TECTONICALLY CONTROLLED FLUVIAL SEDIMENTATION IN THE SOUTH PYRENEAN FORELAND BASIN

"Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor in Philosophy by Antony David Reynolds."

December 1987.

We spent the day on the summit, and I never enjoyed one more thoroughly. Chile