the upper part of the member as described in section 5.3.2.5.

Interpretive synthesis.

A transgressive event at the base of the Vindobonian established offshore conditions along the basin's margins and resulted in the deposition of a thin transgressive facies at the base of the member. It was succeeded by a period of progradation of the gulf's shorelines which is recorded as the passage into progressively more landward facies associations.

In the north-east of the basin at **Esclangon,** the develop ment of a transgressive shoreface lag facies at the base of the member suggests that offshore conditions were achieved by shoreface erosional retreat. The succeeding coarsening-upward mega-sequence, from offshore through brackish bay into a coarse fluvio-distributary facies association records a major progradational event. Palaeocurrents from imbrication and cross-stratification within the gravel facies have a unimodal SSW orientation with a mean 198°N direction indicating that the fluvial system prograded southward and was primarily sourced from Mesozoic carbonate and (Tertiary?) clastic cover sequences of the alpine thrust belt to the the north. The progradational phase is interpreted to have been in response to increased rates of sediment supply due to alpine tectonic uplift of the source area to the north (see discussion section 5.5).

To the west at **Auribeau,** transgressive erosional shoreface retreat also established offshore conditions in the early Vindobonian. This setting was succeeded by a prograding upper estuarine tidal system. Palaeocurrent directions from cross - stratification within the estuarine tidal channels are bipolar with a N355° and S169° trend.

At **St.Symphorien** the development of a transgressive carbonate horizon at the base of the member indicates that the early Vindobonian transgression in this part of the gulf was accompanied by the starving of sediment to the offshore. This was succeeded by a prolonged period of shoreline progradation and the deposition of a progradational mega-sequence. The development of repeated, offshore to inshore tidal channel sequences at the base of the mega-sequence may reflect discontinous shoreline progradation, or that progradation was accompanied by the lateral migration of inshore tidal inlet or estuarine channels. Palaeocurrents from tidal channels of the member are bimodal with NW-SE (328°-112) trends, indicating that the shoreline had a general NE-SW trend. Wave ripple lamination from the tidal shoal facies gives a mean NW-SE wave propagation direction, perpendicular to the inferred shoreline. With continued progradtion, an upper estuarine setting was established.

The extreme thickness of this progradational marine member, requires high rates of subsidence, and sediment supply, with direct evidence for the latter coming from the overlying fluvial dominated (Vf1) member.

Finally at **Sourribes**, the member, though poorly exposed again shows the the development of an estuarine dominated progradational mega-sequence above a relatively thin transgressive sequence. The abundance of coarse cobble grade material within the estuarine channels indicates proximity to a high gradient fluvial system. This is interpreted to record the active supply of clastic material to the Digne-Valensole from the west, associated with the transpressional uplift of the Vaucluse region (Manosque-Forcalquier block) in the Vindobonian (see also section 5.3.3.3)

5.3.2.5 **Member Vf1.**

This distinctive, fluvial dominated member is interpreted to record the southward progradation of coastal fringing, alluvial systems into the north of the Digne-Valensole basin, and to form the top of a progradational mega-sequence (Vm1-Vf1) initiated at the base of the underlying Vm1 member (Figs. 5.2 & 5.4-5.5).

At **St.Symphorien**, this fluvial member is 60m thick and laterally continuous for some 2km (outcrop width). It comprises repeated fining-upward sequences (on a scale of tens of metres) of the alluvial plain facies association, with channelised sandstones (11%) enclosed within the volumetrically dominant overbank facies (~89%), of which fresh water marsh/lake mudstones form (5%). The base of the member is gradational and is marked by the interbedding of fuvial, sheet sandstone facies with upper estuarine tidal flat facies.

At **Auribeau** the 100m thick member comprises sheet sandstone and mudstone facies of the alluvial fan facies association which form a coarsening-upward mega-sequence gradationally overlying estuarine tidal flat facies (Fig 4.80). Gravel ribbon channels developed in the upper part of the member achieve only single storey scale and dm thickness. The gravel clasts of these small channels are predominantly of a distinctive glauconitic, marine sandstone which is clearly identifiable as deposits of the (Burdigalian) Marine Molasse formation.

Mapping in the Auribeau area has demonstrated that as this fluvial member is traced northward from the logged section onto the basins margin (over some 3 km), its gradational base becomes an angular unconformity (Fig. 5.6), as it erosionally overlies Burdigalian marine sequences folded in the footwall of the Melan-Clamensane thrust sheet.

The upper 10m of the member shows a gradational fining upward, sequence with mudstone becoming dominant, and subsequently being erosionally overlain by estuarine facies of the Vm2 member.

At **Esclangon**, reconnaissance work has shown, the development of a some two hundred metre thick coarsening-upward mega-sequence of ribbon channel gravel facies and yellowish-brown continental mudstones ('molasse-jaune' of Gigot *et al.*, 1974) (Figs. 5.2, 5.5 & 5.10) These are arranged in laterally offset, coarse to fine member fining-upward sequences of 5-25m scale. The single and multistorey channels are of 1-3m thickness and dominated by horizontally stratified imbricate conglomerates. The conglomerates are polymodal, sub-rounded and predominantly of upper Jurassic micritic limestone, and have a carbonate sandstone or siltstone matrix.

Synthesis1

This member records the southward progradation, during the Vindobonian, of a number of discrete and locally sourced alluvial systems along the northern margin of the Digne-Valensole gulf (Fig. 5.15) in a continuation of the progradational phase initiated in member Vm1. These fluvial systems are considered to be contemporaneous, and to record the compressional uplift (inversion) of the Melan-Clamensane fault block to the immediate north of the Digne-Valensole basin in response to the translation of alpine compressional strain into the autochthonous foreland (see section 6.6.3).

In the north-east of the basin, at **Esclangon**, the abrupt passage from an alpine sourced, fluvio-distributary system at the top of member Vm1, into a locally sourced alluvial-fan system in member Vf1 is interpreted to record the creation of tectonic relief along the basin margin and the consequent diversion of the alpine sourced system.

The alluvial fan association bears close similarities with the medial fan facies association of the continental, Vancon terminal fan system of the Molasse Rouge formation (section 3.8) and is given a similar interpretation. Palaeocurrents from channel axis and imbrication of the alluvial-fan facies have a mean SSW (216°N) trend indicating sourcing from the immediate north, possibly the Clue de Peroure anticline which presently defines the northern margin of the Miocene basin at Esclangon (Fig.1.2).

Further west, at **Auribeau**, contemperaneous (?) compressional uplift of the basins northern margin, resulted in the folding and subsequent erosion of marine sequences, and the development of a syn-sedimentary unconformity. The erosion sourced a local terminal alluvial-fan system, with palaeocurrents from the sheet sandstone and channel facies indicating that it flowed southward (221°N) into the basin.

In the north-west, at **St.Symphorien**, a discrete phase of progradation by a low-gradient coastal plain, mixed-load fluvial system, is recorded. Palaeocurrents from the channel facies have a mean 214°N direction indicating that the fluvial system prograded south-westward into the basin. The petrography of the fluvial systems at Auribeau and St.Symphorien is similar, but with the latter containing abundant glauconite, and conglomerate grade, glauconitic marine molasse clasts. They are both interpreted to have been derived from the reworking of Tertiary formations, in the latter case, definitely from the marine molasse and to record the tectonic uplift of the Clamensane fault block immeadiately north of the Digne-Valensole basin.

5.3.2.6 **Member Vm2**

This member records the re-establishment of marginal marine conditions in the northern and north-western margin of the basin.

The member shows a marked thickness variations, from only 30m at Auribeau, to 140m at St.Symphorien (Fig 5.2 & 5.4). This may reflect true differences in the period of establishment of marine conditions, and in the subsidence rates in these areas, but, given the lack of physical or bio-stratigraphical correlation, it is possible that it is the result of the incorrect correlation of local marine incursions.

At **St.Symphorien** the base of the member is marked by a gradational fining-upward passage from alluvial overbank facies of the Vf1 member into lagoonal bay facies followed b y a coarsening-upward flood delta facies (see Fig. 4.45) The basal 90m of this member is dominated by two superbly exposed, lagoonal bay to flood delta coarsening-upward sequences (Fig. 4.45-4.46) and which are seperated by upper estuarine tidal flat facies. The upper part of the member shows a progradational passage back into fluvial facies associations (of member Vf2), through an association of upper estuarine tidal channel and flat facies sequences.

At **Auribeau** the member is only some 30m thick and comprises two, tidal channel to supratidal flat, fining upward sequences of the upper estuarine facies association (Fig. 5.11)

In the south of the basin, at **Trois Bastides,** and at **Mirabeau** (Sourribes), the upper part of the Vm1-2 member is exposed and comprises repeated estuarine tidal channel and tidal flat facies sequences.

Interpretive synthesis

This member records the last stages of marine deposition in the Digne-Valensole basin. In the north-west of the basin (St.Symphorien) palaeo-current data from the tidal channels and foresets of the flood tidal-deltas indicates that they prograded toward the NNE-NNW (landward), and that the establishment of marine conditions was achieved by the northward migration of a barrier-inlet shoreline system. The development of two distinct flood-delta sequences in the transgressive basal part may record a discontinous transgression, or simply lateral migration of tidal inlet channels during the transgression. Palaeocurrents from the flood delta and its tidal 'feeder' channels have a bimodal SSW/SSE - NNW/NNE trend, with a clear predominance of the latter (landward).

The upper estuarine setting established was succeeded by the re-establishment of the prograding coastal plain fluvial system (as detailed in the Vf1 member) whose deposits form the Vf2 member.

At **Auribeau**, the member records the re-establishment, for the last time, of estuarine marine conditions. The position of this marine interval above a progradational mega-sequence as at St.Symphorien suggests that it records the same shoreline transgressive event, but given that it involves only upper estuarine facies it is possible that it records only the lateral shifting of estuarine and alluvial fan facies belts.

5.3.2.7 **Member Vf2**

This member gradationally overlies the marine deposits of member Vm2, and marks the resumption (2^{nd} phase) of progradation of southward flowing fluvial systems from the northern margin of the uplifting of the Digne-Valensole basin (Fig. 5.2 & 5.5). This progradational phase culminates in the establishment of the Valensole braided river system which extended across the whole of the depocentre of the Digne-Valensole basin (Debrand-Passard *et al.*, 1984) and deposited the Valensole formation.

At **Auribeau** the Vf2 member comprises a large scale (100m), coarsening-upward megasequence of the alluvial fan facies association (Fig 4.80), with sheet sandstone and mudstone facies at the base succeeded by ribbon channel conglomerate facies. Channel bodies at the top of the sequence attain thicknesses of 2-3m, with cobble-boulder grade clasts of reworked marine molasse sandstone. The top of the mega-sequence is truncated by the Digne thrust sheet.

At **St.Symphorien** the Vf2 member is a 240m thick coarsening-upward succession of channelised sandstones and mudstones of the alluvial plain facies association. Internally the succession comprises repeated, fining-upward, channelised sandstone to floodplain facies sequences of 10-30 m thickness. The top of the member is marked by the abrupt but conformable passage into the massive conglomerate facies of the overlying Valensole formation (Fig. 5.12 - 5.13).

The member is of much reduced thickness (40-60)m at **Mirabeau** where it comprises repeated channelised sandstone and mudstone facies sequences of the alluvial plain facies association. These gradationally overlie upper estuarine channelised sandstone and tidal flat facies with the channel fills vertically becoming progressively more 'uni-directional'.

Interpretive synthesis.

This member is interpreted to represent a continuation of the Vindobonian progradational 'megasequence' as recorded in the Vm1, and Vf1, members, and 'interrupted' by the transgressional Vm2 member. This late Vindobonian progradation of fluvial systems into the Digne-Valensole may be interpreted to record the continued (pulsed?) uplift of the area immediately north the basin and the consequent increase in the sediment supply rate to the basin.

At St.Symphorien palaeocurrents from the channel facies of the mixed bed-load fluvial system have a mean SSW orientation. To the west at Auribeau a similar SSW directed mean palaeoflow was obtained from the channel facies of the terminal alluvial fan system.

5.3.3 **Jabron Basin**

5.3.3.1 **Introduction.**

In the Jabron basin, correlation between the successions in the outliers of Montbrun, Melvouillon and Lange (Fig. 4.4) has proved difficult. Separated by a distance of some 18 kms, the successions comprise distinctly different lithologies and facies associations. The stratigraphy of the successions established by Goguel *et al.* (1964) has been adopted (Fig 1.11).

5.3.3.2 **Burdigalian: Members Bm1 - Bm2**

A 200m thick succession of Burdigalian (undifferentiated) siliciclastic sandstones and mudstones is developed at **Lange.** It overlies poorly exposed Oligocene mudstones and Cretaceous limestones, having a low angle of unconformity with the latter (5-10°) and unknown relationship with the former.

The basal 60m of the succession is dominated by heterolithic tidal flat and channelised sandstone facies of the upper estuarine facies association. Channel fill sequences are small, ranging in thickness from 1m to only 2.5m. Metre scale units of heavily bioturbated sandy, mudstone with small channels, 'choked' with brackish bivalves are particularly characteristic of an upper estuarine setting. The succeeding 40m of the succession is dominated by this brackish mudstone facies.

The very upper part of the succession is poorly exposed but clearly comprises a continuation of the upper estuarine facies associations of the basal section.

At **Montbrun** and **Melvouillon** (Fig 4.4) lower and upper Burdigalian sub-members (Bm1-Bm2) are differentiated (see Fig 1.11).

At Montbrun the base of the formation is marked by a thin shoreface transgressive lag of coarse pebbly sandstone with beds of bored cobbles of Mesozoic limestone and chert.

At Melvouillon the Cretaceous substrate is heavily bored to a depth of about a metre beneath a transgressive shoreface lag facies of interbedded conglomerate (formed of bored clasts of the substrate) and trough cross-stratified clastic sandstone (also substrate sourced).

At both localities, the overlying lower Burdigalian sub-member (Bml^{j}) comprises a distinctive succession of trough cross- stratified bioclastic grainstones of the shelf sandwave facies association (section 4.6.2.1) which attains a thickness of 50m. No internal sequential organisation is apparent within this sub-member.

Upper Burdigalian deposits are restricted (in their preservation) to the Montbrun area in the west of the Jabron basin, where they show a major change from the lithofacies of the lower Burdigalian sub-member. The base of the member is marked by an abrupt, non-erosional passage from coarse grainstones to fine grained, heterolithic siliclastic sandstones and mudstones attributed to an inshore, tidal setting. This facies comprises, metre scale intervals of heterolithic, current ripple laminated sandstone and silty mudstone. Palaeocurrents from the ripple sets are bidirectional, and there is no evidence of any wave or storm activity.

Interpretive synthesis

Lower Burdigalian

The development of a transgressive shoreface lag facies at the base of the member in the west of the Jabron basin (Montbrun and Melvouillon) indicates that low rates of sediment supply and subsidence accompanied the transgression, with the sediment forming the facies interpreted to have been supplied from coastline erosion of the transgressed Cretaceous substrate. The subsequent deposition of the shelf sandwave facies is interpreted to record the establishment of an open marine shelf setting with vigorous tidal currents in the west of the Jabron basin as the Rhone sea extended eastward into the Jabron basin (see Fig 5.18).

After Demarq *et al.* (1974) the entrance to the Jabron seaway is marked by submarine 'palaeo-cliffs'. Interestingly he position of these palaeo-cliffs is noted to be coincident with the position of the major Nimes fault (Fig 5.18).

Analagous, and contemperaneous, carbonate-shelf lithofacies have been described from the Rhone valley by Demarq *et al.* (1984) who suggest that it was restricted in its development to shallow coastal and 'platform' (shelf) regions on the margins of the Burdigalian, Rhone seaway (Fig 5.18). The lithofacies has also been described from borehole core of Burdigalian successions taken from offshore Tertiary basins in the Gulf of Lions (Cravatte 1974) suggesting that it is the characteristic lithofacies of the open Rhone seaway.

The development of carbonate shelf facies such as this, would have required a negligible supply of siliciclastic detritus, and high bio-organic activity to generate the large volumes of bioclastic material involved. In analogous modern-day systems the bulk of the bioclastic sediment is produced in the subtidal zone (Sellwood 1986). Remnants of low energy, reefal facies preserved along the margins of the Rhone valley (Roiure *et al.* 1980) are envisaged as having as been the subtidal 'carbonate factories' which supplied material to the Rhone shelf.

In the Jabron basin palaeocurrent data from the sandwave facies at Montbrun shows bi-modal NNE-SSW trend, with a strongly dominant SSE trend. This trend is parallel to the strike of the high-angle Lanamon (Sault) fault zone suggesting that tidal currents may have been structurally constrained in this area (Fig. 5.18).

At Melvouillon the stratification has a unidirectional, SE trend.

The developement of restricted upper estuarine facies at Lange suggests that the eastern end of the Jabron seaway was headed by a clastic shoreline system as schematically envisaged in Fig 5. The development of the relatively thick clastic sequences required high rates of siliciclastic sediment supply and may have been achieved by direct supply from an alpine sourced fluvial system. Alternatively or may indicate linkage (temporary ?) to the clastic marine, Digne basin which lay some 20km to the east.

Upper Burdigalian

The major change from bioclastic to siliciclastic lithofacies in passing from the lower to upper Burdigalian is interpreted to record a marked reduction in the strength of the tidal regime in the Jabron gulf, with an abrupt cessation of the supply of bioclastic material suggesting reduced connection with the Rhone seaway (Fig 5.19). In the absence of any eustatic sea level fall during the Burdigalian (Vail 1979) the constriction of the Jabron gulf is considered to have been achieved tectonically, through uplift immeadiately along the Nimes fault (Fig 5.19). Active faulting is a characteristic feature of the Neogene of the region (see Chapter 6) and so the mechanism for constraining the gulf is available.

5.3.3.3 **Vindobonian** (Helvetian).

Helvetian deposits are restricted to Montbrun where they form a 40m thick, coarsening-upward **alluvial fan facies sequence** which gradationally overlies upper Burdigalian marine deposits. The base of the member comprises sheet sandstone and ochrous mudstone facies. These coarsen-upwards into conglomerates of the ribbon channel conglomerate facies. The conglomerates are of sub-rounded cobbles and pebbles, and comprise an admixture of Mesozoic carbonates, and Miocene siliciclastic sandstones and grainstones

Interpretive synthesis.

The Helvetian is marked by the cessation of marine conditions and the progradation of an alluvial fan system into the Jabron basin. Palaeocurrent data from gravels of the alluvial facies indicate that the fan system flowed northward (256°N mean palaeoflow), and suggests that active tectonic uplift and erosion of the southern margin of the basin occured during the Helvetian.

5.4 **Source of Sediment of the Marine Molasse Formation.**

The development of thick successions of siliciclastic shallow marine deposits implies high rates of sediment supply. The identification of the source of these sediments is addressed on two levels;

(1) Petrography of the marine sandstones.

(2) Sediment supply.

5.4.1 **The petrography of the marine sandstones**

The petrography of the siliciclastic marine sandstones (see section 4.5) of the Digne-Valensole basin and much of the Jabron basin bears a close similarity to that of the fluvial sandstones of the underlying, Molasse Rouge formation which were interpreted to have been primarily derived from the reworking of older, alpine foreland basins (see section 3.10.1).

The composition of the marine sandstones is also, as might be expected, closely similar to that of fluvial-distributary sandstones of the formation. Clast lithologies in the fluvio-distributary facies, and also in the transgressive conglomerate facies, comprise in decreasing abundance, Jurassic-Cretaceous limestones, and cherts, vein quartz, Permo-Triassic sandstones, granites, and serpentinised gabbro ('roches vertes'). The composition of these clasts indicates that the principal, conglomerate-grade clast source was the Mesozoic cover sequences of the external alps. The presence of Permo-Triassic, igneous and metamorphic lithologies may record erosion of internal, or external, alpine basement massifs, or alternatively the reworking of older Tertiary sequences of the foreland basin.

5.4.2 **Sediment Supply:** - Sediment may be supplied to marine systems by:

(i) direct fluvial supply

(ii) coastal erosion during transgression.

It is obvious that the quantity of sediment within the formation could not have been supplied by coastal erosion alone, whilst there is abundant evidence of direct fluvial supply to the basin, during this period.

The thick, fluvial sequences of the underlying Molasse Rouge formation (Esclangon system) record the development of an 'alpine sourced' (reworked foreland basin) fluvial distributary system in the north of the basin immediately prior to the Burdigalian transgression (section 3.10). In the north-east of the basin (Esclangon), the development of a fluvio-distributary facies (south-westward prograding) within the marine sequences (Bm1 and Vm1 members), clearly shows the continuation of 'internally' sourced, fluvial systems in this part of the basin during the Miocene.

In the North-West of the basin (St.Symphorien) the establishment of a south-westward prograding mixed-load fluvial distributary system (Vf1&2 members) also suggests the maintenance of, or at least renewal of, the Oligocene fluvial system in this area.

5.5 Controls on Transgressional - Progradational Cycles in the Digne and Jabron Basins - A Discussion.

The organisation of facies sequences in the Digne-Valensole and Jabron basin indicates that they had a complex history of shoreline migration. This may have been the result of one, or of a number of, variable factors, namely, eustatic sea level rise, regional tectonic subsidence, basinal fault tectonics, and fluctuating sediment supply rates. In this section each of these controls will be discussed in an attempt to elucidate their role during the Miocene. The controls considered are:

(i) Eustatic sea-level changes.
(ii) Regional tectonic subsidence.
(iii) Basinal faulting.
(iv) Sediment supply rate.

(i) **Eustatic sea-level changes.**

Examination of the regional sedimentation pattern of S.E France during the Miocene reveals a simple trend of early Miocene transgression and late Miocene regression of a 'peri-alpine seaway', which may be related to the Cenozoic eustatic sea level curves of Vail & Harbendol (1977) and Haq *et al.* (1987) (Fig 4.1).

Sea level curves have been the subject of much debate and criticism (Watts. 1982, Watts *et al.* 1982, Pitman 1987, and Haq *et al.* 1988). However given that (1) the Miocene seaways of S.E France had an open connection with the North Atlantic during this period (with the exception of the Messinian; see Cavelier *et al.* 1984) (2) most of the sea level data was derived from the North Atlantic; the curves are considered to provide an important contribution to the understanding of the alpine seaway.

Considered in isolation these curves reflect the 'background' sea level trend, and therefore provide an insight into when such factors as sediment supply, and tectonic subsidence, played an active role in shoreline movement.

Returning to the regional sedimentation history in more detail, the onset of a transgression in the lower Aquitanian and its continuation to a Langhian - Serravallian high in S.E France

(Coppolani *et al.* 1973, and Cavelier *et al.* 1984) is seen to be contemporaneous with the lower Miocene, eustatic sea level rise documented by Haq *et al.* (1987). The upper Vindobonian-Messinian eustatic sea level fall, which resulted in a fall of sea level of some 200-300 m in the Tethyan sea (Mediterranean) (Demarq & Perrieux 1984) and the withdrawal of marine conditions from most of S.E France, correlates with the latter part of the Vindobonian progradational phase in the Digne-Valensole and Jabrob basins.

Therefore, though it may not have been the only active factor, eustatic sea level changes alone, were apparently capable of generating the gross transgressional-progradatational fills of the Durancian basins fills, with the proviso that the onset of progradation in the Digne-Valensole and Jabron basins preceeded that of the eustatic sea level fall.

(ii) **Regional tectonic subsidence.**

In the context of the study area, regional subsidence associated with both the Western Mediterranean rift system, and the alpine thrust belt, need to be considered as possible controls on sea level.

A model of generating regional scale, sea level rises by extensional lithospheric flexural subsidence has been proposed by Watts (1982), Watts *et al.* (1982). In the South of France, the onset, in the Aquitanian, of the lower Miocene transgression is contemporaneous with sea floor spreading and the initiation of the thermal subsidence phase of extension in the Western Mediterranean rift system (see section 1.3.3).

Further evidence of regional extensional tectonic control is provided by the palaeo-morphology of the lower Miocene sea with the major Rhone seaway developing along the axis of the Rhone graben, and the E-W trending Durancian Gulf extending westward toward the alpine thrust front along narrow seaways between 'palaeo-highs' sited above transtensionally reactivated, inherited high-angle faults.

Thrust-sheet driven, flexural subsidence was also active during this period. Flexural subsidence has been appealed to, to account for regional transgressions in the Rocky mountain foreland (Mclean & Jerzykiewicz 1978), and closer to the study area, has been suggested to have played a contributory role (together with eustacy) to the Aquitanian-Burdigalian transgression in the Swiss molasse basin (Homewood *et al.* 1986).

Consideration of the migration history of the S.W alpine foreland basin, together with subsidence curves from the area, suggest that flexural subsidence played an active role in the thrust front, Digne-Valensole basin during the Miocene, but only a minor role further out into the foreland (see section 6.3 for discussion). Subsidence induced by alpine thrust loading may therefore provide a means of generating subsidence phases and consequently shoreline transgressional and progradational events in, and unique to, the Digne-Valensole basin.

(iii) **Basinal faulting.**

Syn-sedimentary faulting as a possible control on transgressional-progradational cycles, has been detailed by Young 1957, Clifton 1958, Weber 1971, Evarry *et al.* 1978, Elliott 1986).

In the Durancian basin, fault control on the palaeo-geometry of the Miocene seaways (Fig 4.3 after Demarq & Perrieux 1984) is suggested as they developed between palaeohighs sited above the position of inherited high-angle, crustal faults (Fig. 6.6.3) which are interpreted as having been transpressionally reactivated during the Miocene (Fig. 5.10).

Direct evidence is given by the Burdigalian sequences of the Apt basin which have been shown to onlap an E-W trending palaeo-high (Jones 1988) developed above the major Luberon fault. In the Digne-Valensole Gulf, the western and northern shorelines apparently remained parallel to the high-angle, Durance and Sorine faults (respectively) thoughout the Burdigalian and Vindobonian suggesting that they played a fundamental role in defining the gulf's geometry.

In the Digne-Valensole and Jabron basins there is also abundant evidence, in the form of lateral thickness variations in basin fill, of active transpressional, and transtensional faulting (see Figs 6.5 & 6.6). Active syn-sedimentary faulting was also speculated to have controlled the palaeo-tidal flow in the lower Burdigalian at Montbrun in the Jabron basin, and also controlled facies changes, including the develop ment of locally sourced alluvial fans, in the upper Burdigalian-Vindobonian at Montbrun (section 5.3.3).

Further evidence of active Miocene faulting, in this case compressional uplift, is provided by the developement of late Miocene (Vindobonian) syn-tectonic unconformities and a system of alluvial fans on the margins of the Digne and Jabron basins (see section 5.3.2.5. & 5.3.3)

As a final point, the Pliocene uplift of the western and southern margins of the Digne fault block within the Digne-Valensole basin (Fig 2.12 - 2.13) by the transpressional reactivation of inherited high-angle faults (the Mirabeau, and Durance faults) suggests that during the late Miocene period the Digne-Valensole basin had been subject to alpine tectonic (transpressional ?) uplift.

Directly attributing movement on a particular fault to a given sedimentary cycle in the Digne-Valensole or Jabron basins is generally not possible, given the limitations of the data and we are unable to take the analysis beyond an appreciation that basinal faulting was an active factor. However, direct fault control can be inferred for the unconformity and associated alluvial fan systems developed on the northern margin of the Digne-Valensole basin (Vf1-2 member) during the Vindobonian. They are attributed to the inversion of the Melan-Clamensane fault block to the imme diate north of the Digne-Valensole basin, along the inherited high-angle Sorine fault zone. The contemperaneous (?) alluvial fan system of the Jabron basin (Vf2 member) is similarly attributed to compressional/transpressional alpine inversion along the Ventoux-Lure fault zone.

Whilst active faulting can be shown to have produced local shoreline shifts, transgressive events which span the basin and a number of internal fault blocks are considered to reflect the influence of a factor of greater areal extent than intrabasinal faulting.

(iv) **Sediment supply rates.**

Variable rates of clastic sediment supply to a shoreline system are capable of generating regional progradational-transgressional cycles (Coleman 1981, Winker 1982, Leckie 1986).

A detailed example is the progradational-transgressional cycles of the Western Interior Basin (of 25-50m thickness) which are interpreted to have been produced by pulses of clastic sediment supplied by spasmodic, thrust induced uplifts in the Rocky mountains (Leckie 1986).

In the Digne-Valensole basin, a similar direct supply from the adjacent alpine thrust belt has been demonstrated by the presence of an alpine sourced, fluvial system in the north-east of the basin (Esclangon) which is an apparent continuation of the pre-cursor, Oligocene age, 'Esclangon terminal fan system' (section 3.10). More direct sediment supply to the basin was apparently restricted to the late Miocene, taking the form of alluvial fans issued directly off the basins margin during their compressional uplift.

The dominance of progradational mega-sequences in the fill of the Digne-Valensole basin, particularly above the lower Burdigalian transgressive megasequences (sub-member Bm1a: section 5.3.2.1) is interpreted to record that sediment supply rates were extremely high, and became progressively greater during the Miocene. This is interpreted to be to the continuous creation of erosional relief in the alpine hinterland, which progressively encroached on the autochthonous Digne-Valensole basin during the Miocene.

Variation in supply from the alpine hinterland fluvial systems (including the 'Esclangon System') as a consequence of discreet phases of alpine uplift (pulsed) may have been important in the shoreline evolution of the basin and generated transgressional, and progradational cycles during the Burdigalian, but this impossible to quantify.

Certainly by the late Miocene, active alpine uplift to the north and west of the basin, as evidenced by syn-tectonic unconformities on the basins margin, resulted in the large scale progradation of fluvial systems into the basin, and the southward displacement of marine conditions.

Summary

During the Miocene sedimentation in the Digne-Valensole and

Jabron basins was subject to eustatic sea level changes, regional subsidence associated with extensional thermal subsidence, and thrust-load flexural subsidence, active basinal faulting, and variations in clastic supply. Whilst the gross transgressional - progradational nature of the Miocene sequences superficially mirrors the eustatic sea level changes of the period:
(1) the basal transgressive event must have been accentuated by regional lithospheric subsidence , both flexural loading and thermal subsidence, and
(2) the early onset of progradation in the basins in the lower Vindobonian is attributed to high rates of sediment supply from the forelandward propagating alpine thrust belt.

The basinal variation in facies sequences may be attributed to intra-basinal faulting, and localised sediment supply, but basin wide transgressive events are considered more likely to reflect pulses of thrust load induced, flexural subsidence.

5.6 Conclusions and discussion.

During the Burdigalian a regional transgression in response to eustatic sea level rise and regional subsidence resulted in the establishment of shallow marine conditions in a circum-alpine seaway which extended into the Digne-Valensole and Jabron basins.

A model of the palaeogeograpy of the Miocene sea of S.E France is presented in Fig 5.20. A marine gulf was established in the Digne-Valensole basin by the northward extension of the Rhone seaway through the Apt basin (Fig 5.14 & 5,20), whilst in the Jabron basin a shallow gulf was established by the eastward extension of the Rhone seaway (Fig 5.18 & 5.20).

Differential subsidence within the Digne and Jabron basins along transpressionally reactivated fault blocks was accompanied by direct sediment supply from the alpine hinterland resulting in the accumulation of predominantly siliciclastic, shallow marine successions which ranged in thickness from 200m to 900m. Facies analysis suggests that the Digne and Jabron basins were discrete marine basins.

The Digne gulf was N-S trending and had mixed, tide and wave influenced, tide dominated clastic shorelines during the Burdigalian and Vindobonian. The tidal regime was diurnal and probably meso-tidal. Wave energy was low to moderate, and the shorelines were storm influenced. Clastic sediment, from the alpine thrust belt, was directly supplied to the northern and north-eastern margins of the basin by fluvial (-distributary) systems which were probably established during the Oligocene. The clastic dominated Burdigalian successions of the (northern) Digne-Valensole basin contrast markedly with those of the Apt-Forcalquier basin which are bioclastic (grainstone) dominated (Jones 1988), and closely comparable to the 'open shelf' tidal facies developed in the western part of the Jabron basin. This east to west transition away from the alpine thrust belt from siliciclastic to bioclastic marine successions is as detailed in the Jabron basin and records a siliciclastic starvation of the more "offshore" (relative to the alpine fluvial systems) areas of the Durancian Gulf during the Burdigalian. This may be interpreted to be the consequence of the net landward sediment transport of tidal waves as they propogate into large bays (Swift 1975), in this case the Durancian Gulf, confining any alpine sourced clastic material to the shoreline systems. Alternatively it may record that a bed-load parting zone, as described from a number of comparable scale estuaries and gulfs, including the Severn Estuary (Kenyon *et al.* 1974 and Harris 1988), operated between the Digne-Valensole and Apt-Forcalquier basins.

The 900m thick Miocene fill of the northern Digne-Valensole basin shows an overall transgressional-progradational trend which is attributed to a combination of

(i) a, Burdigalian-Messinian eustatic transgressional - regressional cycle

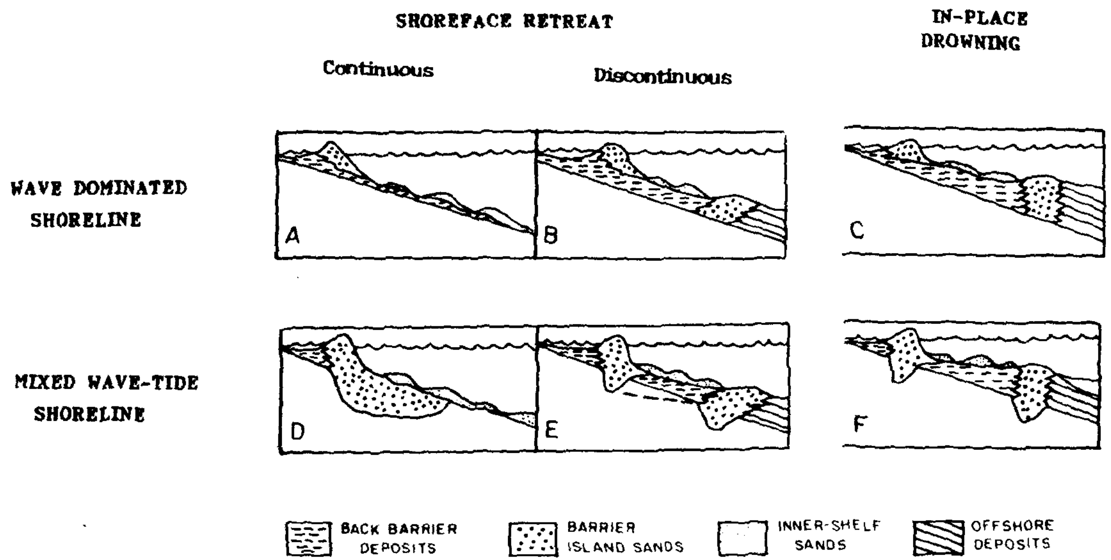
(ii) regional subsidence, and

(iii) variable sediment supply rates.

Within the basin fill four smaller scale transgressional -progradational mega-sequences (100-300m) are defined, of which three are defined by basin wide transgressions attributed to thrust induced foreland loading. The developement of coarse fluvial-systems, and a syn-tectonic unconformity along the northern margin of the Digne-Valensole basin during the Vindobonian is interpreted to reflect that the dominantly progradational Vindobanian fill of the basin was the result of active alpine tectonic uplift immeadiately north of the basin. The restriction of a fourth progradational-transgressional cycle to the north of the Durancian basin records its developement due to the pulsed progradtion of the alpine clastic wedge.

The Jabron gulf was tide dominated and during the lower Burdigalian had an inner-gulf clastic sourced, estuarine system, and an outer-gulf, bioclastic sourced (clastic starved) open shelf sytem, as discussed above.

The basin fill also shows a simple transgressional -progradational cycle, with late Burdigalian-Vindobonian progradation being accompanied by intrabasinal faulting and the uplift of the Vaucluse region to the south of the basin.



A: Erosional shoreface retreat , reworking of backbarrier sequence developement of well-defined ravinement surface and deposition of transgressive sand-sheet.

B: Erosional shoreface retreat punctuated by progradation of barriers.

C: In place growth, almost complete preservation of the backbarrier sequence

D: Erosional shoreface retreat with rapid inlet migration. Scouring by migrating inlets reworks the backbarrier sequence and deposits an inlet-filling sand body. Barrier retreat leads to deposition of a thick blanket of inlet fill deposits.

E: Erosional shoreface retreat punctuated by progradation with rapid inlet migration results in thick barrier sandstone deposits. Resumption of shoreface retreat planes off the barrier and the upper part of the backbarrier sequence.

F: In-place growth and overstep with rapidly migrating inlets. Absence of inlet-fill sediments between drowned barrier and new barrier.

FIG 5.1 Models of stratigraphic sequences along barrier-island coasts undergoing submergence (after Rampino & Sanders 1982).

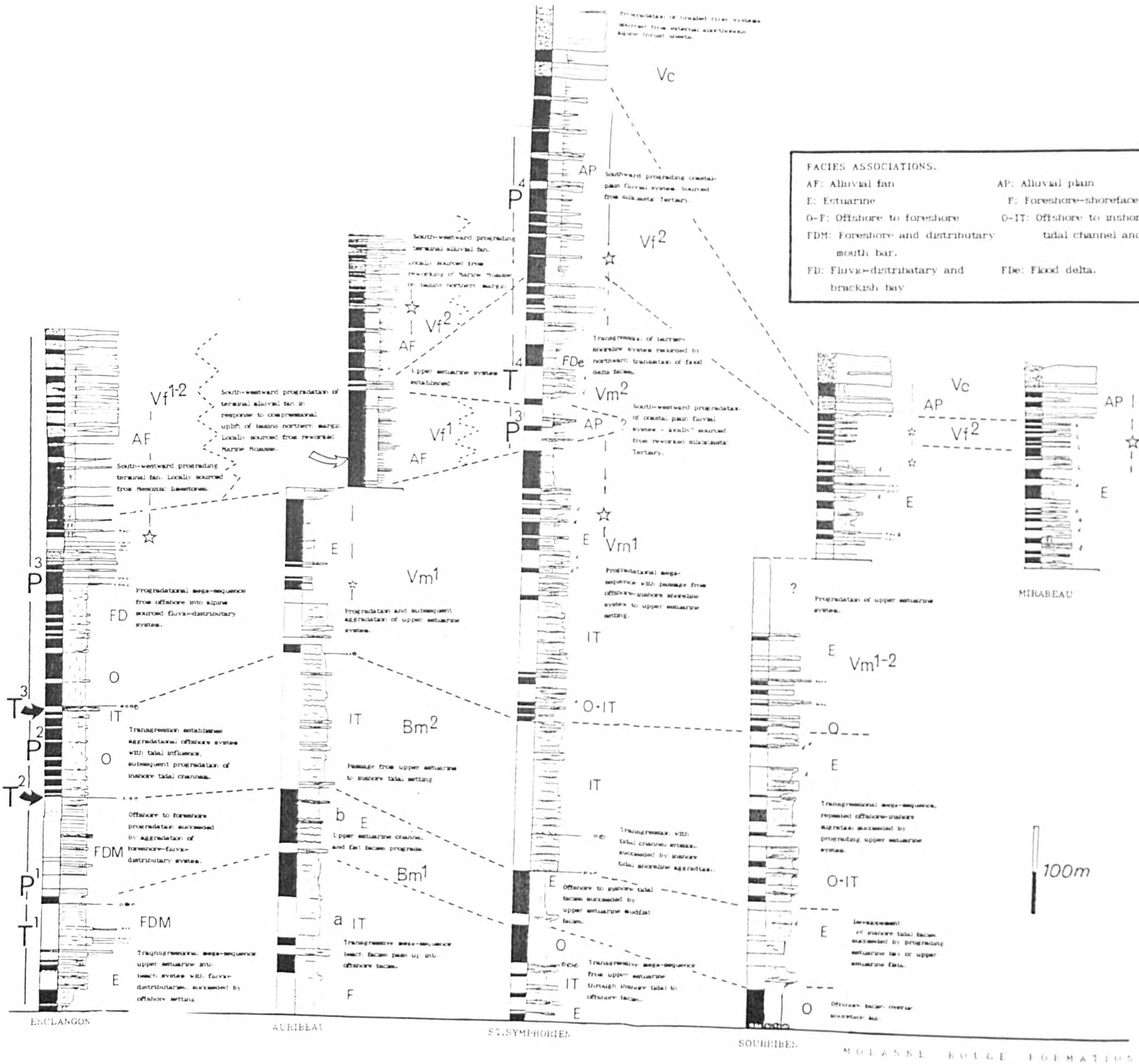


FIG. 5.2 Lateral and vertical distribution of facies associations of the Marine Molasse formation within the Digne -Valensole basin. Members Bm1 - Vf2 are recognised which define four major transgressional-progradational cycles P-T1-4 (mega-sequences). Localities of sections are given in Fig. 4.2.

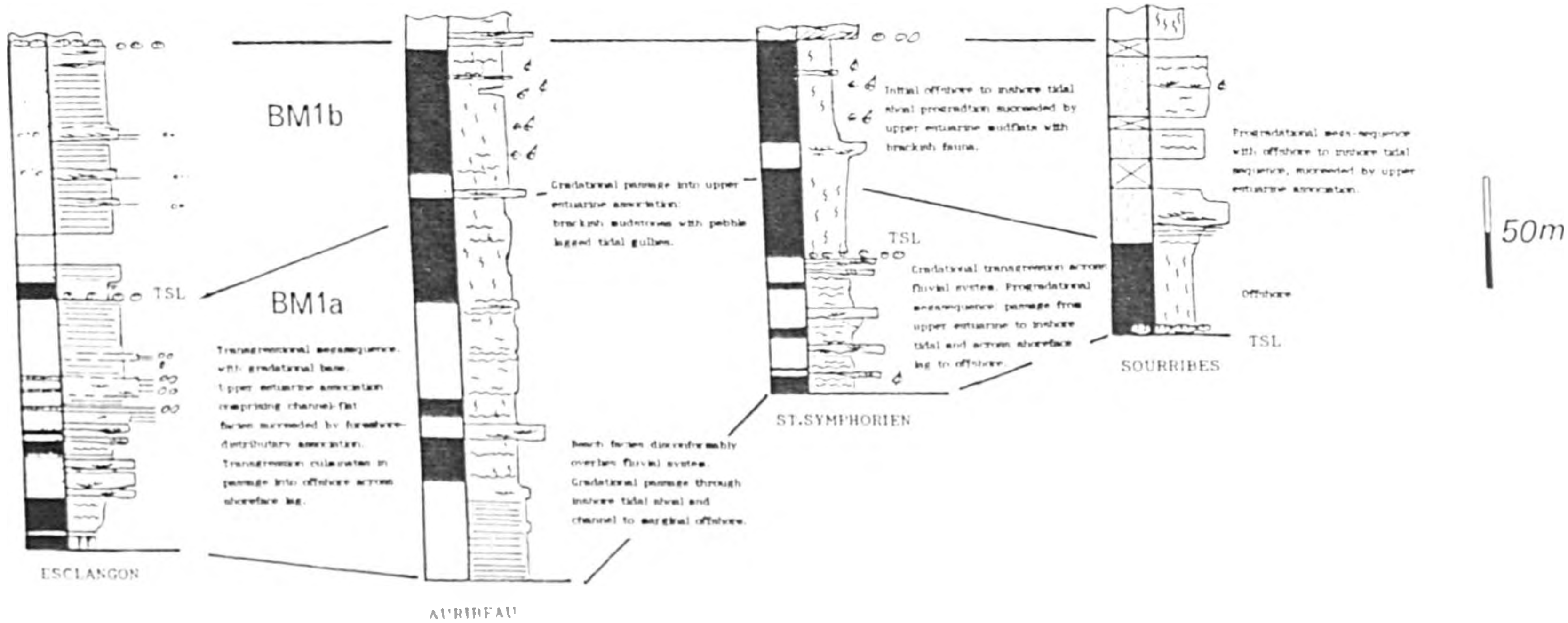


FIG 5.3 MEMBER BM1 - TRANSGRESSIONAL-PROGRADATIONAL CYCLE.
Sub-member Bm1a records transgressive establishment of shallow marine conditions across basin during the lower Burdigalian
Sub-member Bm1b records progradation of wave and tide dominated shorelines.

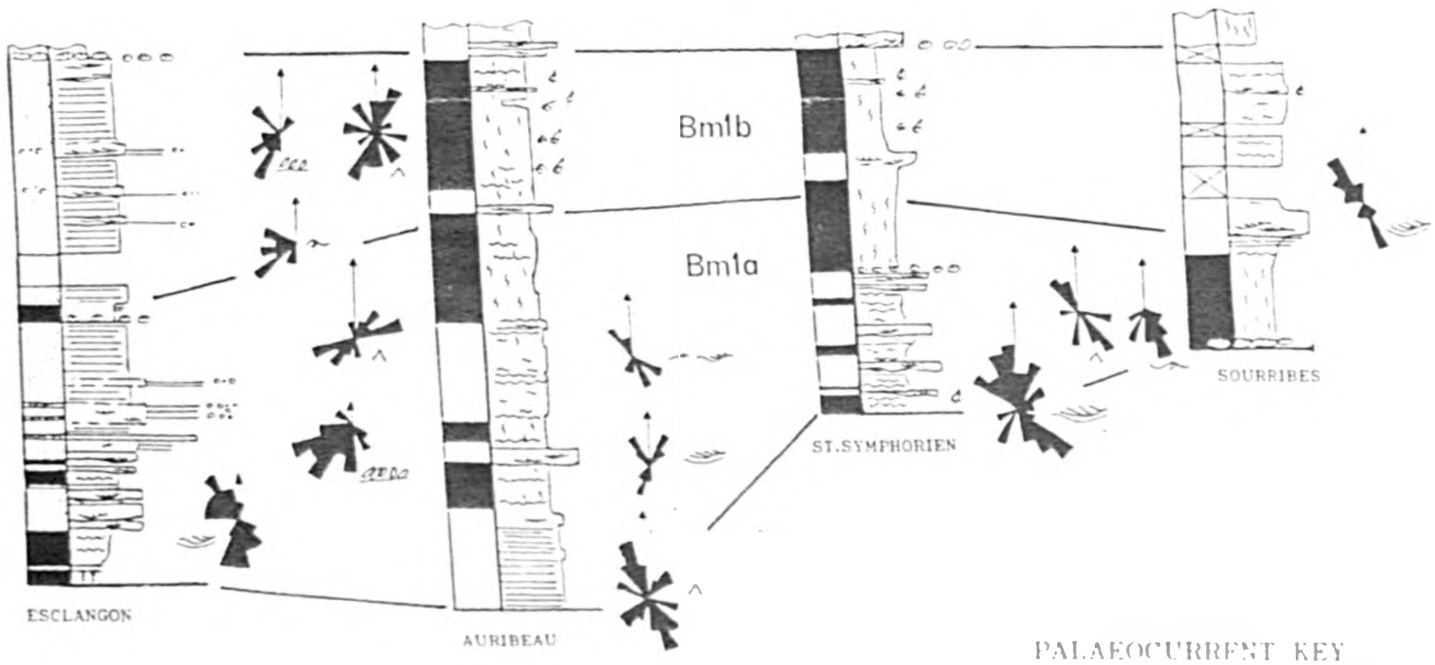


FIG 5.3 (i) Palaeocurrent data.

PALAEOCURRENT KEY
Current ripple orientation
Direction of wave oscillation
Channelised tidal stratification (f: fluvial)
Storm currents
Fluvio-distributary gravel stratification
Alluvial fan gravel stratification

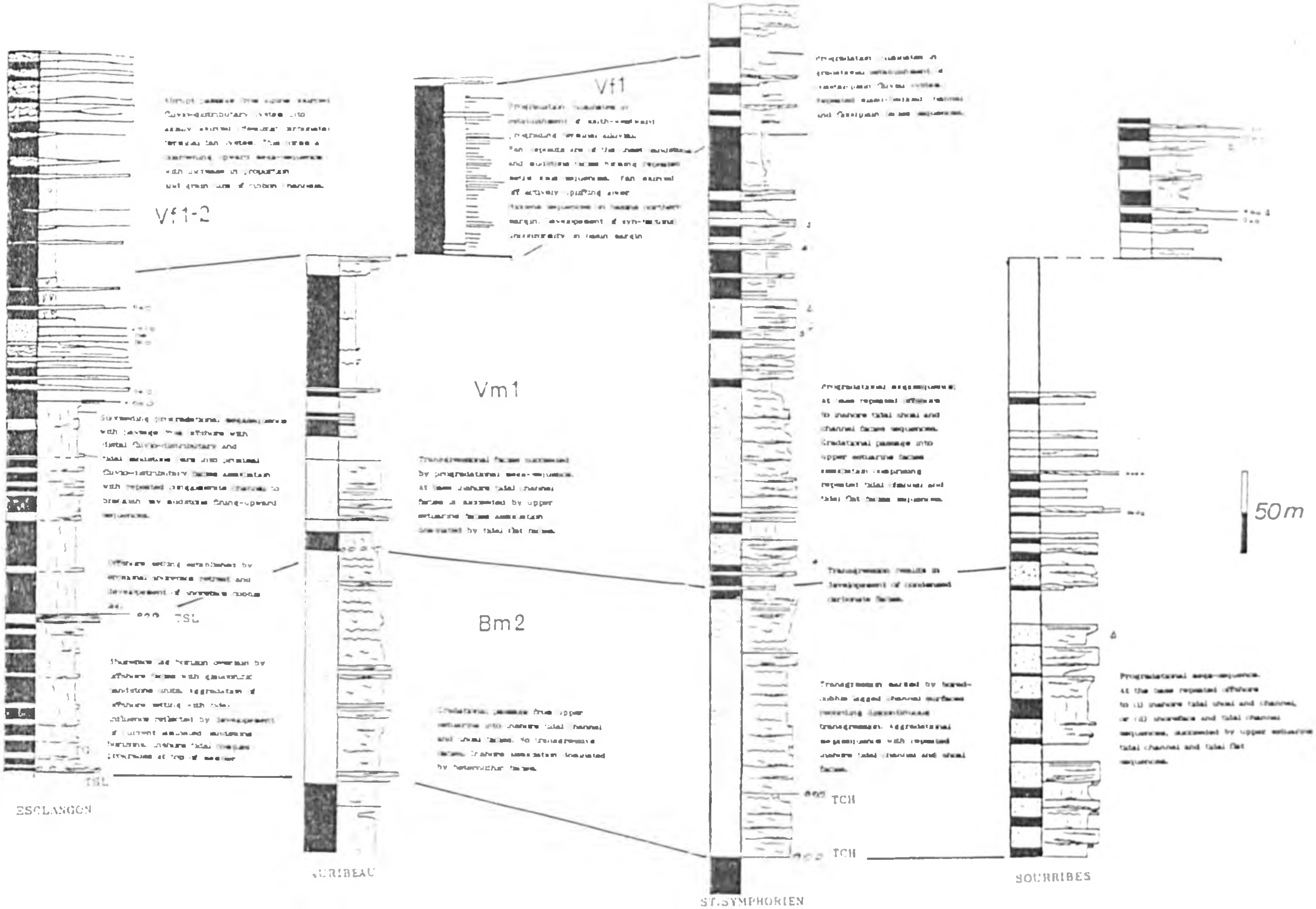


FIG 5.4 MEMBERS BM2, & VM1. TRANSGRESSIONAL-PROGRADATIONAL CYCLES. Members BM2 and VM1 are dominated by progradational megasequences which develop above thin transgressive facies / facies sequences.

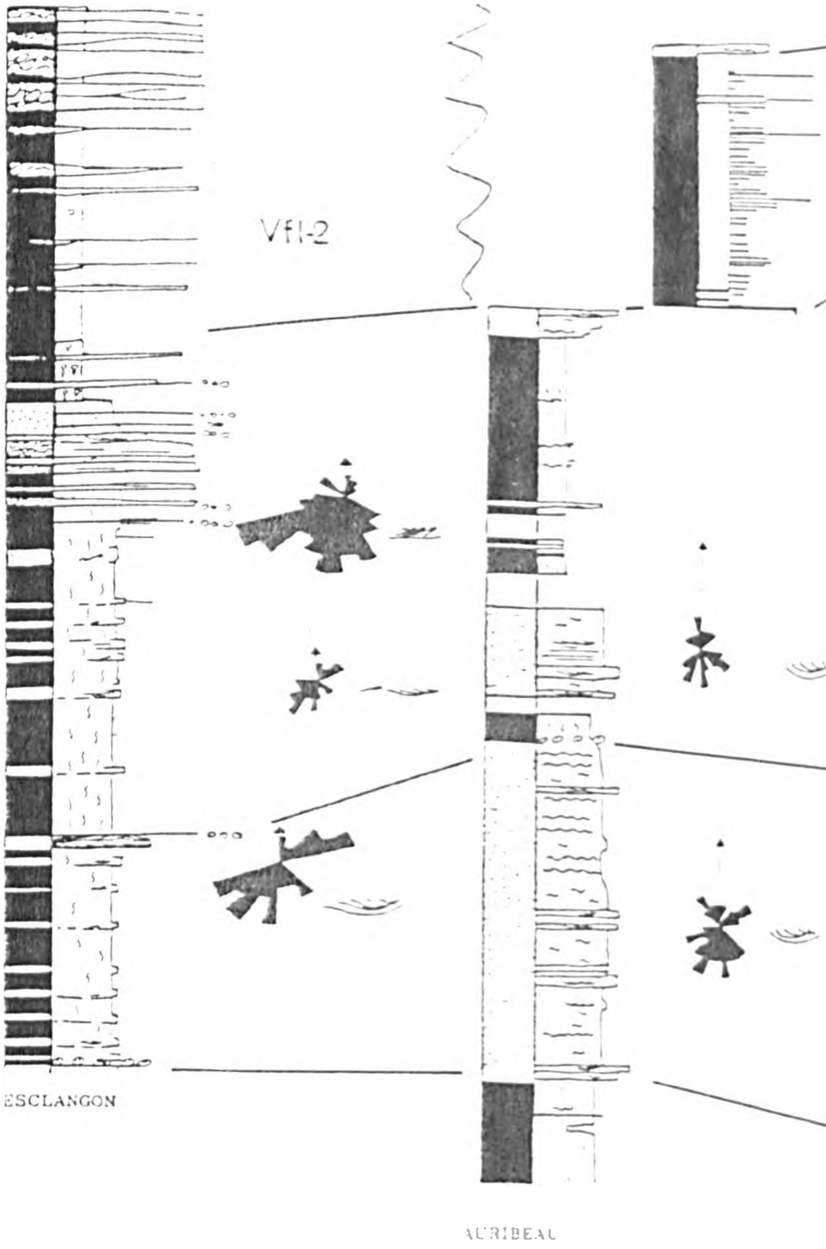
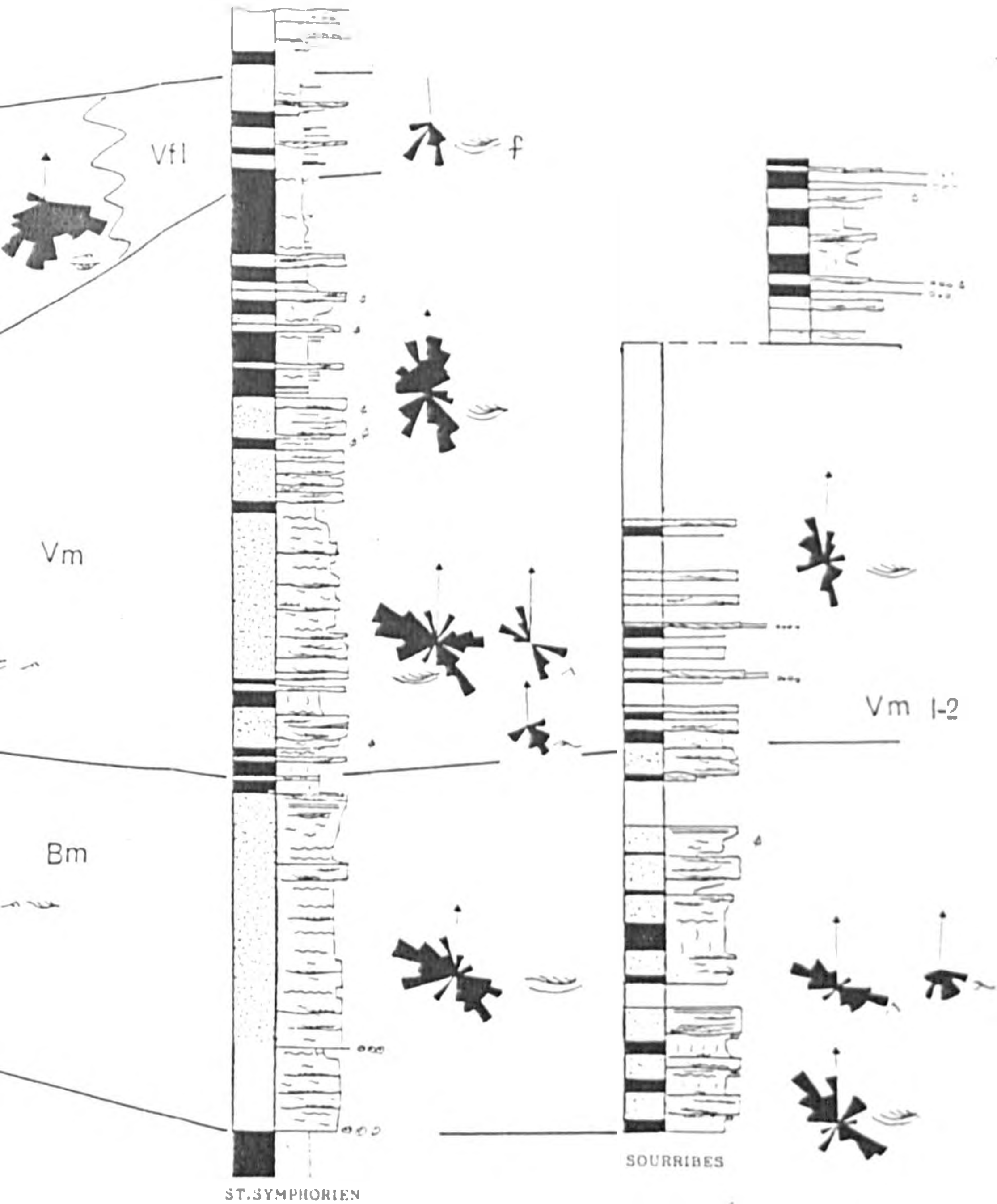


FIG 5.1 (i)



Palaeocurrent data.

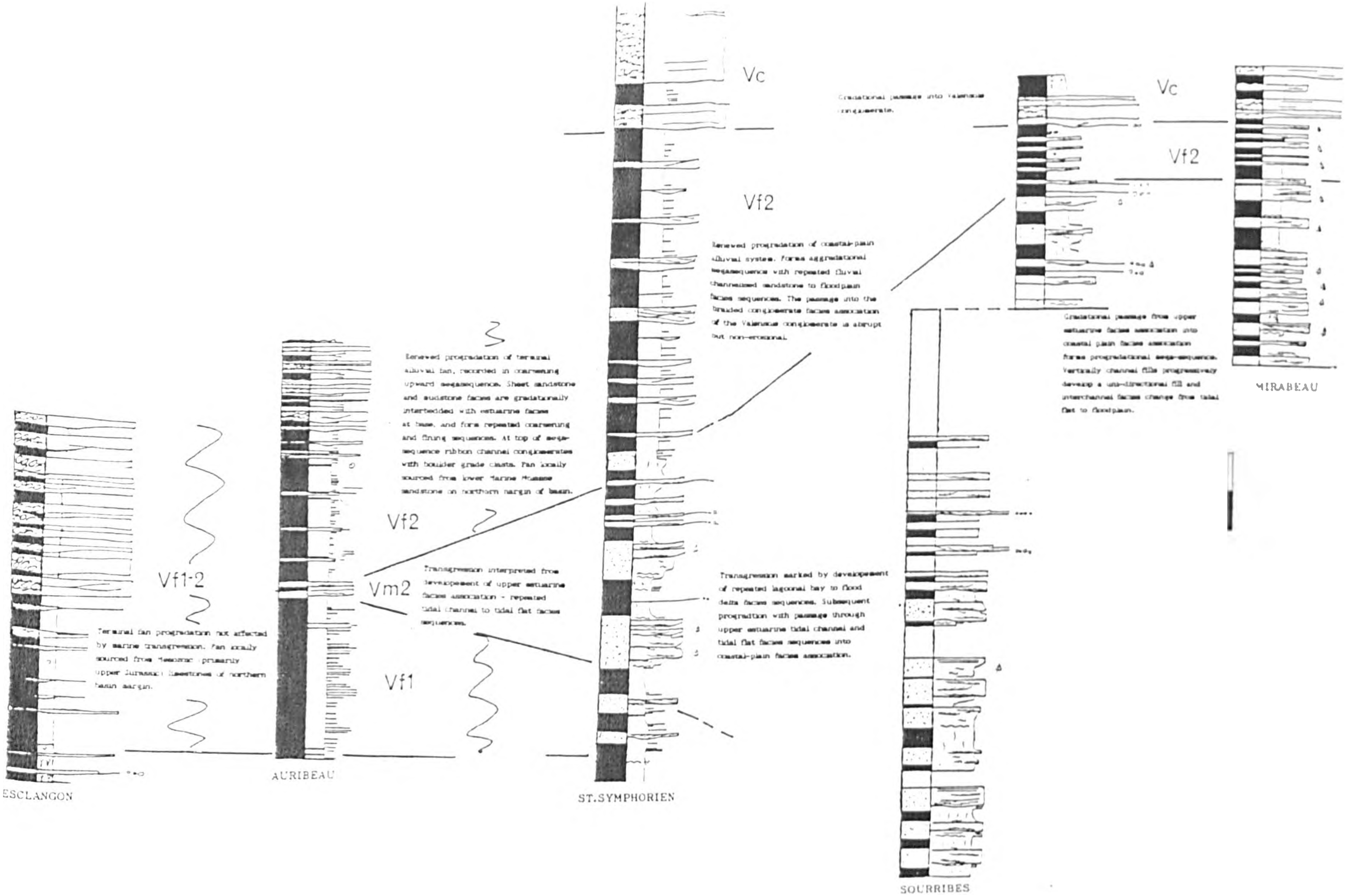


FIG 5.5 MEMBERS VM1/2, AND VF1/2 TRANSGRESSIONAL-PROGRADATIONAL CYCLE. Transgressive member Vm1 fails to extend to Esclangon where locally sourced alluvial fan progrades. Fluvial Member Vf1 restricted to north of basin. Member Vf2 extends south to Mirabeau and is gradationally succeeded by Valensole Conglomerate formation which progrades across whole of Digne-Valensole basin.

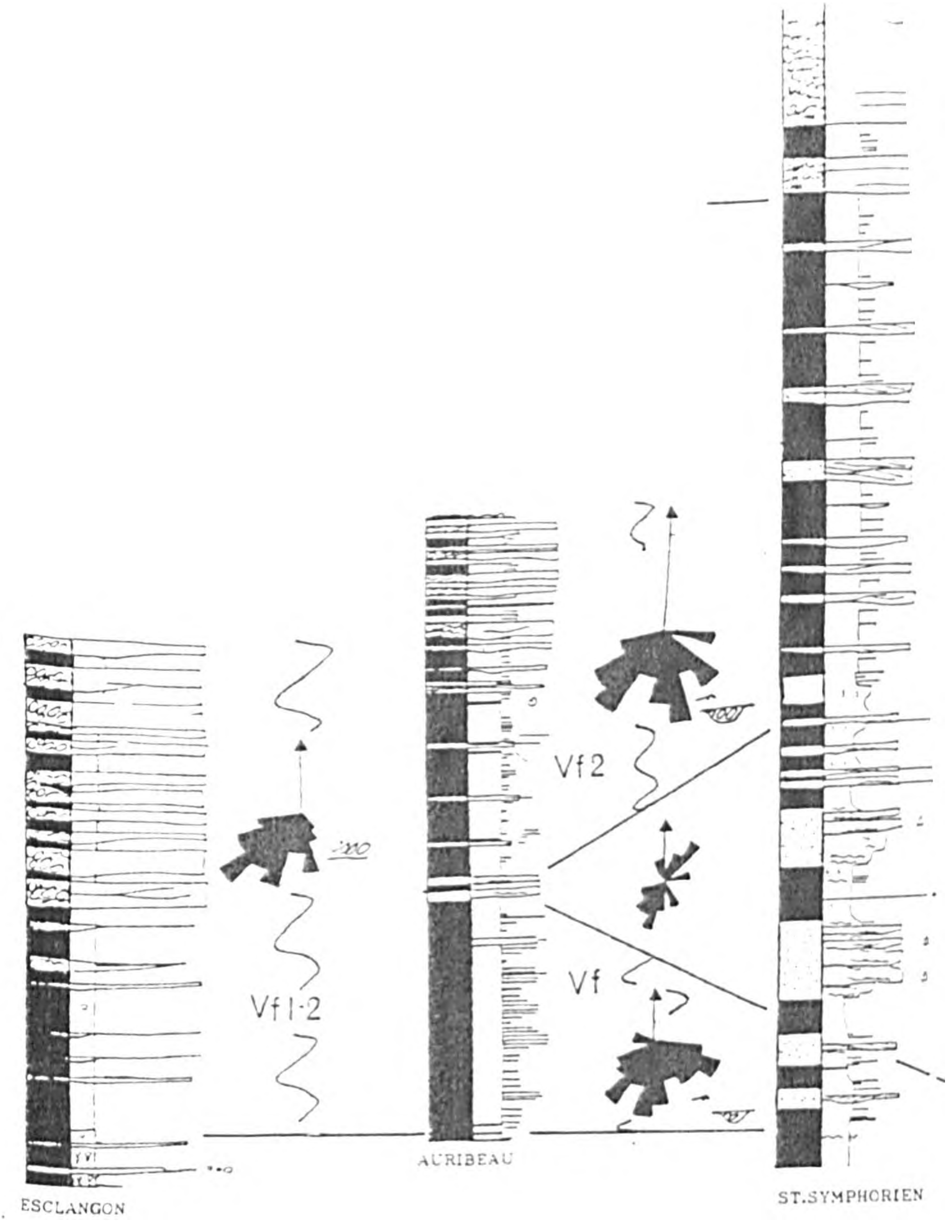


FIG 5.5 (i) Palaeocurrent data.

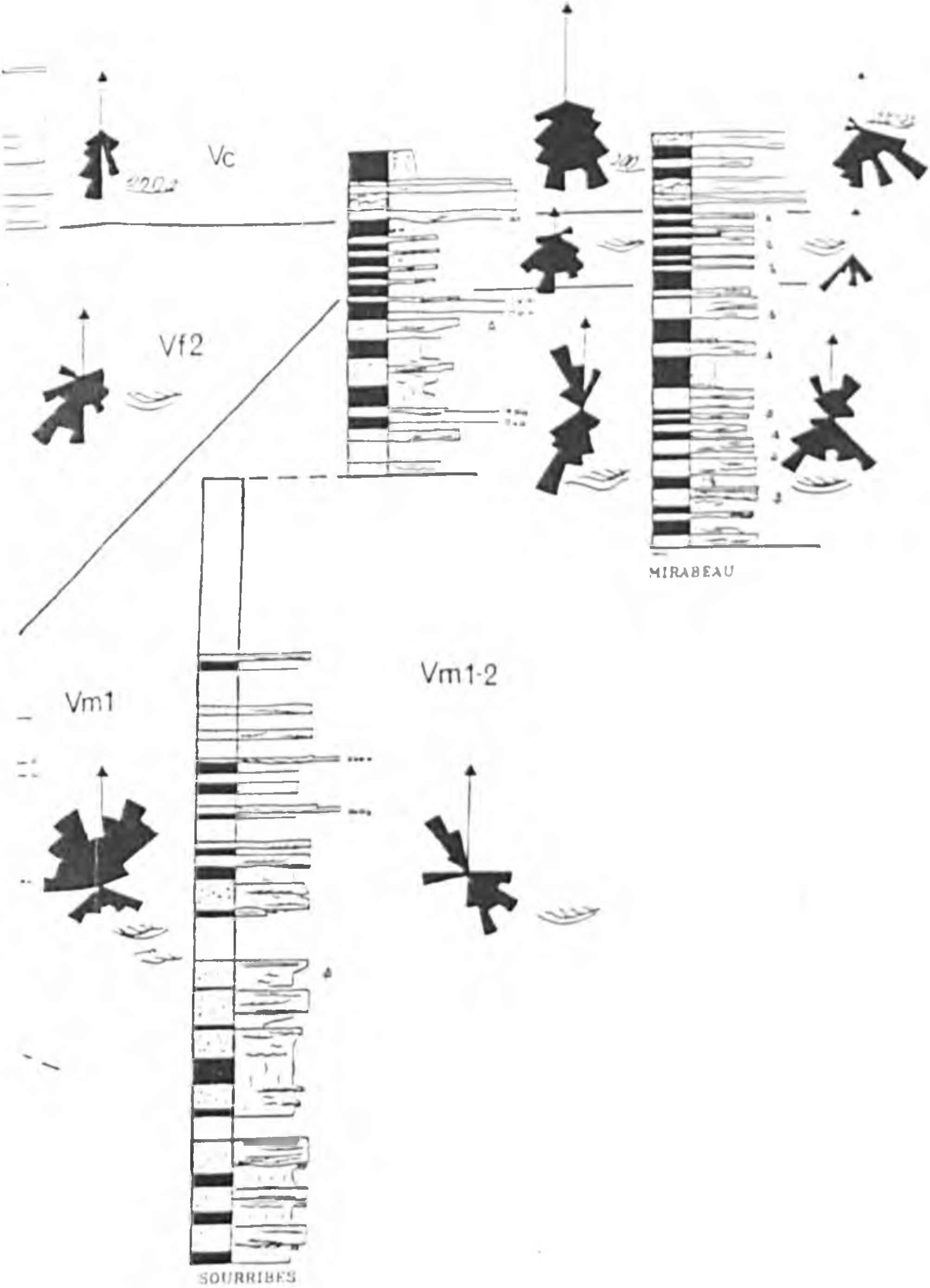

Vc
Vf2
MIRABEAU
Vm1
Vm1-2
SOURRIBES

FIG 5.6 Syn-sedimentary unconformity within the Marine Molasse formation at Auribeau, on the northern margin of the Digne - Valensole basin. Alluvial fan deposits of the Vf1-2 member unconformably overlie shallow marine Burdigalian-lower Vindobonian (?) members (Bm1-Vm1) folded in the footwall syncline to the parautochthonous Melan-Clamensane thrust sheet (fault block). Over a distance of 3km toward the south-west (into the basin) the unconformity passes into a gradational sedimentary contact which takes the form of interbedded estuarine (Vm 1) and alluvial fan (Vf1) facies.

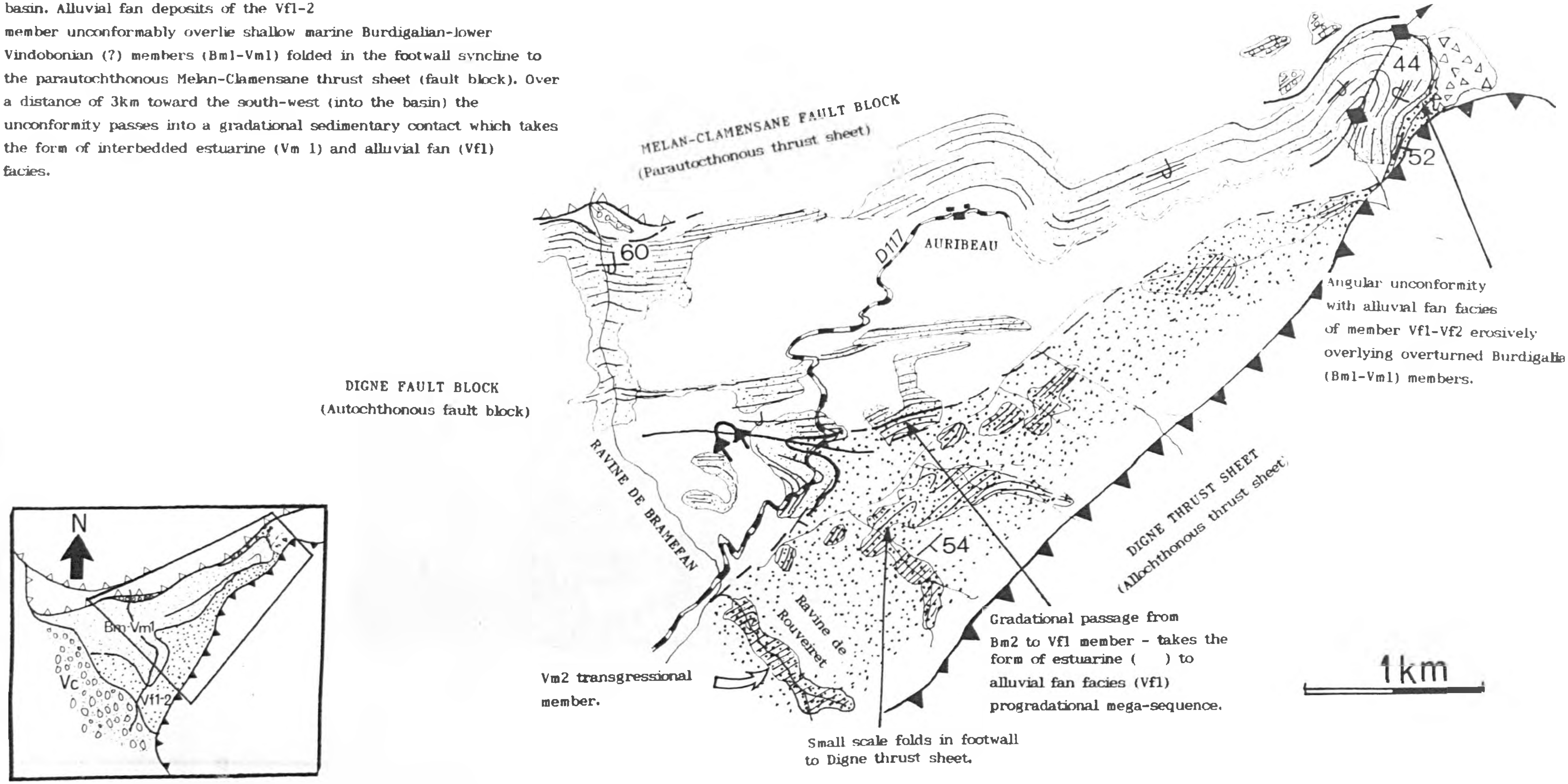


FIG 5.7 Vertically stacked coarsening-upward sequences which pass from offshore muddy sandstones (MSF) to inshore tidal channel facies sandstone bodies (TC), within the Bm1 and Bm2 members at Sourribes.

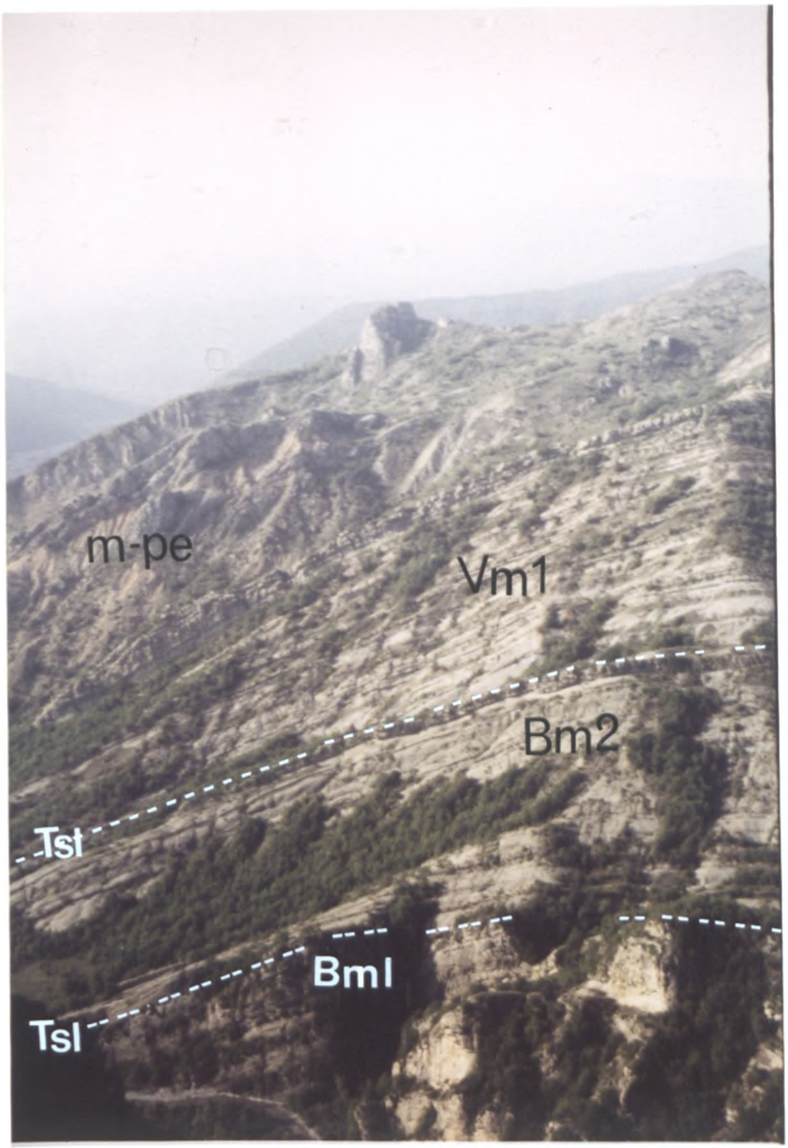


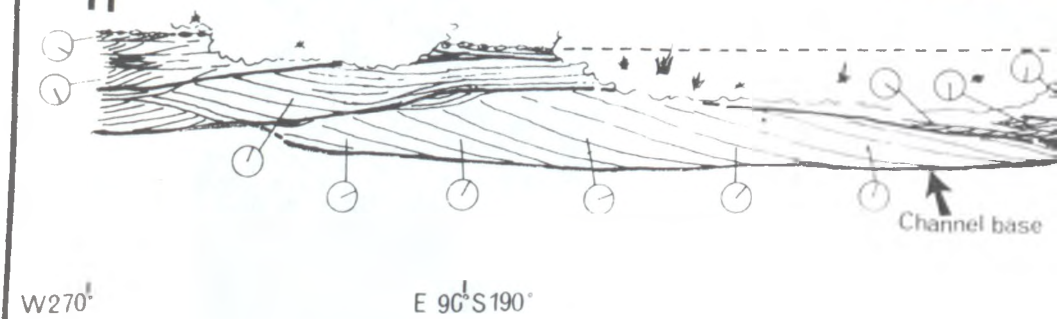
FIG 5.8 Upper part of the Marine Molasse Formation (members Bm1b- Vm1) as exposed at Esclangon.
Member Bm1 - Massive sandstones of the Shoreface-Foreshore facies association at the top of the member. **Member Bm2** - Base of member marked by transgressive shoreface lag facies (TSL) above which offshore mudstones with glauconitic sansdtone units develop. Aggradation of muddy sandstone offshore facies containing laterally extensive, m scale thickness, distal tidal shoal sandstone units. Top of member marked by a laterally extensive sandstone ridge comprising a multi-storey, inshore tidal channel complex (TC) (see also Fig. 5.   )
Member Vm1 - Base of member marked by transgressive shoreface lag facies (TSL) above which offshore mudstones develop. Subsequent progradational mega-sequence with passage from offshore and distal fluvio-distributary/tidal sandstone units into proximal fluvio-distributary facies with brackish bay mudstone facies (see Fig 4.71 for detailed fence log).
In background (m-pE) is a chaotic assemblage of olistoliths (of Cretaceous limestone and Tertiary) and continental mudstones of restricted developement immeadiately beneath the Digne thrust sheet ("bassin residual" deposits of Gigot *et al.* 1974)

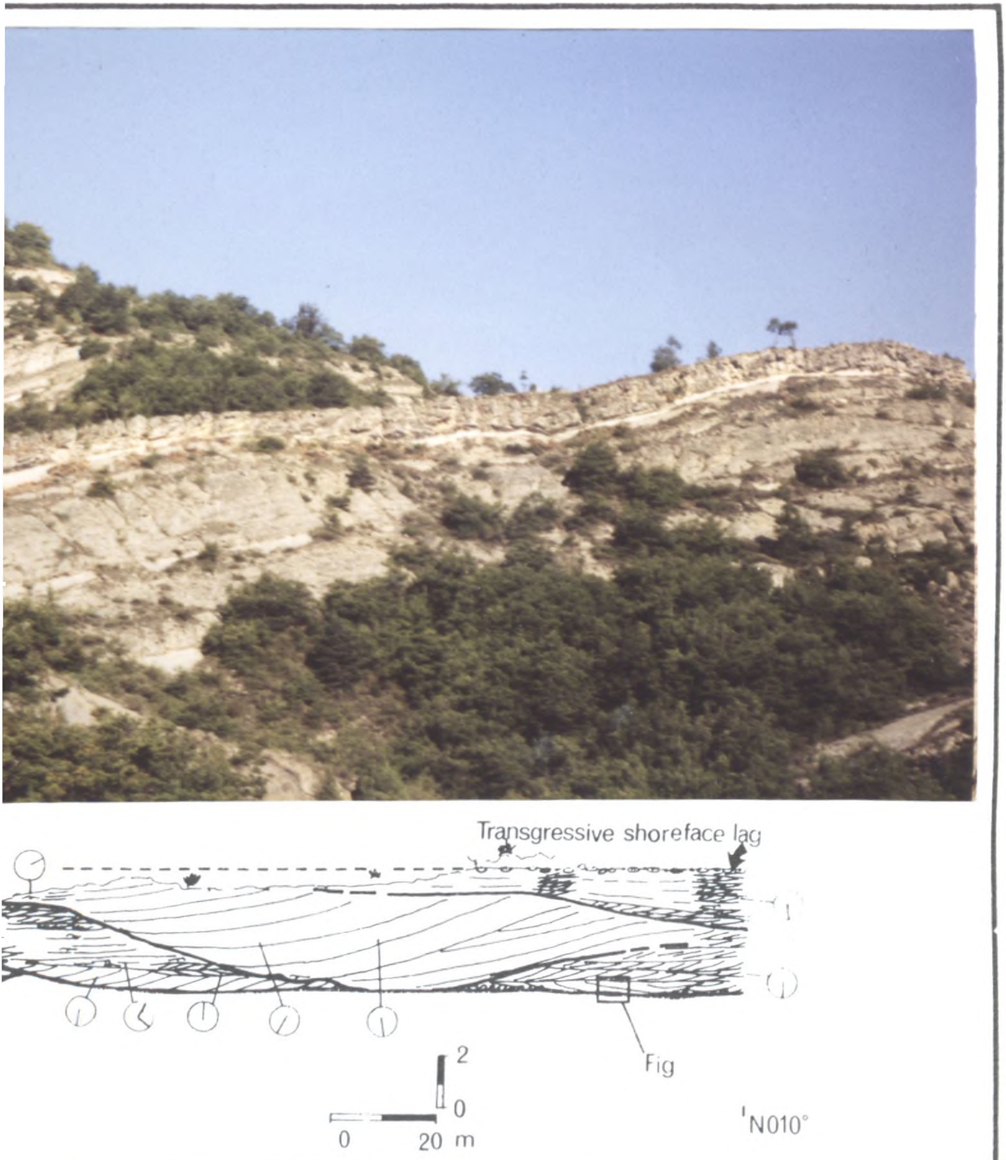
FIG 5.9 (i) Laterally continuous sandstone body of the inshore tidal channel facies erosively overlies muddy sandstones with tidal sandstone bars of the 'offshore' facies at the top of the Bm1 member, Esclangon. The top of the sandstone unit is a transgressive shoreface-erosion surface lagged with bored cobbles marking the base of member Vm1.

FIG 9 (ii) Sketch showing the internal archıtechture of the inshore tidal channel complex at the top of the Bm1 member, Esclangon. Internally, fining-upward sequences from large scale (2-5m) trough cross- stratification into dm scale cross-stratification and ripple lamination are developed above vertically and laterally offset channel erosion surfaces and interpreted to be the fills of laterally migrating inshore tidal channels. Note the bidirectionality of the large and small scale stratification. The vertical scale is exaggerated.

i

ii




Transgressive shoreface lag
2
0
0
20 m
Fig
N010°

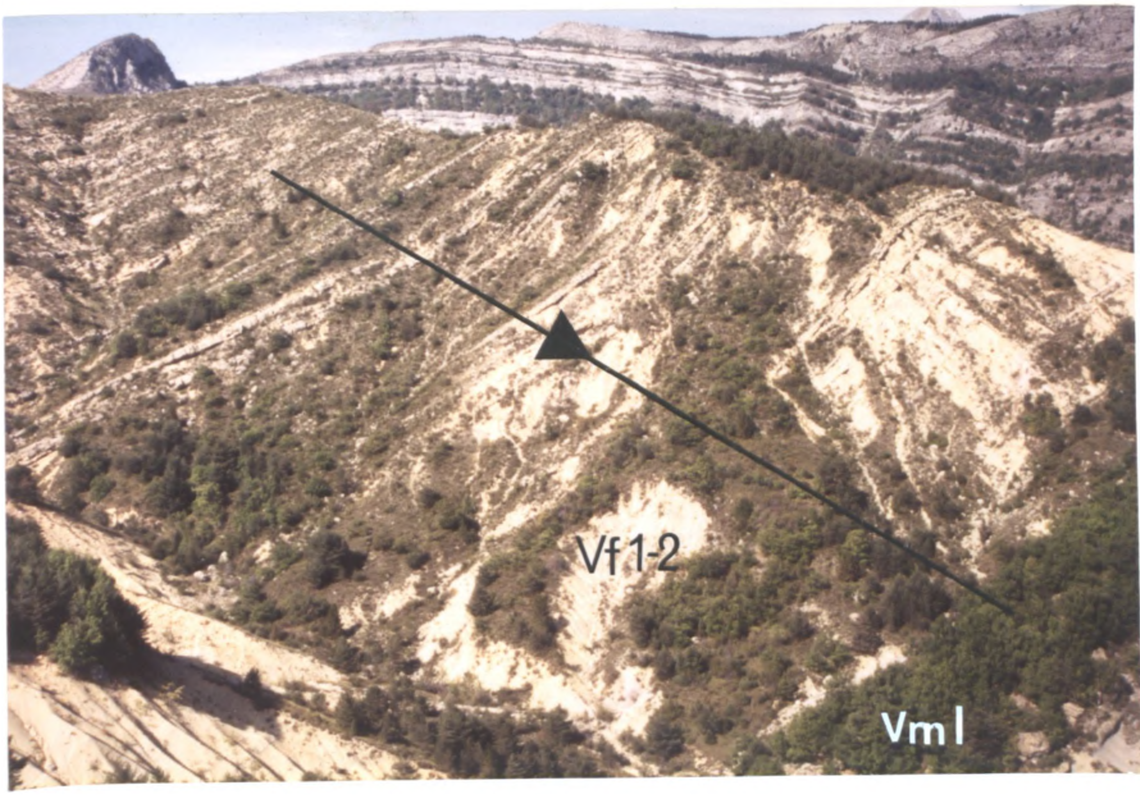


FIG 5.10 Coarsening-upward megasequence of locally sourced, alluvial fan facies of the upper Vindobonian, Vf1-2 member at Esclangon. The mega-sequence coarsens upward above an abrupt but non-erosional contact with mixed carbonate and siliciclastic fluvio- distributary facies of member Vm1.

FIG 5.11 Upper estuarine tidal channel and flat facies of the Vm2 member conformably overlie reddish mudstone and sheet sansdtone facies of the Vf1 member, at Ravine de Rouveiret, Auribeau) (see also Fig 5.6).

FIG 5.12 General view of the upper transgressional-progradational cycle of the Marine Molasse formation and overlying Valensole Conglomerate formation at St.Symphorien. Flood delta and upper estaurine facies associations of the Vm2 member are gradationally overlain by coastal alluvial plain facies of the Vf2 member. The cliff face in the background comprises massive, to horizontally stratified, braided conglomerate facies of the Valensole Conglomerate formation. Palaeoflow in the conglomerates is towards the right (south)

FIG 5.13 Close-up view of the Valensole Conglomerate formation showing that it predominantly comprises massive and horizontally stratified conglomerates forming sheet like units of 5-20m thickness seperated by metre scale intervals of pedogenically mottled mudstone (m).

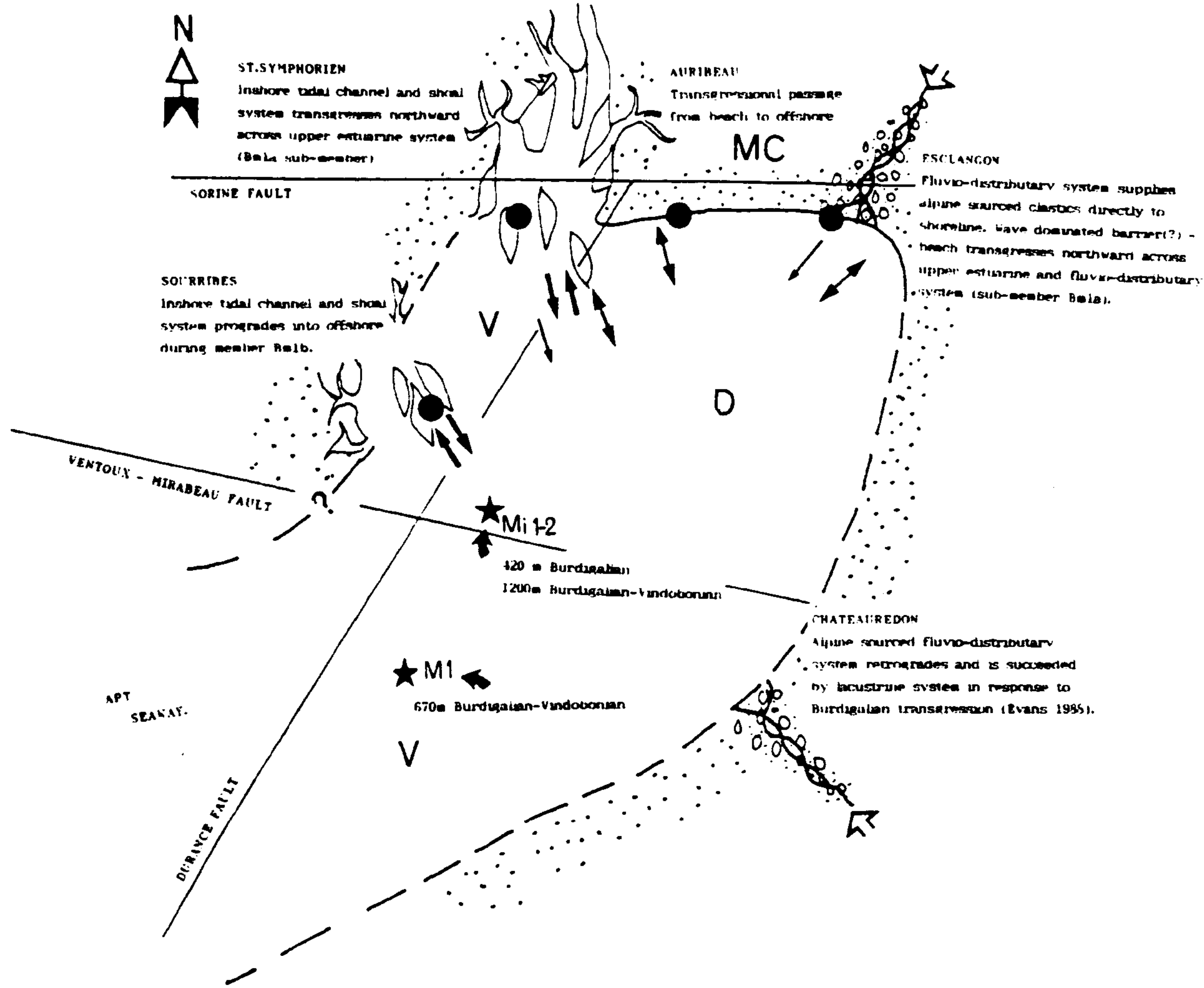


FIG 5.14 Dynamic palaeogeography of the Digne-Valensole Gulf during the lower Burdigalian - Member Bm1. Mixed tide and wave influenced siliciclastic shoreline systems.

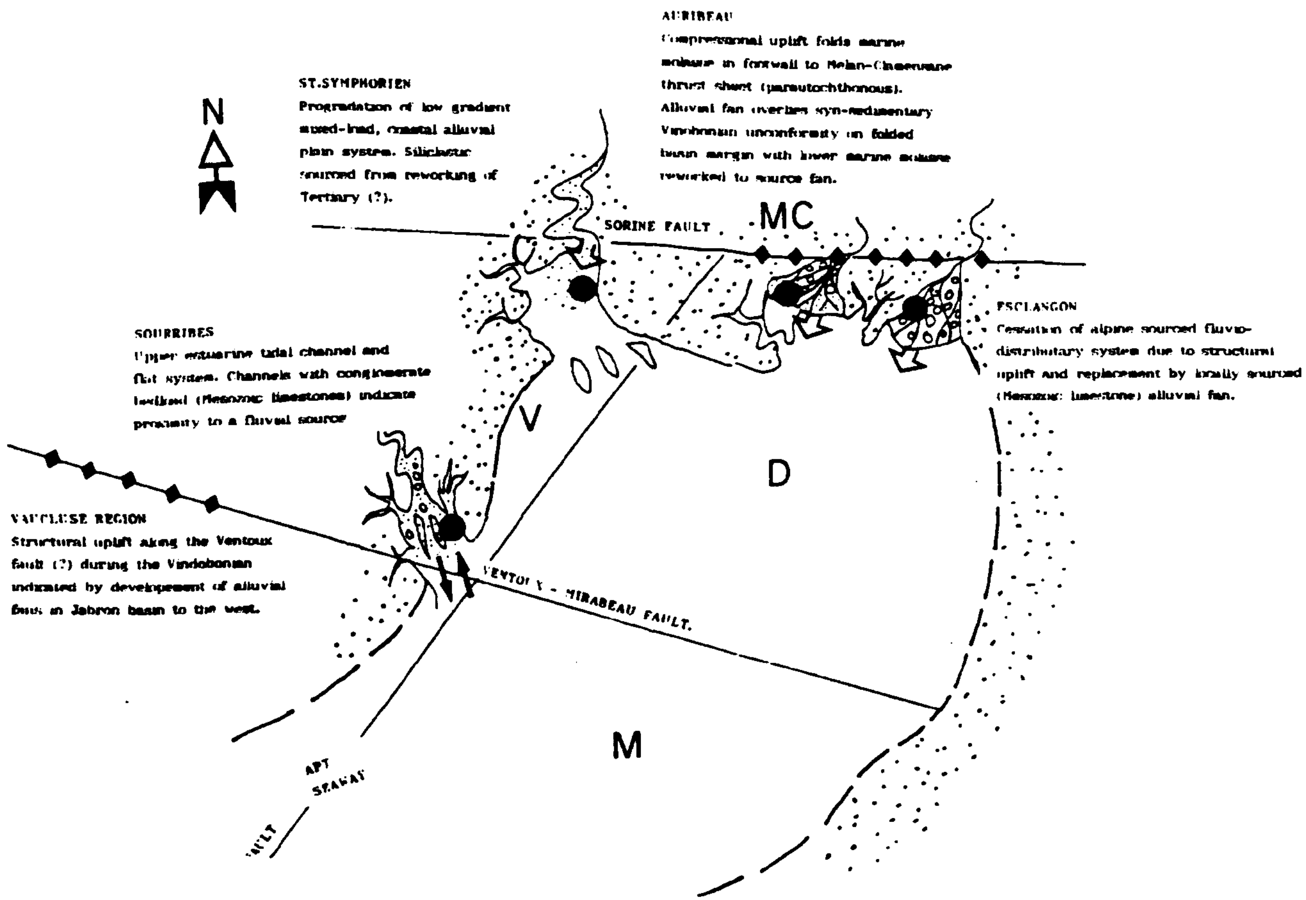


FIG 5. 15 Dynamic palaeogeography of the Digne-Valensole Gulf during the late Vindobonian - Member Vf1-2.

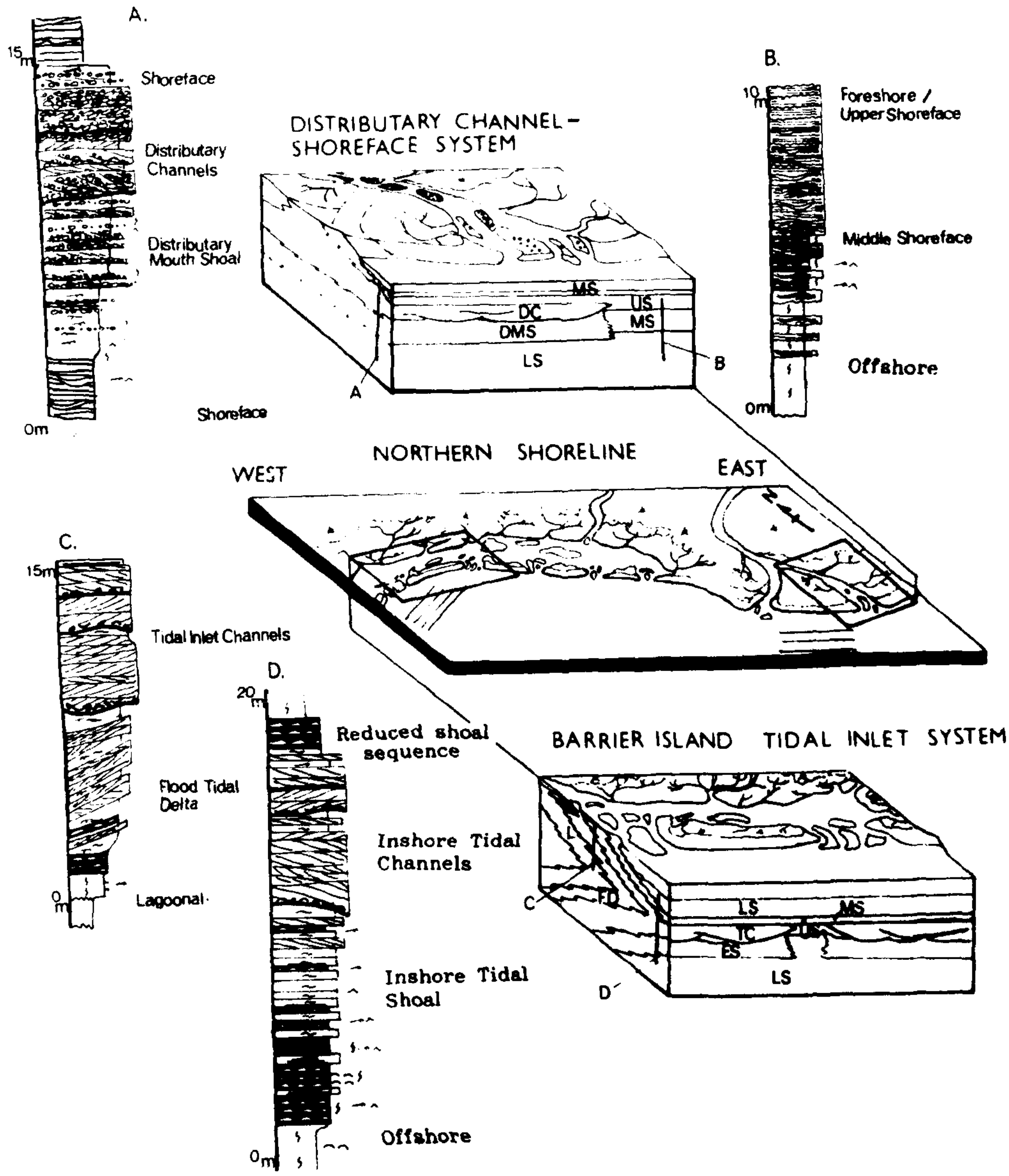


FIG 5.16 Schematic reconstruction of the lateral relationship of depositional systems within the Digne-Valensole Gulf during the Burdigalian-Vindobonian as interpreted from facies sequences of the Marine Molasse formation.

A: Foreshore - Fluvio-distributary facies association developed in a transgressional mega-sequence (sub-member Bm1a)

B: Offshore - Foreshore progradational sequence (base of sub-member Bm1b)

C: Flood delta coarsening-upward sequence attributed to barrier transgression (Vm2 member)

D: Offshore to inshore tidal channel and shoal coarsening upward sequence attributed to progradation or lateral migration of inshore system.

FIG 5.17 Photograph showing the abrupt sedimentary contact of the lower (Bm1J) and upper (Bm2J) Burdigalian members at Montbrun (1km east of Montbrun). F: Fault. Abrupt change in lithofacies from massive grainstone sandwave facies (Bm1J) to siliciclastic siltstone and mudstone (Bm2) interpreted to record major change in marine basins geometry (see text).

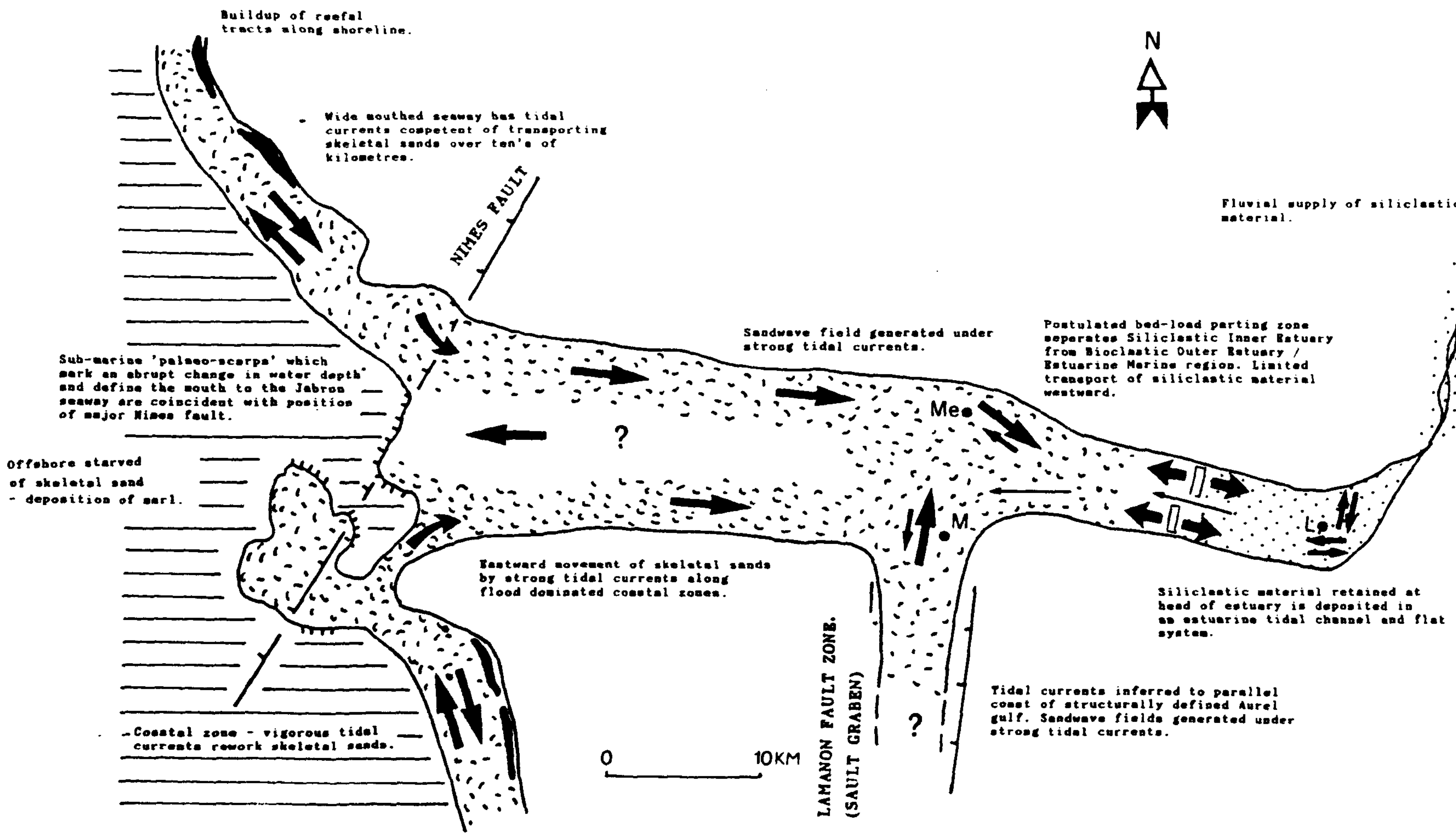


FIG 5.18 Evolution of the Miocene Jabron Basin

Speculative reconstruction of the basin during the lower Burdigalian integrating facies analysis of localities in Jabron basin with lithofacies of the Rhone valley adapted from Demarq *et al.* 1984. Basin interpreted to have been occupied by elongate gulf which extended eastward from the Rhone seaway, with closed estuarine system at head of gulf passing westward into open shelf system starved of siliclastic material.

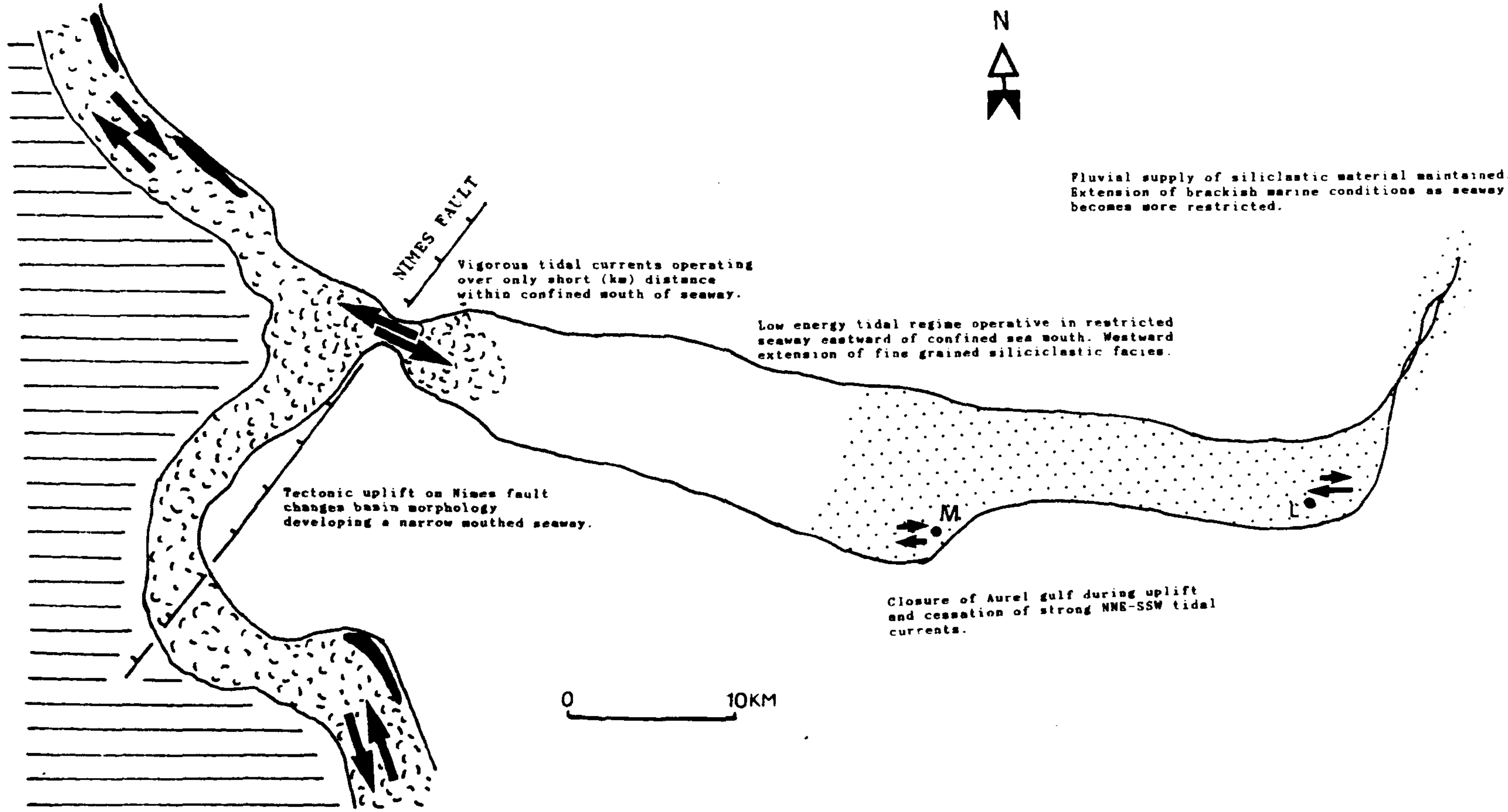


FIG 5.19 Speculative reconstruction of Jabron Basin during the upper Burdigalian. Tectonic uplift along Nimes fault is suggested to have resulted in confinement of Jabron Basin.

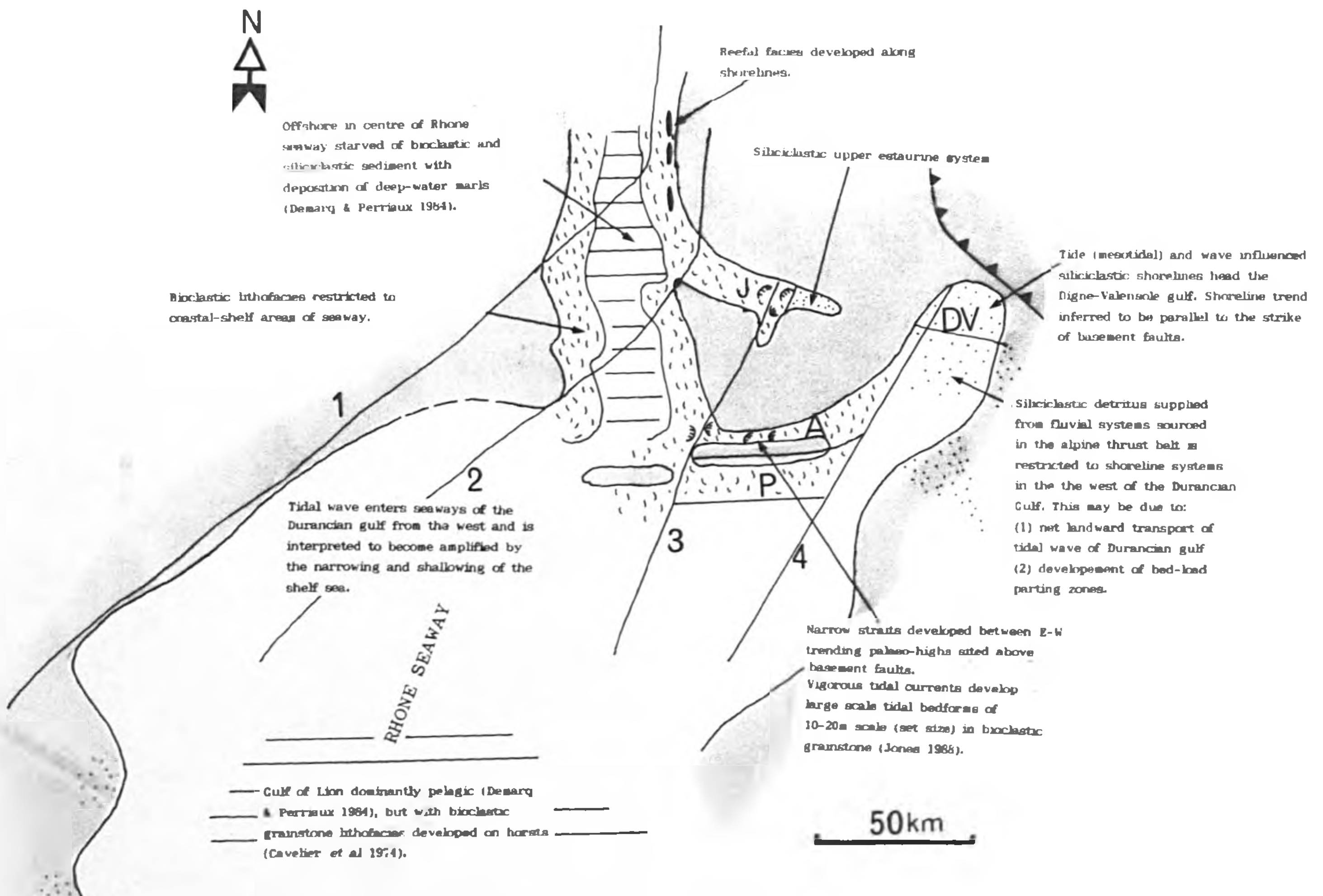


FIG 5.20 Dynamic palaeogeography of the Durancian gulf and Rhone seaway in the Miocene.

1: Cevennes fault 3: Lamanon fault
2: Nimes fault 4: Durance fault
A: Apt seaway P: Pertuis seaway J:Jabron seaway
DV: Digne-Valensole gulf

CHAPTER 6

BASIN EVOLUTION

6.1 Introduction.

Tertiary basins of the external zones of the S.W. Alps have been interpreted by Elliott *et al.*, (1985) and Evans (1988) as the remnants of a foreland basin. In contrast, adjacent and coeval basins in the Rhone and Provence regions have been interpreted as extensional basins of the Western Mediterranean rift system (Steckler & Watts 1980, Cavelier *et al.*, 1984, Bessis 1986).

In the first part of this chapter the characteristics of foreland and extensional basins, and, the evidence for their application to the Tertiary basins of S.E France will be discussed. In the later part of the chapter the characteristics of the Durancian basins will be summarised and compared and contrasted to those of the foreland and extensional basins.

6.2 Foreland Basins - A review of their origin and characteristic features.

Foreland basins form at the site of lithosphere flexure in response to passive loading by supra-lithospheric mass loads superimposed during formation of a thrust belt (Beaumont 1981). The foreland basin is mechanically coupled to the adjacent orogen, with its amplitude and wavelength being dependent on the timing and amount of mass movements in the orogen, the thermo-mechanical properties of the underlying lithosphere, and the amount of sedimentary infill of the basin.

Foreland basins have a number of characteristic features which aid their interpretation (see Beaumont 1982, and Allen *et al.*, 1986 for detailed reviews), namely:

(1) Foreland basins are asymmetric, characteristically having a convex upward flexural shape with maximum subsidence adjacent to the applied load, decreasing forelandward to a peripheral forebulge (Beaumont 1981, Jordan 1981).

(2) Foreland basins migrate in the same direction as their associated thrust belt. Their migration rates are dependent upon

the lateral transfer of mass by the orogen as a whole, including subsurface high density loads, and not simply to the rate of migration of the thrust front (Karner & Watts 1983).

(3) Flexural subsidence migrates across the foreland, with any point in the foreland recording an increasing rate of tectonic subsidence with time. As a consequence, foreland basins are characterised by convex-upward subsidence curves *eg.* the Swiss Molasse basin (Homewood *et al.*, 1986).

(4) Foreland basin fills commonly show an evolution to their sediment fill, with the oldest deposits being fine grained turbidites, and the later deposits predominantly coarse grained, shallow water or continental deposits (molasse) (Allen *et al.*, 1986).

(5) Foreland basin fills show an intimate link with thrusting. In the simplest context, prograding clastic wedges in the basin may be related to thrust events in the hinterland (Wiltschko & Dorr 1983, Elliott et al 1985). A more intimate linkage between the basins and the thrust system is shown by syn-tectonic structures, such as unconformaties and growth folds at inner basin margins.

6.3.1 Evidence for a foreland basin interpretation of the Tertiary basins of Provence - The S.W. Alpine Foreland Basin.

The similarities between the Tertiary basin remnants of Provence and foreland basins are recognised and discussed in investigative studies by Elliott *et al.*,(1985), Apps (1987), and Evans (1987). The basins are interpreted to be the remnant fills of a foreland basin which migrated southwestward in response to flexural loading by the alpine thrust belt (see Fig. 1.2).

The basins show all of the diagnostic features of foreland basins, namely:

(1) the basins are developed adjacent to an orogenic belt, the S.W. Alps, a characteristic of peripheral foreland basins (Dickinson 1974).

(2) the depocentre of the basin migrated foreland-ward (Fig. 1.10), from the Gres d'Annot basins (late Eocene), to the Barreme basin (late Eocene to Oligocene) and subsequently the Digne-Valensole basin (the north-eastern margin of the Durancian basin) (early Oligocene to Pleistocene)

(3) an as ymmetry to the basin is demonstrable with older and thicker successions in the basin remnants to the east progressively thinning and younging to the west.
(4) a foreland-ward passage, from older, turbiditic basins (Gres d'Annot) to younger, continental and shallow marine basins (Barreme and Digne-Valensole basins).
(5) syn-sedimentary deformation can be demonstrated in the Gres d'Annot and Barreme basins.

The construction of a structural section across the external S.W Alps by Graham (in Elliot 1985) identified the key point that the thrust system propagated into the foreland ahead of the basins, with the Barreme and Gres d'Annot basins developing as thrust sheet top basins. A rapid shift in the position of the depocentre, to the autochthonous Digne-Valensole basin (northern-eastern margin of the Durancian basin), is considered to have occurred at the beginning of the Burdigalian (Evans 1988). Using this data Evans 1988) calculated the depocentre migration rate of the basin (time averaged) to have been 2.3mm yr^{-1}.

6.3.2 **Problems with the foreland basin model - A discussion.**

The first problem with the model is that when the Durancian basin as a whole is taken into account, the onset of Tertiary sedimentation across the foreland basins is essentially simultaneous, whilst it is the termination of sedimentation in these basins which shows the more progressive temporal trend.

Perhaps more importantly the model, as a fieldwork based "working-model", was unable to address the problem of the role played by the inherited lithospheric structure. The alpine thrust belt and foreland basin is developed upon a Mesozoic rift margin, which retains pronounced spatial variations in its effective elastic (lithospheric) thickness, and a system of major extensional faults (see Chapter 2). This creates two problems.

The first is the spatial variation in the effective elastic thickness of the lithosphere, with Menard (in Lorenz 1980) suggesting that thinned lithosphere persists, even today, beneath the sites of the inherited Mesozoic basins. The effective change in flexural rigidity that this caused during the Tertiary evolution of the region is debatable as lithospheric strength is a function of geothermal gradient, as well as crustal thickness (Kuznir and Karner 1985, Kuznir & Park 1985, Kuznir and Mathews. (in press)). This problem is pursued no further as it is felt to be beyond the scope of this project, and the present day knowledge of the crustal structure of S.E France.

The second problem is regarding the control played by inherited extensional faults. A 'recent' advance in structural geology has been an appreciation of the role that pre-existing basement structures play in orogenic and extensional rift belts. It is now widely recognised that much intracontinental deformation involves the reactivation of existing discontinuities rather than the creation of new faults (McKenzie 1972, Jackson 1980). This is especially true where a transition from passive to active (compressional), margin occurs, (Cohen 1982). The model of Elliott *et al.* (1985), fails to account for pre-inherited faults, even though studies in the French External Alps have revealed their importance in controlling the orientation and geometry of the thrust sheets (Siddans 1975, Davies 1981, Tricart & Lemoine 1986, Gillchrist *et al.*, in press).

In terms of this study it is important to note these inherited faults can be demonstrsted to have controlled the geometry of the frontal alpine thrust, the Digne thrust sheet. The idea of Siddans (1979) that a high angle basement fault, the Barcilonette fault, acted as a frontal ramp to the Digne thrust, was extended in section 2.4 (Fig. 2.9) to show that the Digne thrust is an inverted Mesozoic 'trap door' graben. In the context of this control by basement faults it is furthermore apparent that the southwestward direction of alpine thrusting is parallel to the strike of the major Mesozoic basement faults and it seems reasonable to infer that these these faults acted as major lateral structures orientating the compressional deformation direction and tramlining the thrust sheets.

Further support for this was found in the southeastern margin of the inverted Digne graben which is coincident with the lateral ramp of the Sapee-Chiran-Beynes imbricate stack, and that the Dome de Remollon hanging-wall lateral ramp fold which is coincident with the Clamensane basement fault (see Fig. 2.9).

Summary

To briefly summarise the foreland basin model of Elliott *et al.*, and Evans (1988) clearly demonstrates that the remnant S.W alpine basins formed a part of a larger foreland basin. Furthermore the trend, and rate of migration of the foreland basin indicates that it should occupy the Digne-Valensole basin (north-eastern Durancian basin) during the Miocene.

However, the work of this thesis suggests that the failure to account for the role played by inherited structures, makes it difficult to directly apply the model of Elliott *et al.*, (1985) to the Durancian basins.

6.4 Extensional Basins - A review of their origin and characteristics.

Extensional rift basins are associated with divergent plate boundaries where extension is dominant (Watts *et al.*,1982). Extensional basin development can be considered in terms of two end member models, namely the McKenzie (1978) model of homogeneous lithospheric stretching and the Wernicke (1985) model of heterogeneous stretching.

Extensional basins typically have a two phase history of development with:

(a) an initial rift (active, or mechanical) phase developed as a consequence of the initial extension of the continental lithosphere, is characterised by rapid subsidence associated with normal faulting of the lithospheric crust.

(b) a thermal (passive, or post-rift phase) characterised by exponentially decreasing thermal subsidence.

The fault controlled subsidence of the rift phase is succeeded by a more widespread and homogeneous subsidence with the areas of greater extension subsiding more rapidly (see Sclater & Christie 1980). In the McKenzie model (Fig. 6.1) (McKenzie 1978, White & McKenzie 1988) this phase produces stratigraphic onlap at the flanks of the rift basin generating a 'steershead' basin geometry.

As a consequence of their two phase developement, extensional basins are characterised by a concave upward subsidence curve. This comprises an initial, steep section (rift phase), followed by an exponentially shallowing section (thermal phase).

6.5 Evidence for an extensional basin interpretation of the Tertiary basins - The Western Mediterranean Rift Systems.

Coeval with the development of the S.W. alpine foreland basin an extensional rift system, the Western Mediterranean System, was initiated in S.E France in response to a change in the convergence direction of Africa and Europe (Ziegler 1982, Burrus 1984).

A linked system of extensional grabens defined by reactivated, Mesozoic high-angle faults, extended northwards from the Provence Basin (Ligurian), as the Rhone, Bresse, Rhine and Ruhr grabens (Fig. 1.6) (Rehault *et al.*,1983, Ziegler 1982).

In the south of France four principal NE-SW trending faults, namely the Cevennes, Nimes, and Durance fault are interpreted to have divided the 'Rhone system' into elongate, regional fault blocks (Fig. 1.3)

These N.E - S.W faults are interpreted to have been linked, and or, offset, by E - W trending 'transfer zones', which reactivated Cretaceous (?) age extensional faults. A principal example is the Luberon fault which is interpreted to 'offset' the Camargue graben faults, and translate displacement onto the Durance fault system (Fig. 2.10). The principal rift grabens of the Rhone system, and isopachs of their late Eocene (Priabonian) - Oligocene fills are detailed in Fig. 2.10. Maximum basin fill thicknesses of some 1500-2000m are achieved in the rift axis - Camargue graben, and the more marginal Ales and Forcalquier grabens.

The grabens had a continental fill, characterised by the development of locally sourced alluvial fan sequences adjacent to the footwall block.

A two phase history of extension has been shown for the Western Mediterranean rift system (Fig. 1.8) (Steckler & Watts 1980, Burrus 1984, Bessis 1986) which may be applied to the Rhone system.

(1) **An active rift phase,** initiated in the Chattian (30-35my) and maintained until the Aquitanian (23-24 my). This phase was characterised by the extensional reactivation of Mesozoic age faults, to form a system of NE-SW trending, regional horsts and grabens (Fig 6.4 (i)).

Map studies show that the principal grabens were elongate features developed in the hanging wall of NE-SW trending faults. They were terminated at their northern and southern limits by subordinate E - W trending transfer faults.

(2) **A thermal subsidence phase.** The beginning of the Aquitanian marked the onset of passive thermal subsidence. The Gulf of Lions margin experienced strong subsidence associated with this thermal phase, with the Neogene, marine sequences onlapping westward toward the margins hingeline (Steckler & Watts 1975) giving a steers head basin (Dewey 1982) geometry (Fig. 6.1 - 6.2). Stratigraphic onlap is also demonstrable in the Rhone rift system.

The relationship of the onlap stratigraphy to tectonic subsidence is not, however, simple because the period of onlap was also one of a eustatic sea level rise (Vail and Harbendol 1979, Haq et al., 1987; see Fig. 4.1). However, the fact that marine conditions were established during this period, but not the Palaeocene when global sea level had been higher, suggests that the transgression and associated onlap were primarily due to tectonic thermal subsidence.

The maximum extent of the Miocene sea in the south of France (Serravalian period) superimposed upon an isopach map of the preserved Miocene (post-Aquitatian -pre Messinian) sequences and the principal extensional structural elements is shown in Fig. 6.4 (ii)

Two principal points are shown by the figure.

The first is that the structurally isolated basins of the Oligocene have become stratigraphically connected, a characteristic feature of extensional basins in their rift to thermal phase transition.

The second is that subsidence, whilst more extensive than in the Oligocene was by no means homogeneous. Sediments achieved maximum thicknesses of some 1000-3500m in the Gulf of Lions (Cravatte *et al.*, 1974), Camargue and Valreas, above the sites of NE - SW trending Oligocene basins. In contrast, the depocentres of the Durancian region show shifts in the positions of their maximum sediment fill, which developed in the footwall of the E - W trending faults.

Summary

An extensional basin system clearly developed in the Rhone region, and extended eastward to the alpine thrust front through the reactivation of Mesozoic, high angle faults. Extensional basins of this system should show a classic, two-phase extension history and a convex upward, subsidence curve.

6.6 The Tertiary Durancian Basins - Foreland, Extensional, and Strike-Slip Basins. A Tectono-Sedimentary Model of Hybrid Basins.

6.6.1 Introduction

The evolution of the Durancian basins will be considered in a series of sequential time steps which allow a breakdown of their multi-phase deformation history. These are:

(1) Late Eocene to early Miocene.

(2) Early to Late Miocene.

(3) Late Miocene to Pleistocene.

The data presented for each of these time periods will take the form of:

(i) the European, Tertiary palaeostress fields (Fig. 6.3 (i-iv) from Bergerat (1987).

(ii) isopach maps of the Tertiary remnant basins of S.E France (adapted from Debrand-Passard 1984) superimposed onto a structural map of the region (Fig. 6.4 (i-iii)).

(iii) a fence diagram of the Tertiary fill of the Durancian basin superimposed onto a structural map of the basins inherited fault framework (Fig. 6.5).

(iv) a sequentially restored section across the Vaucluse fault block in the north of the Durancian basin (Fig. 6.6 (i-iv)).

(v) a resume of the sedimentological evolution of the Durancian basins.

(vi) subsidence curves (Fig. 6.10). An attempt to derive the subsidence history of the northern Durancian basin has been made by decompacting the thickest and most complete Tertiary sequences of the basin, namely those of the Vaucluse fault block. This curve is compared with an uncompacted one from the Manosque graben (from Jones 1988). Decompaction was carried out using a simple computer programme (J. Marshall *pers. comm*). Tectonic subsidence was not differentiated from sediment load induced, subsidence, so that the curve gives only an indication of net subsidence. Loading by the water column is considered to have been negligible given the shallow marine conditions established during the Miocene.

6.6.2 **Late Eocene to Early Miocene.**

In the Durancian basins, the late Priabonian to Aquitanian period was marked by the deposition of continental sequences whose lateral stratigraphic thickness variations highlight a system of intra-basinal, trap-door grabens and horsts, defined by extensionally reactivated, NE-SW and E-W/WNW-ESE trending, high-angle faults (Fig. 6.4 (ii) - Fig. 6.5)
Regionally this was coincident with:
(1) the onset of active rifting in the Western Mediterranean rift system.
(2) out of sequence thrusting behind the 'pinned', alpine (Digne) thrust front.

Palaeostress reconstructions for the European platform (Fig. 6.3 (i) shows that during this period the region of Provence was being subjected to an E-W extensional regime.

In the north of the Durancian basin, grabens of St.Geniez, and Vaucluse grabens (Fig. 2.12) formed the northernmost part of a NE-SW trending regional half-graben, the Forcalquier Graben, developed on the hanging wall block of the major Durance fault (Fig. 6.4ii). This regional graben is dissected along its length by a series of active, E-W trending transfer faults which defined a series of discrete basins (trap-door grabens) namely the Aix, Manosque and Vaucluse basins (Fig. 3.61). Displacement on the Durance was apparently at a maximum in the region of Manosque, decreasing to the north and south suggesting that it was bound by an ellipsoidal zero displacement contour (see Barnett et al. 1987).

In the north of the Durancian basin maximum subsidence was in the Vaucluse, and St. Geniez grabens, whose northern and southern margins were defined by the Sorine, and Ventoux transfer fault zones respectively (Fig. 6.5 - 6.6ii). These are marked by a pronounced thinning of stratigraphy, and in the former case, by the extrusion of a remobilised Triassic diapir which was erosionally reworked by the Tertiary basin fill (Fig. 6.6 (ii) & 6.8)

Active subsidence on the Durance fault throughout the Oligocene is marked by the maintenance of a fault scarp fringing fan system (Vancon fan system) (Fig. 3.60–3.61), and the developement of a continuous fill in the grabens. Uplift on the footwall block to the immediate east of the fault (the Valensole horst) (Figs. 3,60 & 6.5) is indicated by the absence of pre-Burdigalian sequences (Mirabeau boreholes) and the erosion of the Mesozoic cover sequences down to the level of the Oxfordian, as opposed to lower Cretaceous within the grabens. This was a tilted block with a thick sequence of alpine sourced, fluvial deposits (Esclangon System) preserved along its downtilted eastern margin (Fig. 3.60). This block was dissected by 'E-W' transfer faults, in a similar fashion to the regional Forcalquier graben, as indicated by the developement of the Rancure basin (Fig. 6.5).

An important point to emphasise is the spatial variation in the fill of the Tertiary grabens of the Durancian basin (see section 3.11 for full discussion). Whilst the grabens are characteristically filled by terminal alluvial fan systems sourced from the Mesozoic carbonate sequences of the intra-basinal horsts (*eg* Vancon fan system), those grabens in the north of the Durancian basin are filled by siliciclastic sandstone supplied from the alpine thrust belt (*ie* Esclangon fan system) (see Fig. 6.5, and 3.60). The sediment supplied by the thrust belt was restricted to the immediate vicinity of the thrust front by extensional basins in the foreland which acted as sediment traps ("sinks") into which the alpine fluvial systems terminated.

Subsidence curves for the Vaucluse and Manosque basins show that both are characterised by the abrupt onset of high subsidence rates in the late Eocene (Fig. 6.10).

Compressional tectonics are not evident in the study area, or any of the Durancian basins during this period. The alpine thrust front was defined by the Digne thrust which is interpreted to have lain inactive, some 5–8 km to the NE of the Digne basin (see Fig. 3.60), as displacement was taken up behind the thrust front.

Summary

The Oligocene history of the Durancian basins is one of the development of extensional, tilted, fault blocks. The regional grabens were defined by NE - SW trending, reactivated, high angle faults, and were sub-divided into a series of sub-basins by reactivated E-W trending faults. Subsidence curves for these basins are characteristic of extensional basins. In terms of a tectono-sedimentary model, the basins show the characteristics of continental half-grabens with axial, and interior drainage (Leeder & Gawthorpe 1987). The principal role played by the compressional tectonics was in generating an erosional domain from which sediment was supplied to the extensional basins in the immediate vicinity of the thrust belt.

6.6.3 **Lower to Middle Miocene.**

The lower Miocene of the Durancian region was marked by the stratigraphic onlap of sequences on to Oligocene structural highs, and the consequent stratigraphic linkage of previously isolated basins (Fig. 6.4 (ii) & 6.5).

Considered in its regional context this change in the basins was also coincident with:

(1) the Burdigalian 'eustatic' sea level rise (see section 5.5 for discussion) and the establishment of shallow marine conditions in the basin.

(2) the onset of thermal subsidence throughout the Western Mediterranean Rift System.

(3) the cessation of a prolonged phase (late Chattian-Aquitanian) of out of sequence thrusting in the sub-alpine thrust belt which culminated in the structural uplift of the Barreme basin and the switching of the foreland depocentre to the thrust-front Digne-Valensole basin (Fig. 1.10).

(4) the onset of a NE-SW orientated stress field in the European platform (Fig. 6.3 (iii) & 6.4 (ii)) (Bergerat 1987).

In the study area active sedimentation occured across the fault block immediately to the east of the Durance fault (the Oligocene Valensole horst) linking basins separated by it. Thickness variations in the marine sequences of the Digne-Valensole basin (Fig. 6.5) show that it comprised a series of differentially subsiding sub-basins (depocentres), defined by transtensionally reactivated high-angle faults.

In the Digne-Valensole basin up to 900m of predominantly shallow marine clastic deposits were accumulated within the subsiding fault blocks. These form three basin-wide transgressional-progradational cycles with the transgressions tentatively attributed to thrust-load induced flexural subsidence. The volumetric dominance of progradational megasequences in the basin fills is interpreted to record the high rates of clastic sediment supply to the foreland from the adjacent alpine thrust belt.

An important contrast to the basins of the Oligocene is that the depocentres developed parallel to E - W trending, rather than NE - SW trending faults. In the Digne-Valensole basin depocentres developed adjacent to the Mirabeau and Sorine faults, whilst to the south west, they formed parallel to the E-W trending Luberon fault (R.Jones pers comm) (Fig. 6.5 & 6.6).

In the Digne-Valensole basin, subsidence was contemporaneous with compressional uplift along its northern margin. Here uplift in Vindobonian was marked by the developement of a syntectonic uncomformity (Col d'Ainac), overlain by Vindobonian fanglomerates containing reworked Miocene clasts (see section 5.3.2.4 & Fig 5.6). To the east, at Esclangon, a coeval alluvial fan succession records contemporaneous uplift of the Mesozoic on the north eastern margin of the basin. This uplift has been attributed to alpine compressional inversion of the Melan-Clamensane fault block along the high-angle Sorine fault zone, with Gigot *et al.* (1974) noting the developement of a higher level of thin skinned thrust folds in the Mesozoic cover.

Uplift in the forelandward propagating alpine thrust-fold belt to the east and north of the northern Durancian basins resulted in progressively increasing rates of sediment supply to the differentially subsiding autochthonous basin.

Compressional uplift was not restricted to the east of the alpine thrust front but also occured along the E - W trending, high-angle inherited faults in the foreland, namely the Luberon fault (R.Jones *pers. comm.*), and Ventoux fault (section 5.3.3.3).

The coeval development of extensional and compressional structures within an area (Fig. 6.4 ii) is characteristic of strike-slip deformation (Crowell 1975, Christie-Blick & Biddle 1985), and suggests a change in deformation style to that of the Oligocene. This change is coincident with one in the regional stress field with the onset of a NE-SW (56-256° in Provence) stress to the European plate in the early Miocene (Fig 6.4 (ii)).

A N50-230° strain field in the Digne thrust sheet during the Aquitanian phase (Gigot *et al.*, 1974), is attributed to an alpine compressional phase, but nevertheless it may have translated strain into the foreland.

It appears therefore that two simultaneous and parallel, NE-SW trending stress fields were established in the Provence region. However generated, such a field of stress would have subjected the NNE-SSW trending Durance fault to a strike-slip regime, whilst E-W /WSW-ENE faults would have been in a transpressional or transtensional regime. Along length changes in the strike of the Durance may have created 'restraining' and 'releasing' bends (after Crowell 1975). An example may be the offset of the Durance along the Hongrie fault creating the overlapping of the Digne, Vaucluse and Clamensane Blocks (Fig. 2.12 & 6.5). Uplift of the Clamensane block and northern margin of the Digne-Valensole basin along the Sorine fault at this restraining bend, was coeval with extension and subsidence of the Mees block on the Mirabeau fault some 20km to the south.

The translation of alpine compressional stress into the northern Durancian basin was not the sole contribution of the alpine belt to the evolution of the basin. The uplifting thrust belt also created an erosional domain from which siliciclastic material derived from the erosional re-working of earlier foreland basins (see section 5.4) was actively supplied to the subsiding strike-slip basins where it formed siliciclastic shoreline systems.

This is in complete contrast to the more western of the Durancian basins *eg* Jabron, and Apt basin (Jones 1988), and also the extensional basins of the Rhone graben and Gulf of Lions margin basins (Cravatte *et al.*, 1974) which are characterised by intra-basinally generated bioclastic grainstones (see Fig 5.20) in an open shelf sytem.

The subsidence curves for the Manosque basin (Fig. 6.10), show that subsidence was at a low rate relative to that in the

Oligocene giving the basin a concave upward 'apparent' subsidence curve, characteristic of extensional rift basins. In contrast, the curve from the Vaucluse graben shows the maintenance of high subsidence rates throughout the Miocene with the highest apparent rates being achieved during the late Miocene - Pliocene period. This maintenance of high subsidence rates during the Miocene in the Vaucluse, relative to the Manosque basin may reflect the additional effects of thrust sheet induced flexural subsidence, as well as higher rates of sediment supply, with proximity to the thrust front.

Summary

The Miocene was marked by a continuation of differential subsidence along re-activated, high-angle faults. A change from an extensional to transtensional regime is marked by a change in the geometry of basins, and by the coeval development of E-W trending, pull-apart basins and compressional highs, along the length of the regional Durance fault. This is attributed to the translation of alpine compressional deformation into the foreland reactivating inherited high-angle faults. The alpine thrust belt further contributed in supplying the siliciclastic fill to the northernmost Durancian basins. In comparison to the Oligocene, the alpine sourced siliciclastic detritus was restricted to the east of the foreland basin, but during this period was 'stored' within a shoreline-system.

The development of basin-wide transgressive events in the Digne basin (see Chapter 5) is interpreted to record spasmodic, thrust induced loading of the foreland. Further evidence for the onset of foreland flexure is provided by the steep subsidence curve of the Digne basin.

6.6.4 Late Miocene to Pliocene.

The Messinian was marked by a eustatic sea level fall (Haq *et al.*, 1987), and in the Durancian basins, by another major change in the basinal geometry and sedimentation patterns. For the first time during its Mesozoic and Tertiary history the Valensole block acted as a regional depocentre as the Durance fault changed its polarity (Fig. 6. 4iii).

Continental conditions, in the form of a braided river system were established across the Durancian basin. Deposition was confined to the Digne-Valensole basin which developed on the Valensole block to the east of the Durance fault and to the west of the alpine thrust front (Figs. 6.4 (iii) & 6.5). Sediment was actively supplied to the subsiding basin from the uplifting alpine chain. Maximum subsidence occured parallel to, and immediately to the east of the Durance fault (Fig. 6.4 (iii)). Subsidence of the Valensole block (Fig. 2.12) to the immeadiate east of the Durance fault was accompanied by tectonic uplift of the ManosqueForcalquier block to its west with the compressional inversion of the E-W trending, high-angle, Luberon, and Ventoux faults (Fig. 6.4 (iii)). Horizontal displacement on these was minimal, with inversion being primarily achieved by vertical uplift along the high angle reverse faults.

Differential subsidence within the Valensole block, is highlighted by the palaeoflow directions of the braided river deposits of the Valensole conglomerate within the Digne-Valensole basin fill (Valensole Conglomerate) (Fig. 6.9 i-ii). These rivers paralled the alpine thrust front, or traversed it at 'structural re-entrants' (Eisbacher *et al.* 1978) before entering the Digne-Valensole basin where they flowed parallel to the strike of the basement faults, a situation such as envisaged for the Oligcene fluvial system (see figure 3.60).

During the late Pliocene the Digne thrust sheet (alpine thrust front) overthrust the eastern margin of the Digne Valensole basin, travelling across a palaeo-land surface (Gigot *et al.*, 1974). North of Digne (Esclangon), alluvial fanglomerates and olistoliths, were shed westward off the emergent thrust, and unconformably overlie Oligocene and Miocene sequences folded in the footwall of the thrust sheet (basin residual deposits of Gigot *et al.*, (1974) (see Fig 5.8).

Compressional displacement subsequently shifted out into the autochthonous foreland. In the region of Mirabeau - les Mees a series of WNW-ESE trending, southward verging tip folds (Fig 1.2 (ii)) mark the uplift of the southern margin of the Digne block by positive inversion of the Mirabeau basement faults (see Fig. 2.13). This was accompanied by sinistral, transpressional deformation on the southern margin of the block with the development of a series of WSW - ENE striking en-echelon folds

(Figs. 1.2 (ii), 2.12 & 6.6).

Continued uplift on the northern margin of the Digne-Valensole basin during the Messinian-Pliocene resulted in the overthrusting and folding of the Valensole conglomerate in the footwall to the Clamensane sheet, and the positive inversion of the St. Geniez block along the Sorine fault zone reactivating the Sisteron anticline (Fig. 6.6i).

Summary

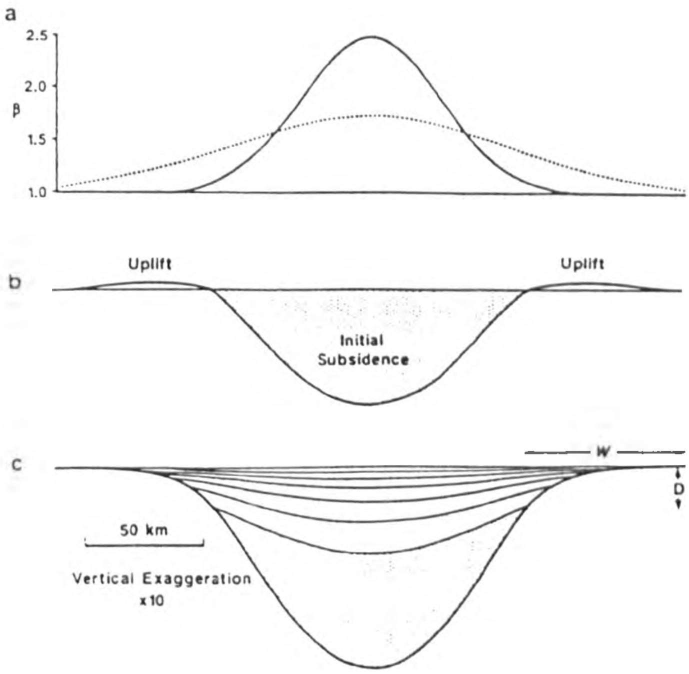
This Messinian-Pliocene period was characterised by the progressive change from a transtensional to transpressional regime in the north of the Durancian basin.

The subsidence of the Valensole block for the first time in its history suggests that this inherently buoyant block was subjected to thrust induced loading. This is supported by the high subsidence rates derived from a subsidence curve from the basin (Fig. 6.10). The major increase in thickness of the Messinian-Pliocene to the east of the Durance fault, intuitively suggests that it was reactivted as a strain decoupler to the alpine flexural loading of the foreland. However it is clear that the fault also acted as a major alpine strike-slip fault, partitioning a region of transtensional subsidence to the east, from one of transpressional uplift to the west.

Present Day

As a final point to the tectono-sedimentary evolution of the area, it is interesting to note that the present day drainage pattern in the Digne-Valensole basin is similar to that of the Pliocene and continues to reflect active fault controlled, differential subsidence (Fig. 6.9).

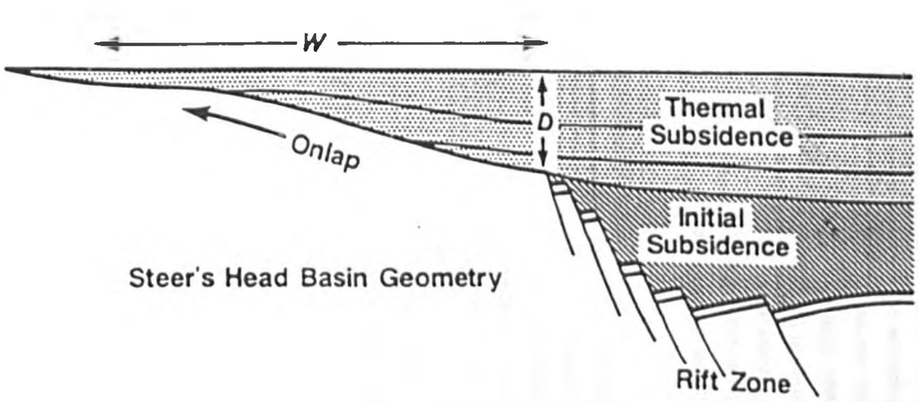
I



a: β_c, stretching factor in crust (solid line), and β_m, stretching factor in upper mantle (dotted line) are plotted as function of distance using equations 1 and 2 (see text); $\sigma_m = 60$ and $\sigma_c = 30$, as in Figure 2. In center of basin, $\beta_c = 2.5$ as before, but now $\beta_m = 1.75$. All other parameters are identical to those used in Figure 2. b: Initial subsidence immediately after stretching. Note localized uplift along flanks of basin. c: Total subsidence 150 m.y. after rifting. Thermal subsidence is represented by lines drawn every 25 m.y. Onlap produced by thermal subsidence at basin margin results in steer's head geometry. *W* is horizontal extent of onlap and *D* is its maximum thickness.

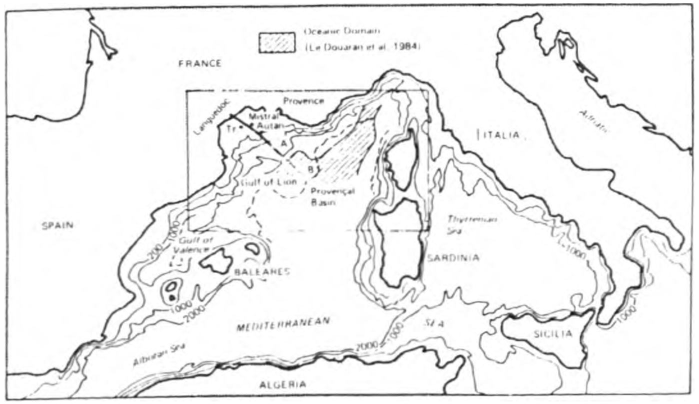
(i) Differential stretching of crust and upper mantle (a) result in generation of a steer's head basin with rift (b) and thermal subsidence (c) phases.

II



(ii) Idealised steer's head basin geometry showing postrift stratigraphic onlap at basin margin. Note the close similarity of this idealised geometry with that of the Gulf of Lions margin detailed in Fig 6.2 (i) which forms the western margin to the Ligurian sea of the Western Mediterranean rift system.

FIG 6.1 (i-ii) Steer's head basins (from White & McKenzie 1988)



(i) Cross-section of the Gulf of Lions rift margin showing that syn-rift (Palaeogene) sequences are restricted to grabens, whilst thermal rift phase sediments (Miocene) onlap, and progressively fill horst topography to form a laterally continous blanket of cover sequences. An erosion surface (er) seperates the Miocene and Pliocene sequences, generated during the Messinian lowstand.

FIG 6.2 (ii) Subsidence curves derived from back-stripping a series of sections across the Gulf of Lions margin. Within each graph the amount of Messinian erosion is indicated (hatched lines). (from Bessis 1986). Note the the, very different curves, particulary the initial rift-phase of the curve, generated from the Autan graben (G) and adjacent Autan horst (H) sections.

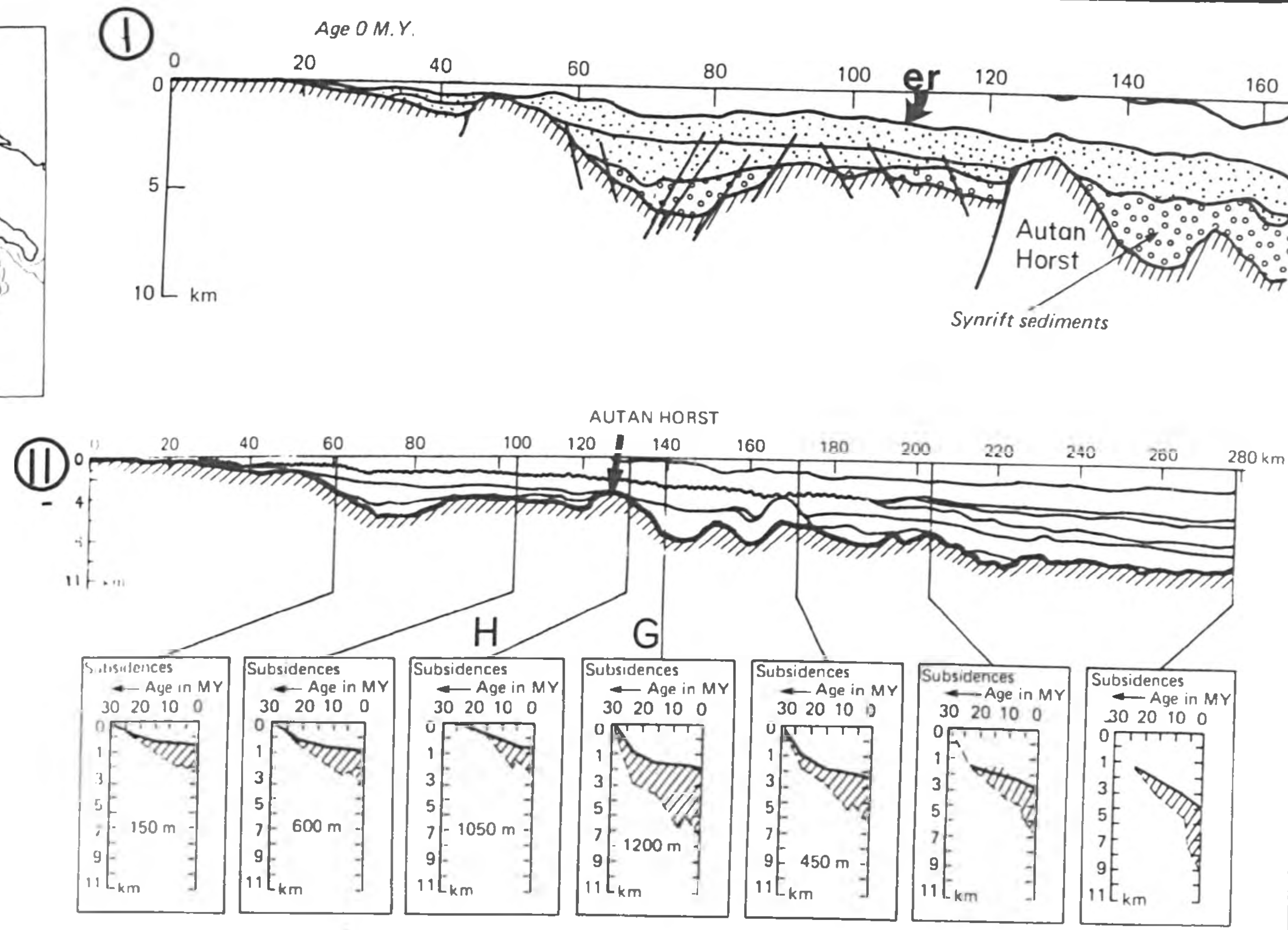
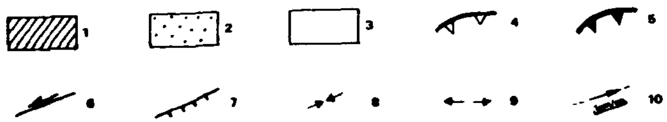


FIG 6.2 (i-ii) Gulf of Lions margin



1: oceanic crust
2: thinned continental crust
3: continental crust
4: subduction
5: overthrust

6: strike-slip fault
7: normal fault
8: main maximum stress
9: main minimum stress
10: vector motion (cm/yr)

IB: Iberian/Eurasian rotation pole.
AF: African/Eurasian rotation pole.
D-V Digne-Valensole basin

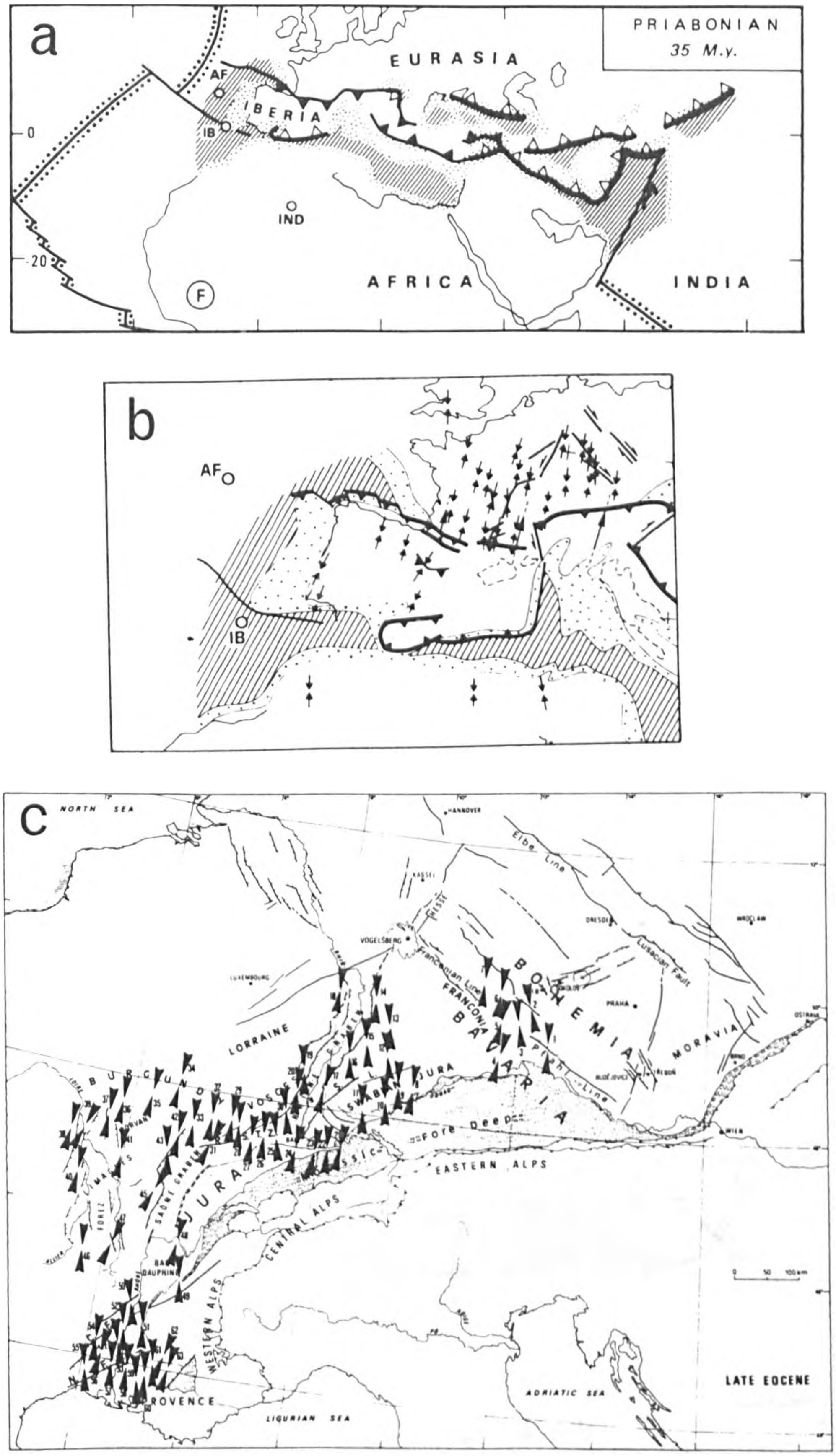


FIG 6.3 (i) N-S stress field in Provence associated with the northern Pyrenean (Pyrenean-Provencal) deformation.

FIG 6.3 (i-iv) Tertiary geodynamic evolution and palaeo-stress fields of the European Platform for (i) Late Eocene (ii) Oligocene (iii) early Miocene (iv) late-post Miocene (from Bessis 1986)

The figures show (a) the African - Europen continental plate framework (b) European platform palaeostress fields (c) detailed analysis of the western European palaeo-stress fields

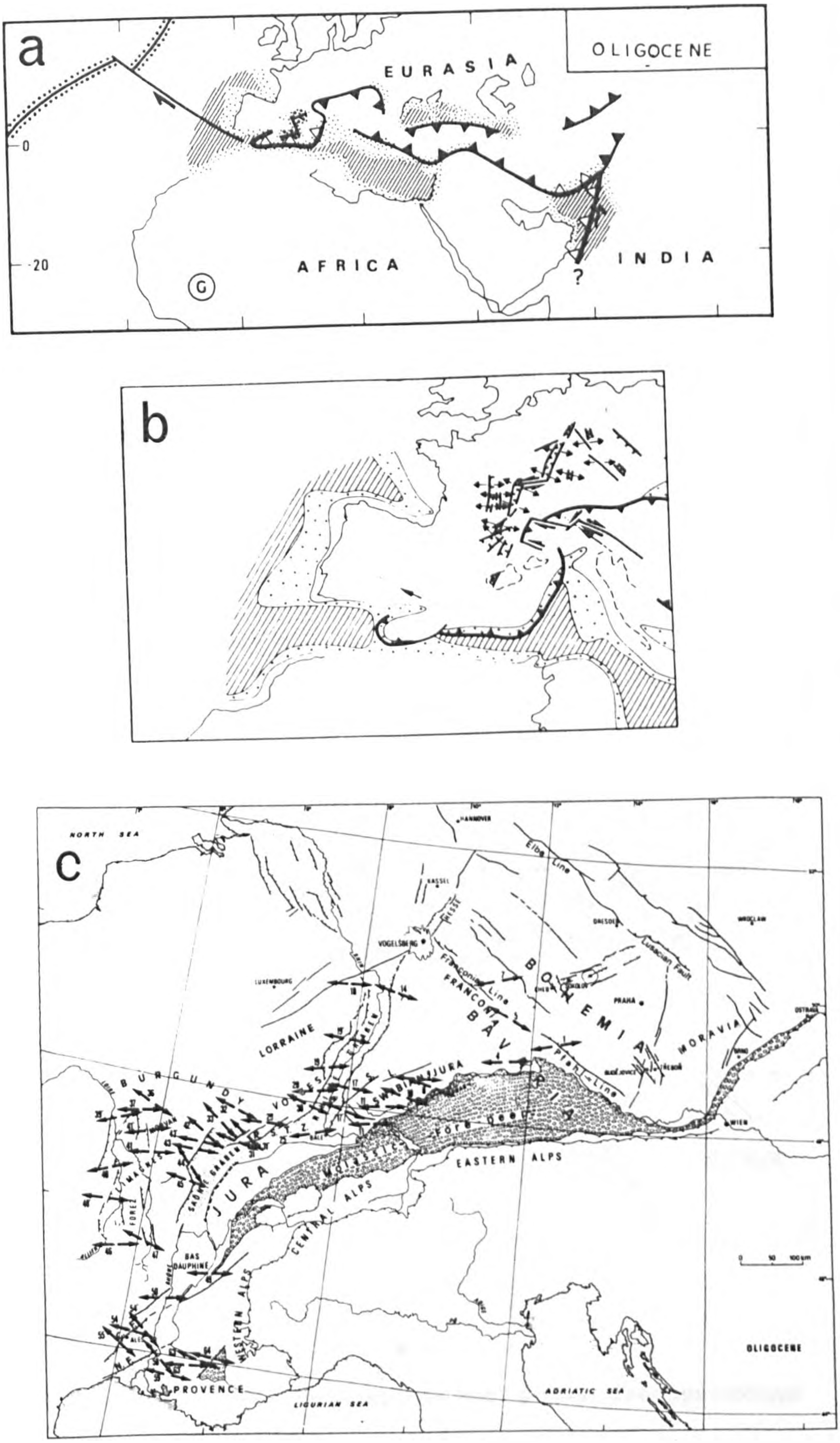
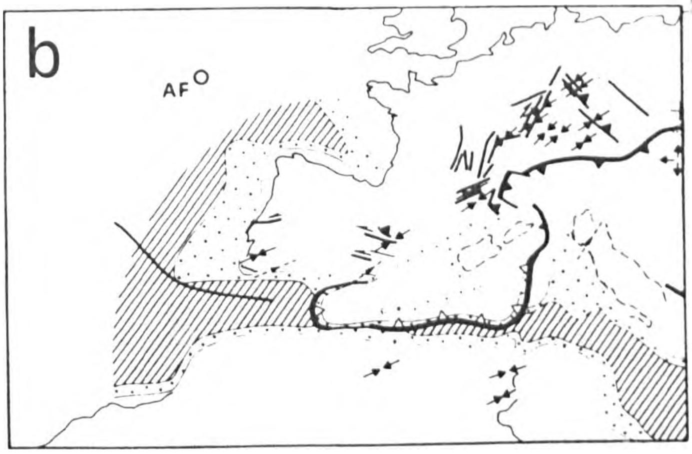


FIG 6.3 (ii) Oligocene extensional phase: E-W extensional regime in Provence



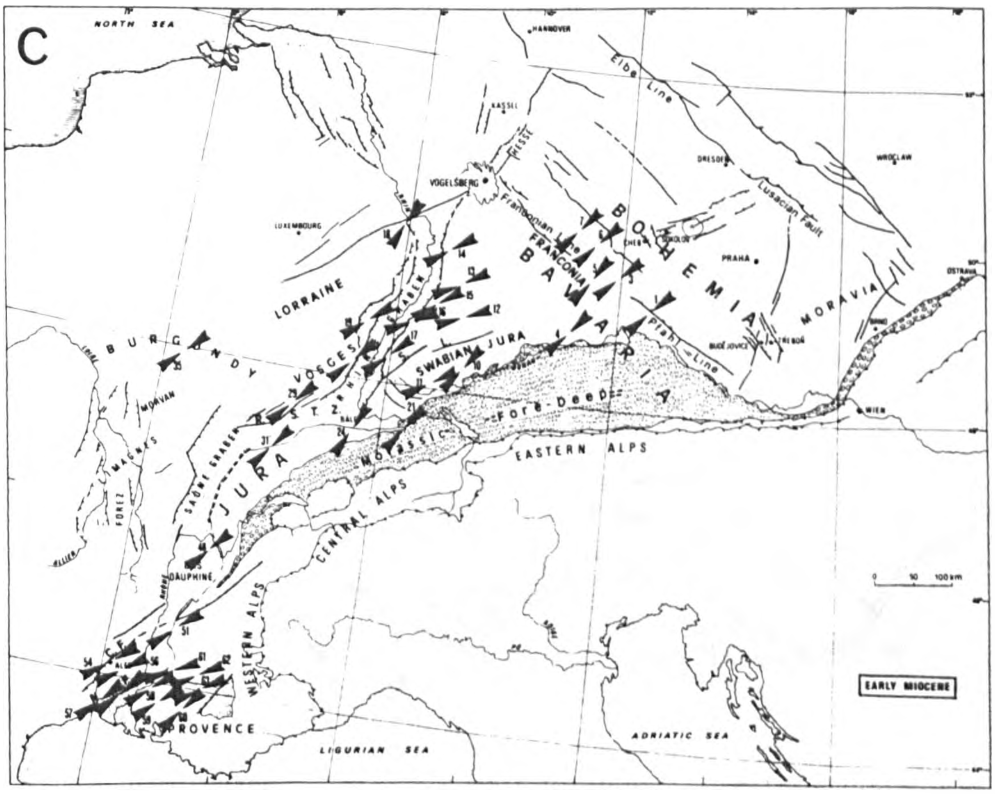


FIG 6.3 (iii) Lower Miocene compressional phase: Transpressional regime in Provence.

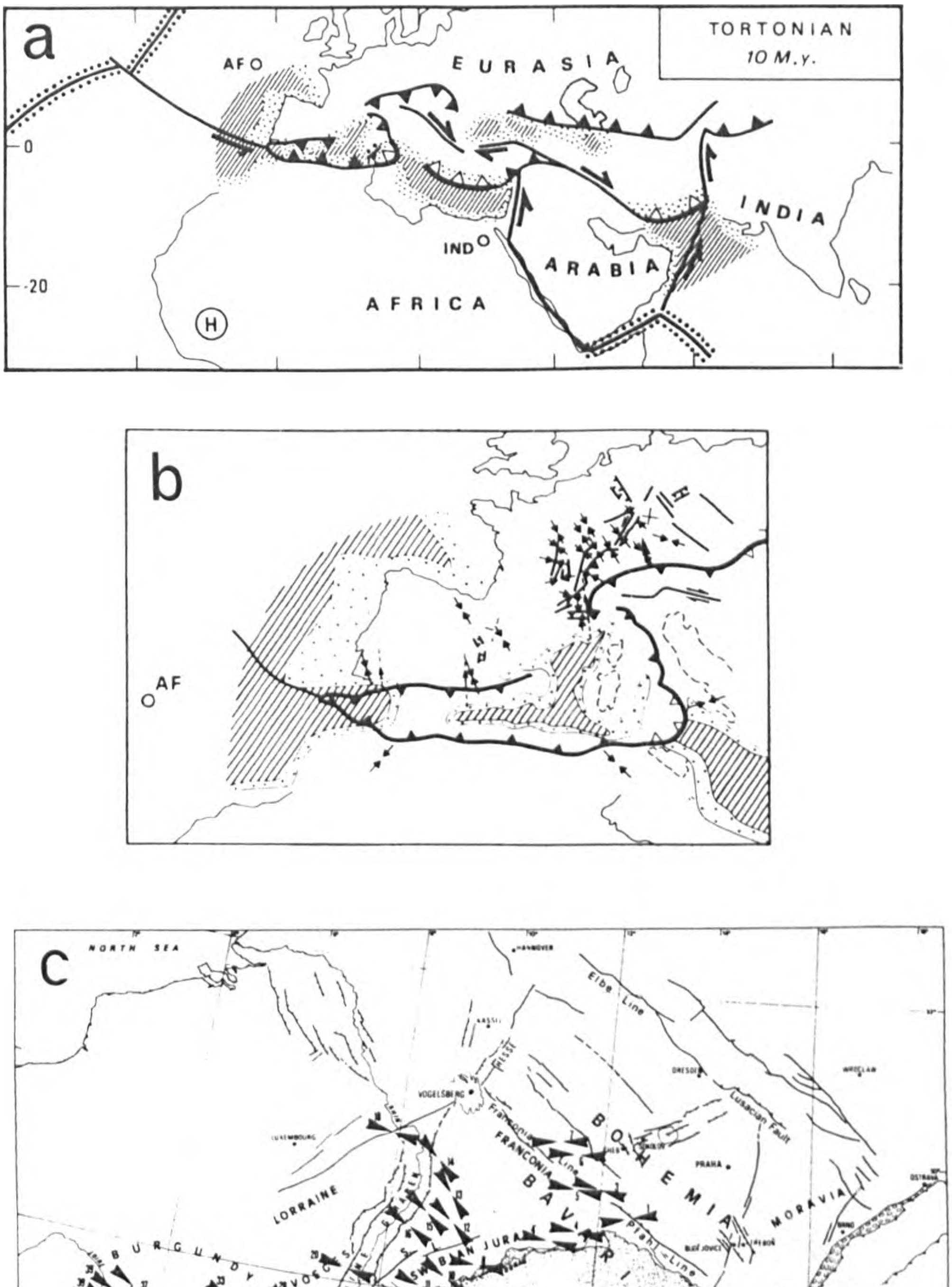


FIG 6.3 (iv) Late Miocene–Pliocene compressional phase. The figure fails to show the NE–SW compressional stresses which were operative in the S.W Alps.

FIG 6.4 (i–iii) Tertiary remnant basins of S.E France superimposed onto structural framework of region. Isopachs of Tertiary are from Debrand-Passard *et al.* (1984)

A -> Alpine compression (Graham 1985)

E -> <- , <- -> : European plate principal compressional, extensional stress (Bergerat 1984)

S.W Alpine foreland basins ////
Alpine thrust front
High-angle faults
HERCYNIAN CRYSTALLINE BASEMENT + + +
P: Pelvoux M-C:Massif Central
A: Argentera M-E: Maures Esterel

GL:Gulf of Lions
V:Vaucluse graben D-V:Digne-Valensole basin
*Au:Autan horst
M:Marseille

Note the E-W extensional stresses which were operative in Provence. The Vaucluse graben (V) is interpreted to form the northern part of a regional graben, the Forcalquier graben, developed in the hangingwall of the Durance fault. Apparent maximum regional subsidence was in the Forcalquier graben - this may reflect lack of data from the Gulf of Lions rift basins.

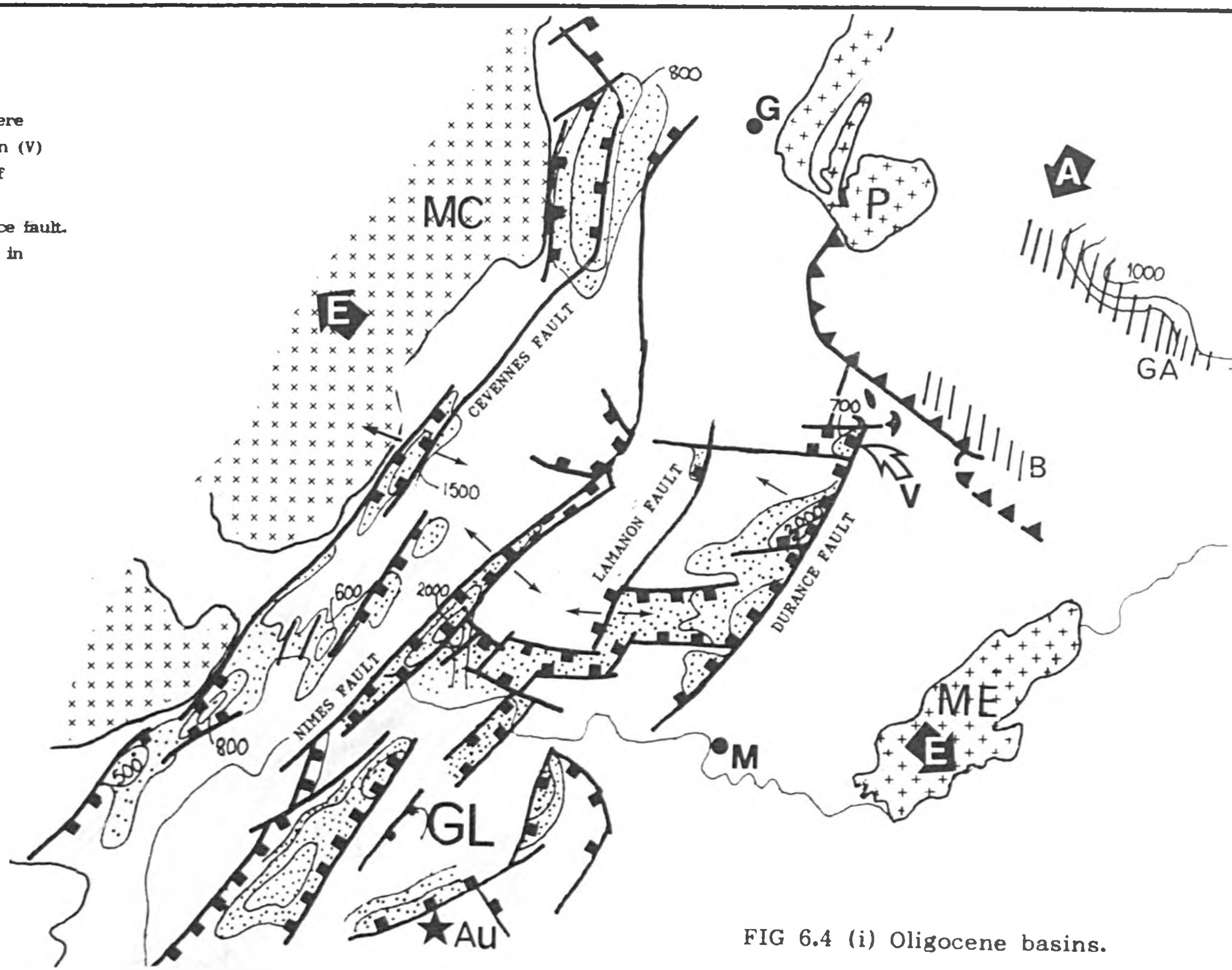


FIG 6.4 (i) Oligocene basins.

Note the NE-SW compressional stress field in the European plate, and also in the S.W Alpine chain. Note that the basins to the west of the alpine chain form a more laterally continous basinal system compared with the small rift basins of the Oligocene. This is a consequence of (i) onset of thermal subsidence (ii) eustatic sea level rise. Maximum subsidence was achieved in the Gulf of Lions where some 3 - 4,000m of Miocene was deposited (Cravatte *et al* 1974).
Note also simultaneous uplift and subsidence on faults in the Digne-Valensole basin.

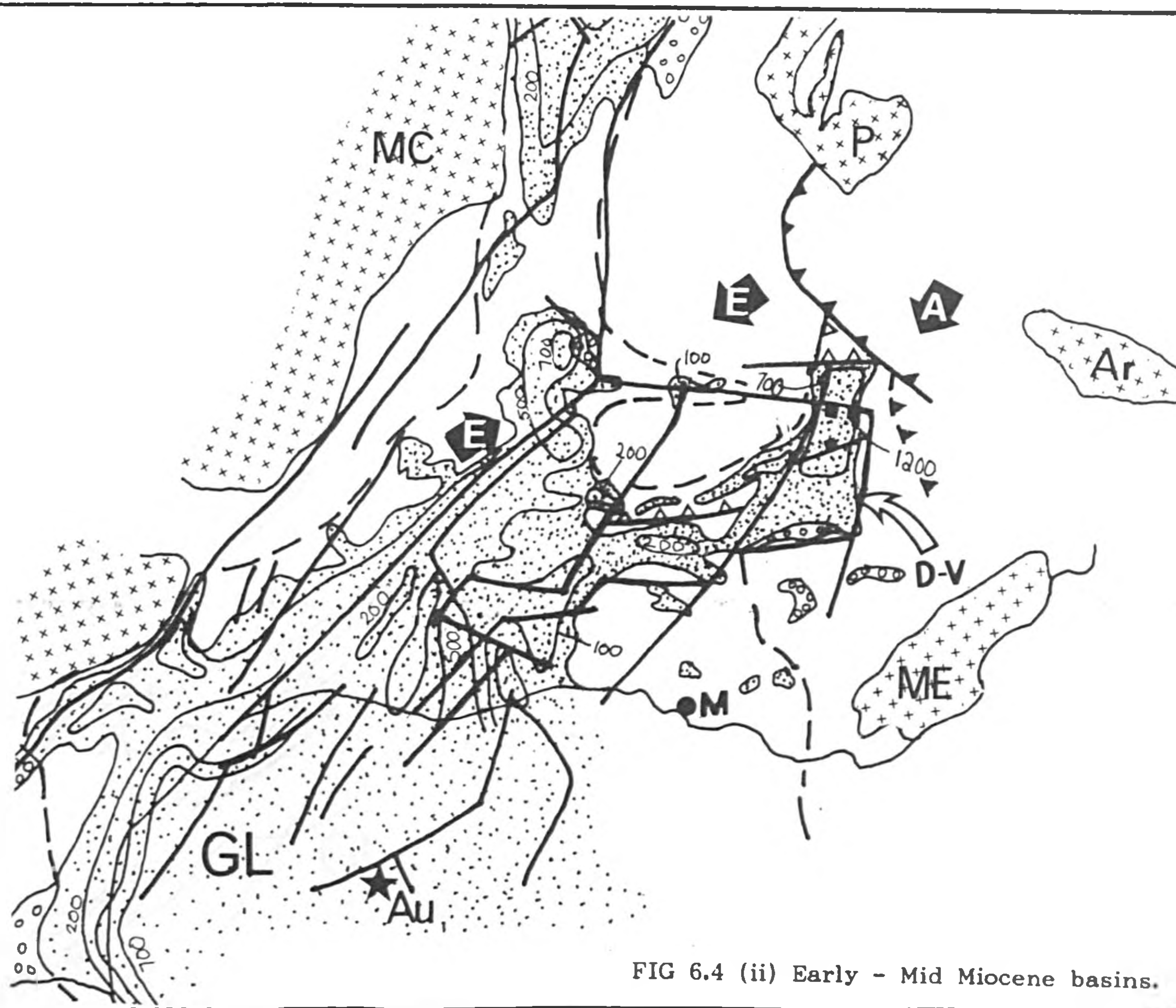


FIG 6.4 (ii) Early - Mid Miocene basins.

Compressional uplift to the east and west of the Digne-Valensole basin as alpine compressional uplift translates into Durancian basin. Sinistral movement along Durance in response to NE-SW alpine compressional stress. Retreat of marine conditions coincident with Messinian sea level fall.

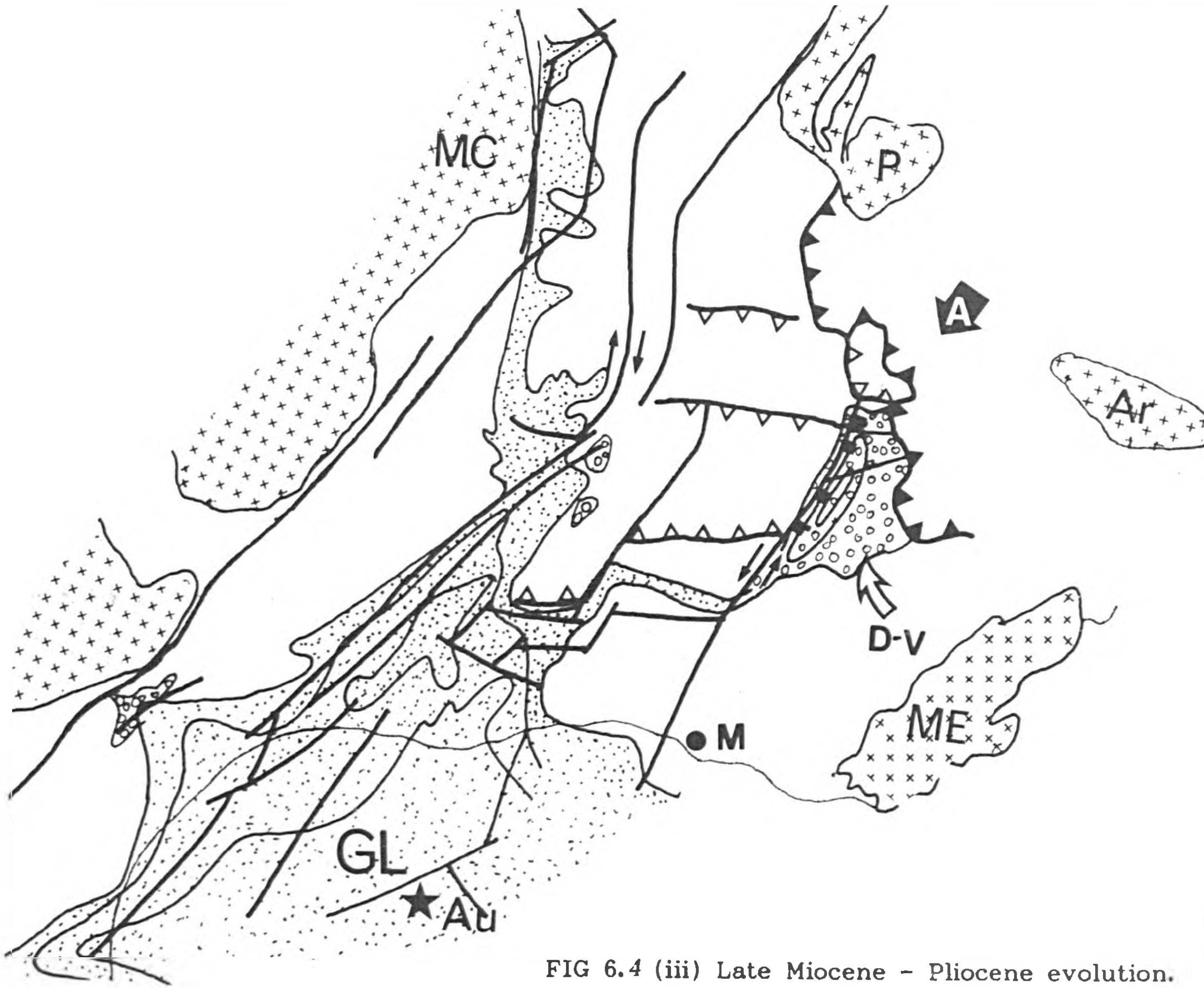
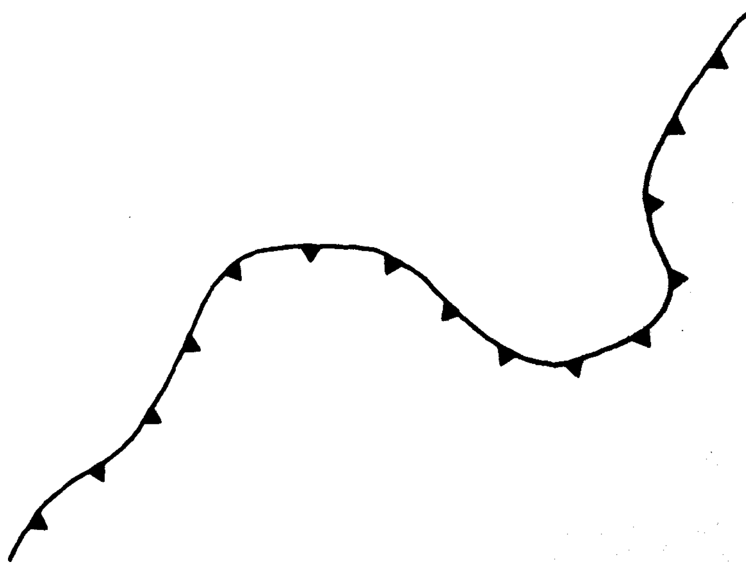


FIG 6.4 (iii) Late Miocene - Pliocene evolution.



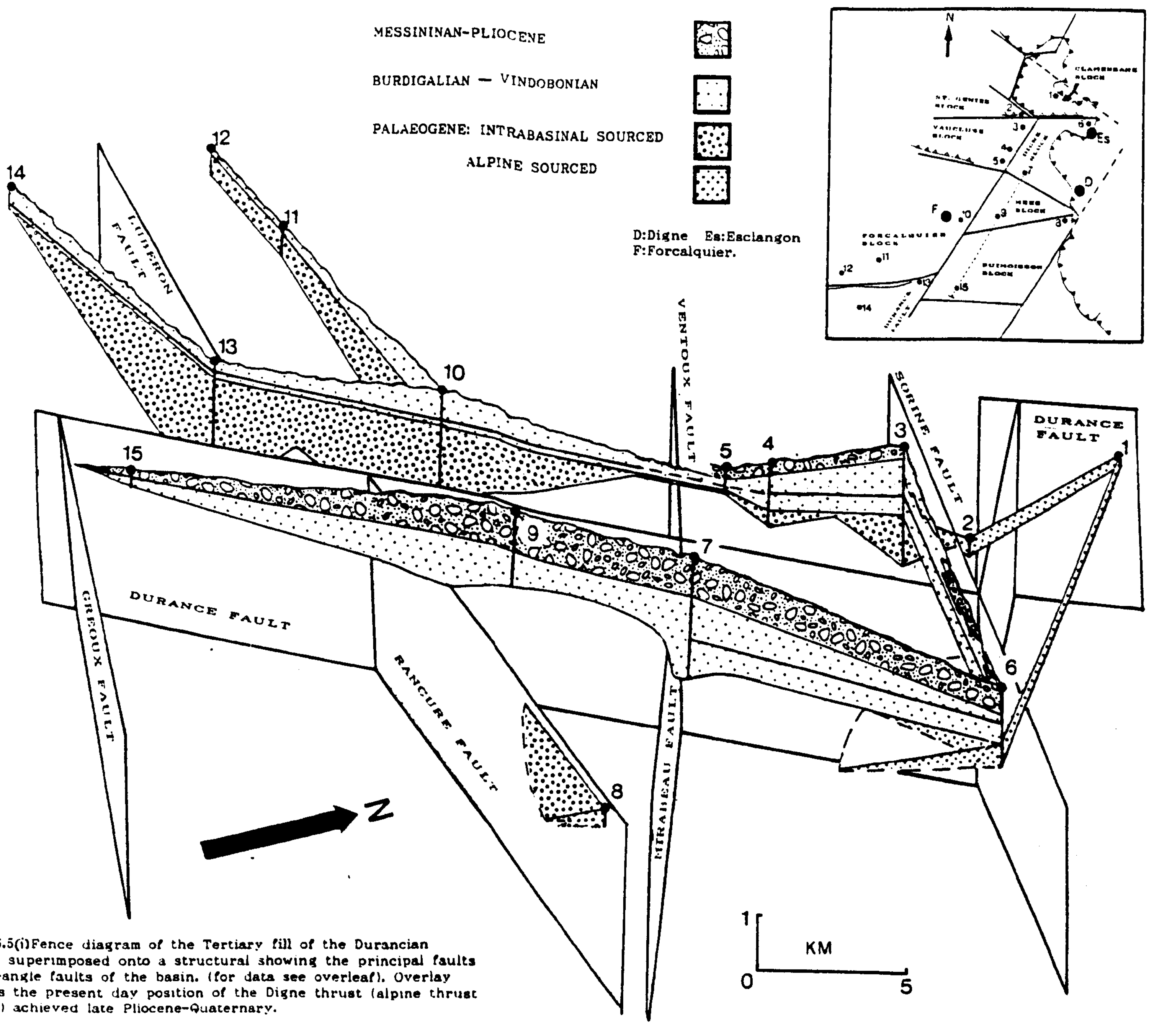


FIG 6.5(i)Fence diagram of the Tertiary fill of the Durancian basin superimposed onto a structural showing the principal faults high-angle faults of the basin. (for data see overleaf). Overlay shows the present day position of the Digne thrust (alpine thrust front) achieved late Pliocene-Quaternary.

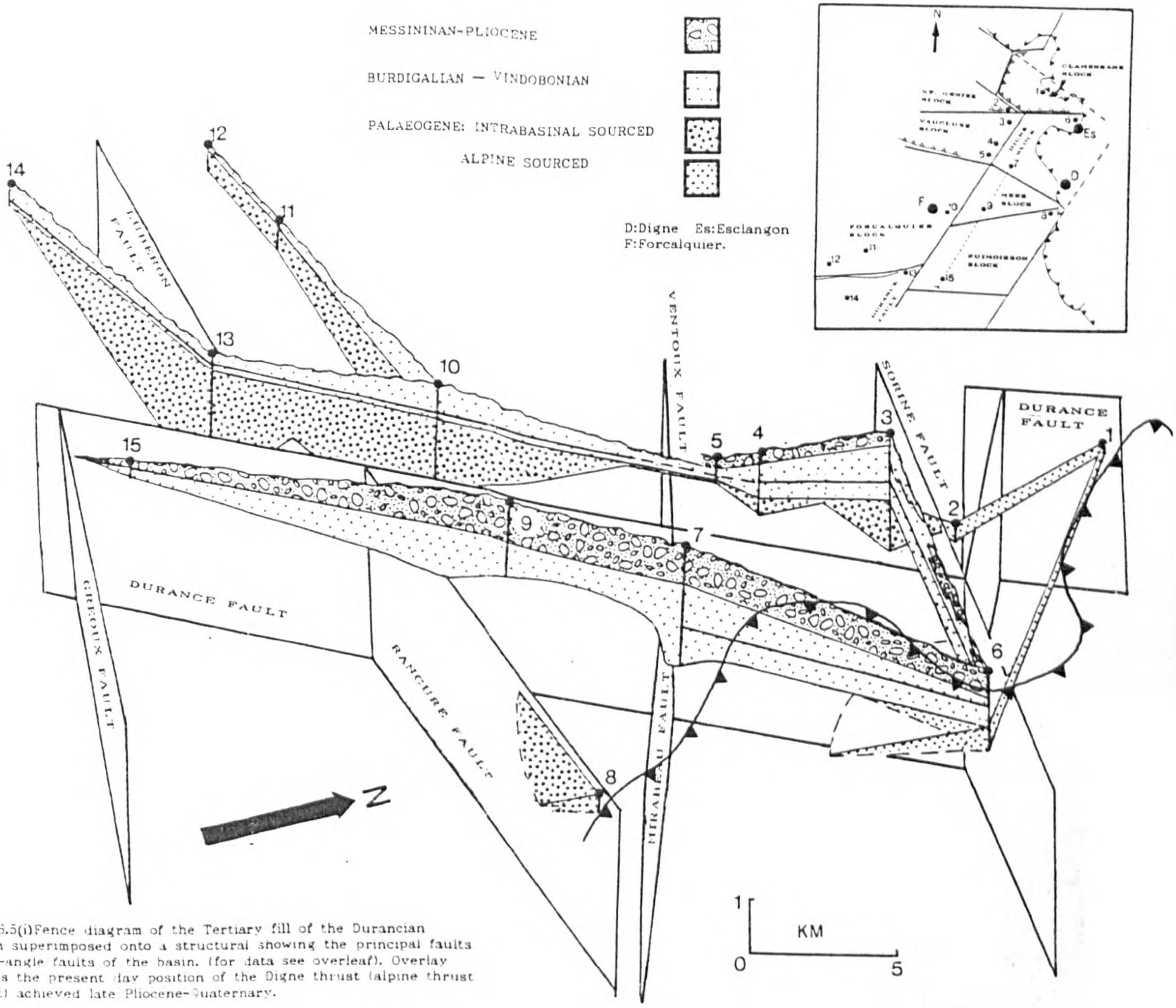


FIG 6.5(i)Fence diagram of the Tertiary fill of the Durancian basin superimposed onto a structural showing the principal faults high-angle faults of the basin. (for data see overleaf). Overlay shows the present day position of the Digne thrust (alpine thrust front) achieved late Pliocene-Quaternary.

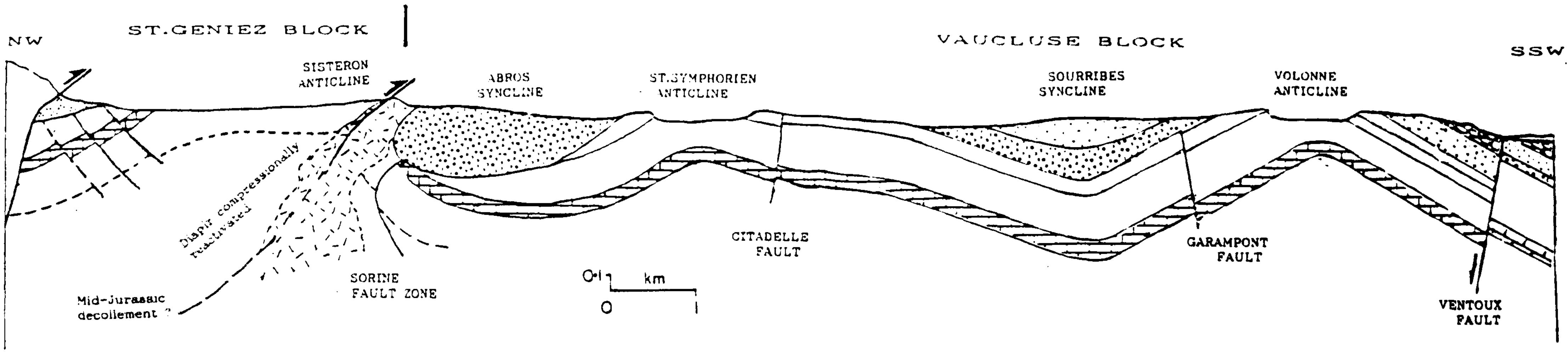
LOCATION	1 MELAN CLAMENSANE BLOCK (Esparron)	2 ST GENIEZ BLOCK (St.Geniez)	3 VAUCLUSE BLOCK NORTH (St.Symphorien)	4 VAUCLUSE BLOCK CENTRAL (Sourribes)	5 VAUCLUSE BLOCK (Volonne)	6 DIGNE BLOCK NORTH (Esclangon)	7 DIGNE BLOCK SOUTH (Mirabeau)
DATA SOURCE	A	A	A	A	A	A	B+D
Messinian-Pliocene	-	-	~300er	~300er	~300er	300	800er
Vindobonian	-	-	700	340	250	400	840
Burdigalian	-	?	300	300		360	420
Priabonian-Aquitanian	180	316	700	300	140	1060	-
TOTAL THICKNESS	180	316	2000	1240	690		2060

8 RANCURE BLOCK	9 LE MEES BLOCK	10 FORCALQUIER BLOCK EAST	11 FORCALQUIER BLOCK CENTRE	12 FORCALQUIER BLOCK WEST	13 MANOSQUE BLOCK NORTH	14 MANOSQUE BLOCK SOUTH	15 GREOUX BLOCK
D	B+C	E	E	E	E	E	C
-	838er	-	-	-	-	-	100er
-	670	720er	30er	100er	250er	330ER	
?		100	100		30	20	100
300	-	1300	500	450	3100	—	—
?	1508	2120	630	550	3380	350	200

FIG 6.5 (ii) Fence diagram data base.

Data Sources: A=Authors field data B=Borehole data from Debrand- Passard *et al.* 1984 C=Seismic reflection data from Dubois & Curnelle (1978) D=Grazianscky *et al.* 1982 E=Jones 1988.

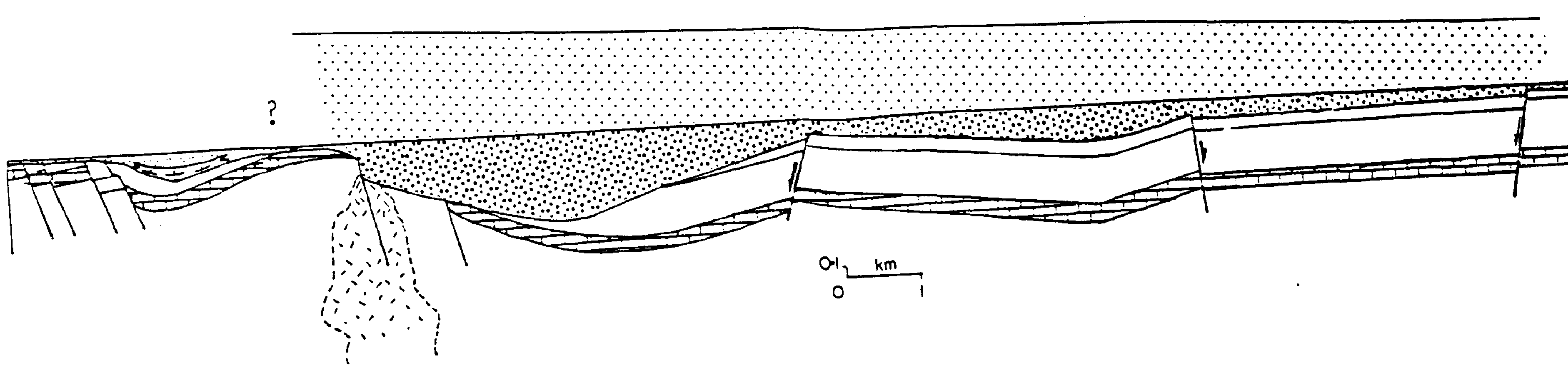
FIG 6.6 (i) PRESENT DAY



Late Miocene- Pliocene structures.

Section shows that the E-W trending Sisteron anticline was reactivated in the late Tertiary alpine phase to form a southward verging anticline with an overturned southern limb. Tertiary sequences are wrapped around the fold passing northward into the Chardavon syncline and southward into the St. Symphorien syncline. The core of the anticline is breached by a southward directed thrust which pierces a remobilised Triassic gypsum diapir. Data of J. Hossack 1987 (*pers. comm* from examination of proprietary seismic data) indicates these late alpine thrust faults in the Barronies region alpine fold - thrust belt, which includes the Sisteron anticline, had a high level, mid-Jurassic decollement. The series of synclines and anticlines to the south of the Sisteron anticline developed as en-echelon folds along the length of the Durance fault indicating a predominantly sinistral strike-slip sense of movement.

(ii) PRE-MESSINIAN

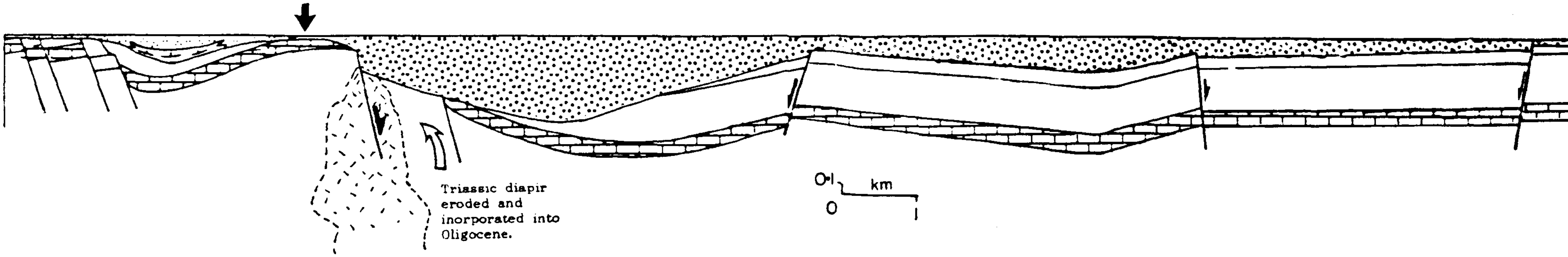


Lower - Middle Miocene structures.

Thickness variations in the Marine Molasse formation are not as pronounced as in the Oligocene but indicate that maximum subsidence continued to occur in the north of the Vaucluse block.

FIG. 6.6 (i-iv) Sequentially restored sections across the Vaucluse fault block for the (i) present day (ii) pre-Messinian (iii) pre-Burdigalian (iv) pre-Priabonian

FIG 6.6 (iii) PRE-BURDIGALIAN

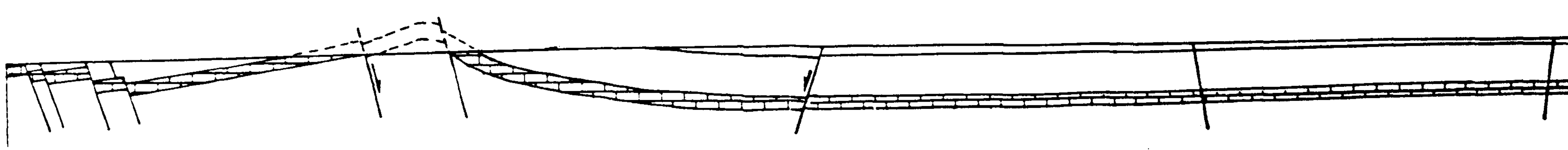


<u>Late Eocene - Lower Miocene Structures</u>

Lateral thickness variations in the Molasse Rouge demonstrate that the folded Mesozoic substrate was transected by a series of high-angle normal faults trending E-W / ENE-WSW. The principal faults were the E-W trending Sorine and Ventoux faults which acted as transfer faults to the regional NE-SW trending Durancian fault. The St. Geniez, Citadelle and Garampont faults are interpreted as accomadation faults. The best exposed faults are those of the St. Geniez fault set (Fig 6. 7). Note how the locally sourced fluvial successions of the Vancon fan system and the alpine sourced Esclangon fluvial system are seperated by the Sorine fault zone.

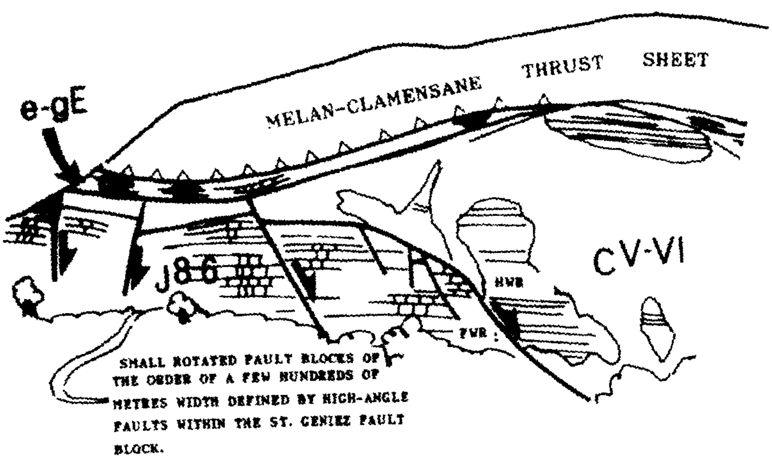
The Triassic is diapirically emplaced. Menard (1979) (Fig 2.4) has shown that in this region the depth to the Triassic is some 5 - 6km. Under high overburden pressures the imposition of inhomogeneties such as extensional faults would have allowed the upward movement of the relatively buoyant salt layers. The developement of diapirs may have been triggered by periods of extensional faulting. During extensional faulting the overburden upon the footwall block is known to decrease as a consequence of hangingwall unloading and footwall wall block erosion, as a consequence of which salt diapirs commonly form in the footwall block to extensional faults.

FiG 6.6 (iv) PRE-PRIABONIAN



<u>Late Cretaceous - Eocene Structures.</u>

The restored base Tertiary section shows that the Cretaceous basement was folded with the developement of an E-W trending and southward verging anticline, the Sisteron anticline. The anticline is developed above the inferred position of the Sorine fault zone, and is interpreted to be the result of the positive inversion of this zone during the N-S directed Pyrenean - Provencal compressional phase.


MELAN–CLAMENSANE THRUST SHEET
e-gE
J8 6
CV-VI
HWR
FWR
SMALL ROTATED FAULT BLOCKS OF THE ORDER OF A FEW HUNDREDS OF METRES WIDTH DEFINED BY HIGH-ANGLE FAULTS WITHIN THE ST. GENIEZ FAULT BLOCK.

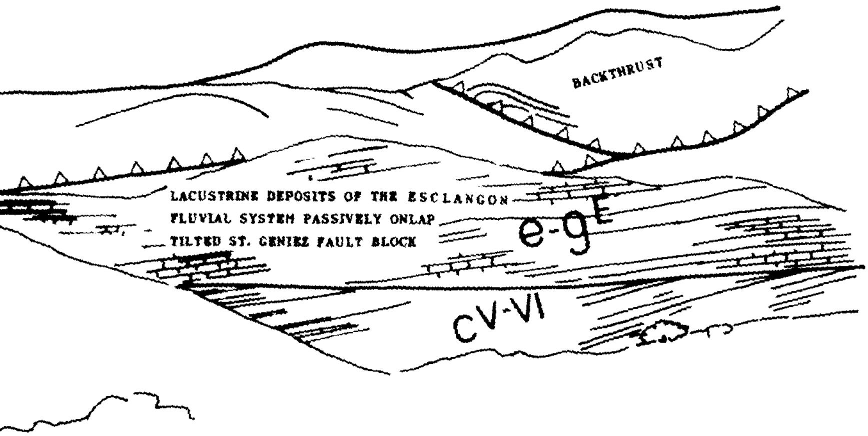

BACKTHRUST
LACUSTRINE DEPOSITS OF THE ESCLANGON
FLUVIAL SYSTEM PASSIVELY ONLAP
TILTED ST. GENIEZ FAULT BLOCK
e-g[E]
cV-VI

FIG 6.7 Tertiary deposits (Molasse Rouge formation, e-g^E, St.Geniez section) onlap a substrate of faulted Cretaceous (cv-v1) - upper Jurassic (j8-6) on the north-west margin of the St. Geniez graben.
The faults form a ENE-WSW trending set and have southward dips of some 50 - 70°, and a spacing of some 0.2 - 1 km. Ramp and flat sections may be distinguished on some of the faults. The faults progressively downthrow to the south with apparent displacement being of the order of 10 -100m.

The Tertiary sequences are cut-off in the footwall of the Melan-Clamensane thrust sheet, a parautochthonous thrust sheet (Gidon *et al.* 1978).

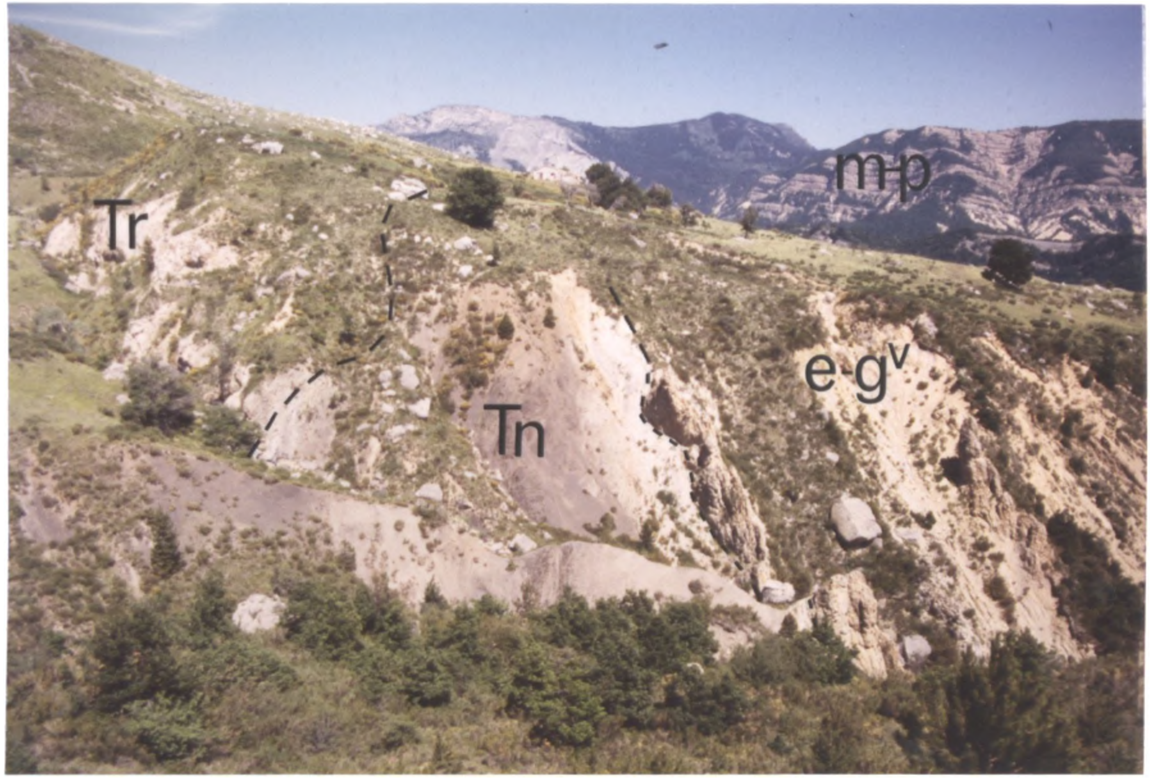


FIG 6.8 Gypsiferous Triassic diapir (Tr) in the core of the Sisteron anticline has an irregular intrusive contact with mid-Jurassic (Tn - Terres Noires) shales which are themselves erosionally overlain by Tertiary fanglomerates (e-g^V) of the Vancon fan system. Abros. Clasts of Triassic siltstone and mudstone are erosionally incorporated in the Tertiary sediments.

The Valensole conglomerates (m-p: Messinian-Pliocene) of the north of the Vaucluse block form the background.

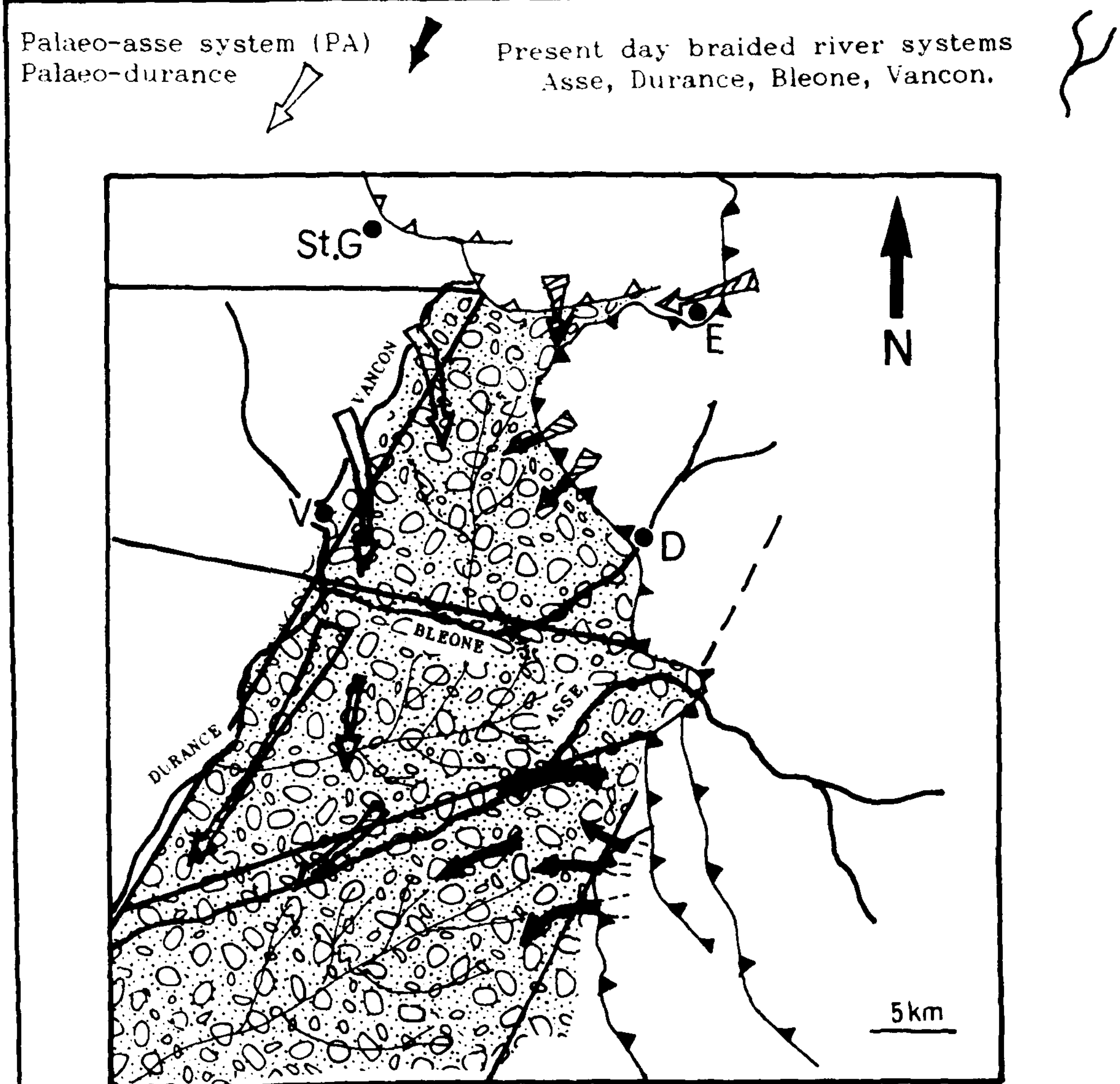
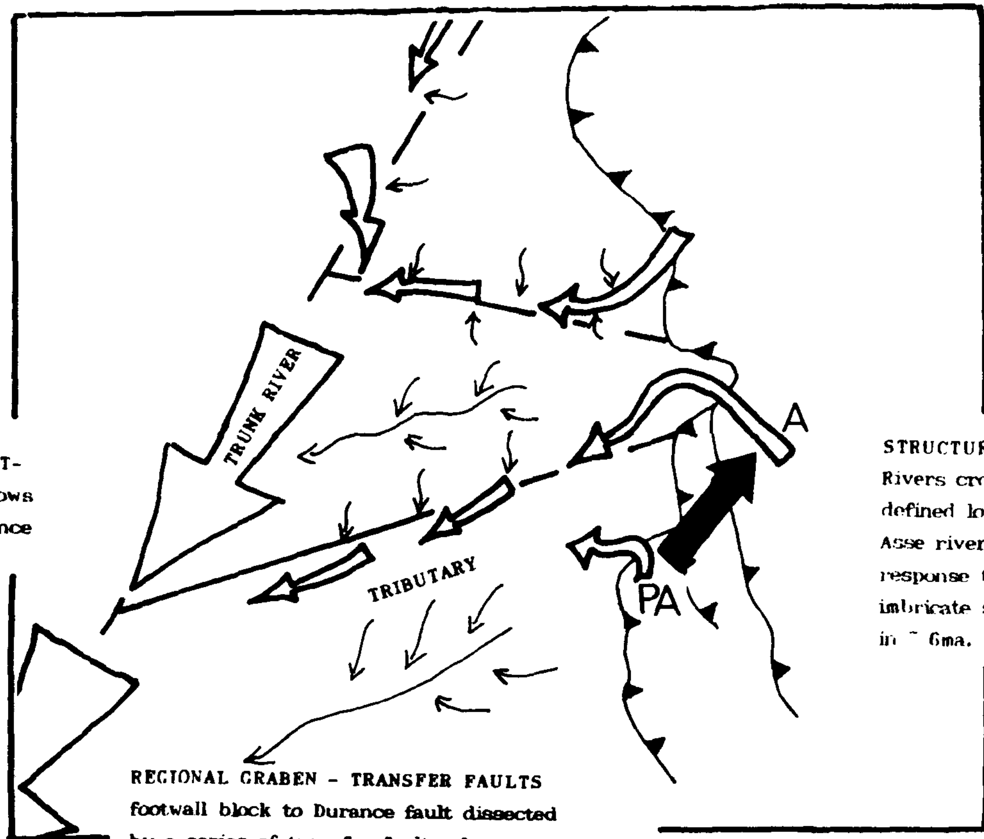
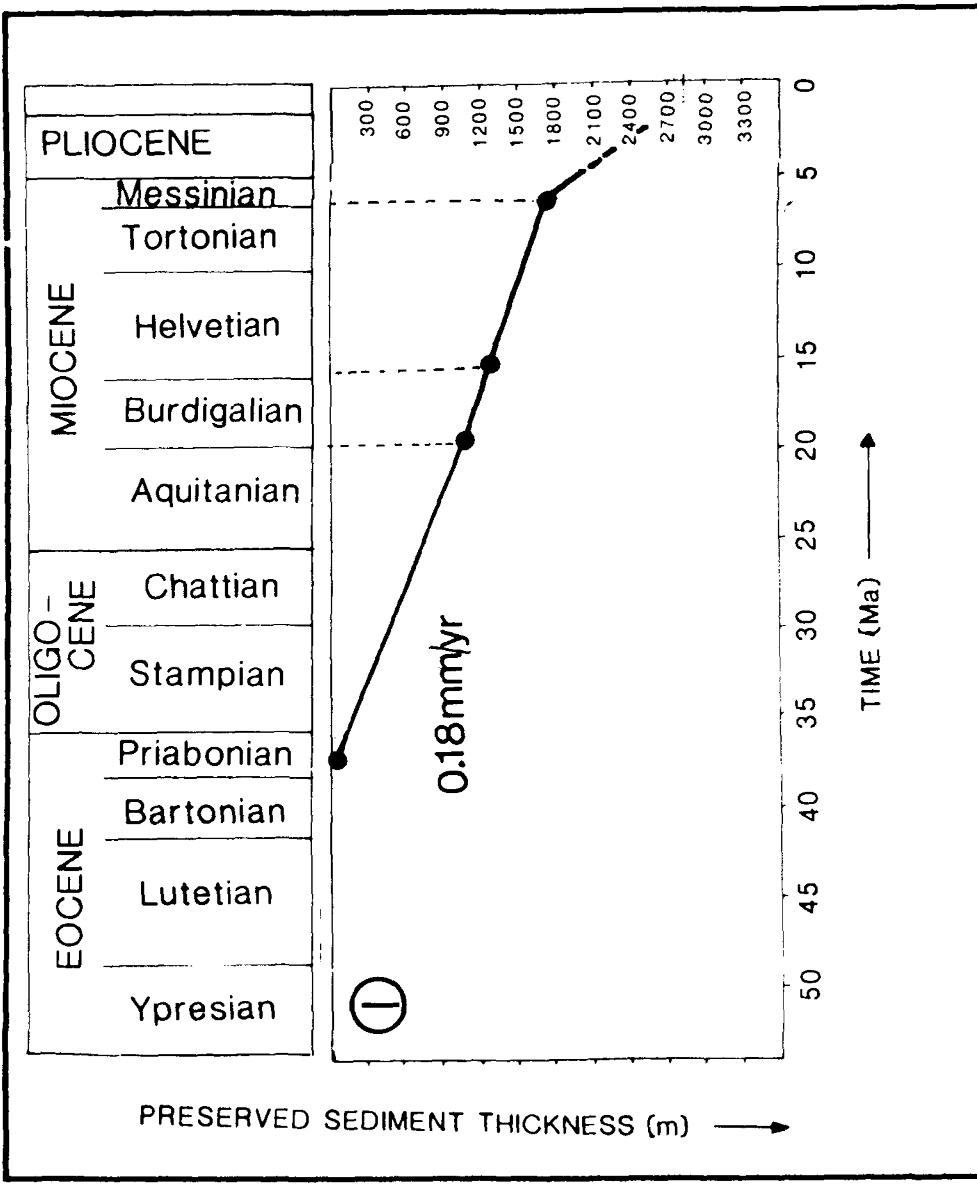


FIG 6.9 (i) Note (i) within the Durance basin palaeo- and modern rivers flow parallel to the strike of the buried high-angle faults. (ii) Note how the Palaeo-asse and present day Asse river cross the Digne thrust front at "structural re-entrants" defined by the position of lateral ramps. The Asse river has migrated northward as the imbricate stack system beneath the Digne thrust has developed Beynes-Sapee-Chiran imbricate stack of Graham 1985) The Asse river system issues into the Digne-Valensole basin where it flows parallel to the high-angle angle Rancure fault prior to intersecting with the regional Durance fault *ie* flows across regional hanging-wall block parallel to transfer fault prior to flowing parallel to regional fault - Fig 6.9 (ii).

FIG. 6.9 (i-ii) Palaeo-flow of Messinian-Pliocene braided stream systems of the Valensole formation (after Grazianscky *et al.* 1982) superimposed onto a structural map of the inherited high-angle faults. Also marked on are the present day rivers of the region.

FIG 6.9 (ii). Sketch interpretation of stuctural controls on Messinian - present day river systems in the Durancian Basin (see caption Fig 6.9 (i) for discussion).





PLIOCENE
MIOCENE
Messinian
Tortonian
Helvetian
Burdigalian
Aquitanian
OLIGO-CENE
Chattian
Stampian
EOCENE
Priabonian
Bartonian
Lutetian
Ypresian
300
600
900
1200
1500
1800
2100
2400
2700
3000
3300
0
5
10
15
20
25
30
35
40
45
50
0.18mm/yr
TIME (Ma)
PRESERVED SEDIMENT THICKNESS (m)

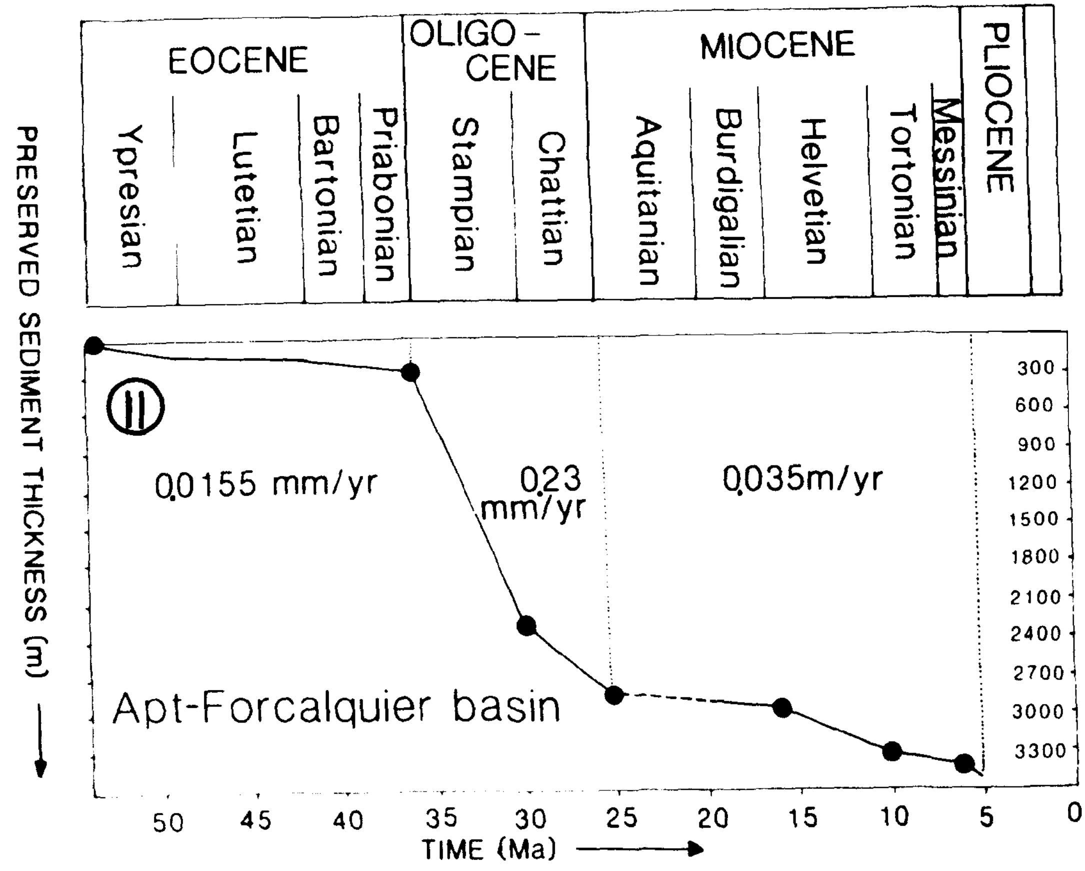


FIG 6.10 Subsidence curves from the Durancian Basin (i) Northern Vaucluse Block, St. Symphorien. Decompacted using programme of J. Marshall *pers. comm* (ii) Forcalquier Basin (uncompacted; from Jones 1988).

CHAPTER 7

CONCLUSIONS

7.1 General.

During the Tertiary the inherited structural framework of S.E France acted as a principal control on the structural and sedimentological evolution of the Durancian remnant basins of the S.W alpine foreland. The prime structural elements of the foreland were a series of NE-SW trending high-angle crustal faults which were initiated during the Hercynian and which were characterised by a multiphase, Mesozoic-Tertiary history of reactivation under both compressional and extensional regimes.

Integrating structural and sedimentological data it has been argued that the Durancian remnant basins developed as a series of discrete sub-basins within a system of differentially subsiding fault blocks defined by inherited faults.

Sedimentary analysis of the Tertiary sequences of the northern Durancian basins show that they had a three stage fill, namely;

(i) a **Priabonian - Aquitanian continental stage** : characterised by the deposition of laterally discontinous continental deposits, the Molasse Rouge formation, by terminal alluvial fan systems.

(ii) a **Burdigalian - Vindobonian marine stage** :characterised by the deposition of more laterally continuous, and dominantly marine successions, termed the Marine Molasse formation, within tide and wave influenced, shallow marine gulfs.

(iii) a **Messinian - Pliocene continental stage**: characterised by the deposition of a thick succession of continental conglomerates, the Valensole Conglomerate formation, by a braided stream system.

Structural analysis, lateral thickness variations and facies association variations in the fill of the basins shows that the above stages relate to a progressive change from an extensional, through transtensional into transpressional regime in the alpine foreland. Inherited high-angle faults were differentially reactivated as extensional or transtensional fault blocks or as thrust sheets under the changing stress regimes.

7.2 Priabonian - Aquitanian Period.

Sedimentary analysis of the Molasse Rouge formation has demonstrated a number of points, namely that:

(i) the principal depositional systems of the northern Durancian basins during the Priabonian-Aquitanian period were continental terminal alluvial fan systems.Two types were recognised, namely:

(a) mass-flow dominated alluvial fans - these comprised the Vancon terminal fan, and the Peroure and la Tour palaeovalley fills.

(b) a stream dominated terminal fan system called the Esclangon system.

(ii) terminal alluvial systems were either intra-basinally or extra-basinally sourced.

The petrography of the mass flow, and stream dominated alluvial systems is markedly different. The Vancon, Peroure and la Tour systems comprise immature carbonate sandstones and conglomerates intra-basinally sourced from Mesozoic limestones, whilst the Esclangon system comprises siliciclastic sandstones having a mature heavy mineral assemblage and sourced extra-basinally from the alpine thrust belt.

(iii) intra-, and extra-basinal alluvial systems were deposited within a system of differentially subsiding extensional fault blocks.

The Esclangon system flowed south-westward across the alpine thrust front into a foreland transected by extensional grabens. These were defined by inherited faults reactivated during the rift phase of the Western Mediterranean extensional system. The fluvial system flowed parallel to the axial strike of the grabens, and to intra-graben structural salients in the form of transfer faults. Offset channel domains within the deposits of the system are interpreted to show that rivers tended towards the point of maximum subsidence in the grabens.

Lateral thickness variations, and palaeoflow trends of the

Vancon fan system clearly demonstrate that it was deposited within a 'trap-door' graben (the Vaucluse graben) on the hanging-wall block of the Durance fault. The lateral limits of the Vancon fan were defined by E-W trending transfer faults, to which the fan system flowed sub-parallel. Tectonic uplift of the footwall block to the Durance fault generated an intra-basinal source for the Vancon fan system.

(iv) extensional fault blocks tectonically separated extra-, and intra-basinally sourced fan systems.

The Esclangon fan system was restricted to the north-eastern part of the Durancian basin, where it was deposited as alluvial fans which terminated into extensional grabens in the immeadiate vicinity of the alpine thrust front.

In contrast, the Vancon fan system was restricted to the south-west of the study area where it was deposited within the Vaucluse graben on the hanging-wall block to the Durance fault. Regional considerations show that this fan was a part of a NE-SW trending system of intra-basinally sourced terminal fans which fringed the western margin of the extensional Durance fault.

(v) deposits of the Vancon fan system form a fining-upward mega-sequence which is interpreted to reflect either the lowering of erosional relief, source area retreat through back-faulting, and/or, a decreasing rate of basinal subsidence, following rapid subsidence with the onset of late-Eoecene / Oligocene rifting.

The deposits of the Esclangon fluvial system show either a fining-upward, or a coarsening and then fining-upward trend which is attributed to the progradation and retreat of the alluvial system in response to differential rates of uplift in the alpine hinterland

In both cases the onset of a regional marine trangression in the Aquitanian may have contributed to the retreat of the alluvial system.

7.3 Burdigalian - Vindobonian Period

Shallow marine conditions were established in the Durancian basin during the Burdigalian to Vindobonian period as the Rhone seaway extended eastward toward the alpine thrust belt in response to a eustatic sea-level rise and the onset of regional subsidence (thermal phase extensional subsidence, and foreland flexural subsidence).

A series of shallow marine gulfs and seaways were developed in a system of differentially subsiding, inherited fault blocks within the Durancian basin. The seaways were structurally constrained lying between palaeohighs developed above transpressionally (?) reactivated high-angle faults and included the Digne-Valensole basin (gulf), and the Jabron basin (gulf).

7.3.1 The Digne–Valensole basin

The Digne–Valensole basin was occupied by a NE-SW trending gulf which connected to the Rhone seaway through the Apt seaway.

Facies analysis of the Miocene marine succession has shown that the gulf had mixed, tide and wave influenced shorelines. Tide dominated shorelines were characterised by an inshore tidal channel and shoal system, whilst wave dominated shorelines were characterised by a barrier(?)-beach system transected by either tidal, or fluvio-distributary channels. The shorelines fronted an upper estuarine tidal channel and flat system which locally passed landward into a coastal alluvial plain. The tidal regime in the gulf was diurnal and probably mesotidal, with the wave regime being of low to moderate energy but storm influenced.

The 900m thick (maximum), Burdigalian to Vindobonian age fill of the basin shows an overall transgressional - progradational trend from shallow marine to coastal fluvial systems which is attributed to a combination of

(i) a Burdigalian-Messinian, eustatic, transgressive -regressive cycle

(ii) regional subsidence

(iii) variable sediment supply rates from the alpine thrust belt.

The basin fill comprises four smaller scale transgressive - progradational cycles of 100–300m thickness, of which three are defined by basin wide transgressions attributed to thrust induced foreland flexure. A fourth cycle restricted to the north of the basin is associated with a syn-tectonic unconformity on the basins margin and records the pulsed, southward progradation of an alpine fluvial clastic wedge into the basin.

7.3.2 Jabron Basin

Facies analysis of the remnant, Burdigalian-Helvetian (Vindobonian) fill of the Jabron basin indicates that it was occupied by a tide dominated gulf. Facies analysis shows that the basin fill comprises a progradational mega-sequence developed above a thin transgressive sequence.

During the lower Burdigalian an E-W gulf extended eastward from the Rhone seaway into the Jabron basin . The gulf was headed to the east by an estuarine tidal system with a tide dominated, open shelf developed in the outer gulf.

The late Burdigalian saw a reduction in the tidal energy in the Jabron basin which is attributed to intra-basinal faulting resulting in reduced connection to the Rhone seaway. The top of the basin fill records the progradation of a locally sourced alluvial fan off the structurally uplifting southern basin margin.

In a regional context the Jabron, Digne-Valensole, and Apt gulf systems have pronounced spatial variations in their lithofacies, with alpine sourced, siliciclastic shoreline facies developed adjacent to the alpine thrust belt passing westward into intrabasinally sourced, bioclastic grainstone facies. This reflects a siliciclastic starvation of the more 'forelandward' offshore areas either as consequence of the net ladward transport of the tidal wave in the Durancian Gulf, or due to bed-load parting zones within the gulf.

7.4 Messinian-Pliocene Period

The Messinian was marked by a eustatic sea level fall, and a progressive change from a transtensional to transpressional regime in the northern Durancian basins. Alpine depormation translated out into the foreland reactivating NE-SW inherited basement faults as strike-slip faults and E-W faults as thrust ramps or normal faults .

In the Durancian basin continental conditions became established which took the form of a braided river system sourced from the advancing alpine thrust belt. Deposition in the Durancian basin was restricted to the Valensole 'block' to the immediate east of the high-angle, Durance fault and to the west of the alpine thrust front.

7.5 General Points.

As a final point the study of the autochthonous Tertiary basins of the S.W Alps has demonstrated a number of key points which are of general application, namely:

(1) Probably the most important point that this study has highlighted is the critical role that major high-angle crustal faults can play in controlling the structural and hence sedimentological evolution of an area. The study lends full support to the concept of "resurgence tectonics", and ephasises the need to understand the inherited structural framework of a region in order to gain a full understanding of the sedimentary history of a basin.

(2) The Durancian basins are perhaps quite unique in comprising extensional basins which developed within the foreland to a genetically distinct thrust-fold belt. As a consequence the basins display the characteristics of both foreland and extensional basins. The extensional basin characteristics were predominant in the early stages of the basins history with the foreland basin characteristics becoming progressively dominant as the alpine thrust belt propagated forelandward.

(3) Following on from the first two points, it has been shown that the inherited high-angle framework controlled the geometry and style of deformation under both extensional and compressional regimes. The orientation of these structures re lative to that of the deformation is the critical factor in controlling which structures are reactivated and in what fashion. As an example NE-SW trending high-angle faults in the autochthonous foreland were reactivated as normal faults in the Oligocene extensional phase, and as strike slip faults in the Miocene when alpine compression was translated along them into the foreland. This concept can be extended into the allochthonous alpine belt where it is suggested that the cover thrust sheets comprise inverted half grabens.

(4) From a sedimentological point of view the inherited structures can be shown to have controlled the sediment dispersal patterns and sites of sediment deposition within the Durancian foreland basins. The reactivation of the fault network of the foreland during the Tertiary created a system of differentially subsiding fault blocks within which continental alluvial systems and shallow marine tidal systems developed. Extra-basinal sourced systems flowed parallel to the strike of both regional and transfer faults prior to terminating within a fault block. In contrast shallow seaways were confined between structural highs developed above inherited faults.

(5) In the somewhat unique case of the S.W Alpine foreland basin, clastic material supplied from the thrust belt to the foreland was trapped in extensional basins adjacent to the thrust front starving those more forelandward sections of the basin which were dominated by intra-basinally sourced detritus. In the Oligocene these intrabasinal deposits were dominantly of lacustrine facies (Jones 1988) but also included extensional fault fringing alluvial fan facies. In the Miocene the intrabasinal deposits were bioclastic grainstones. These are markedly different from the siliciclastic alpine sansdtones ('molasse') which characterises the Oligocene - Miocene fills of the autochthonous foreland basins which fringed the thrust front.

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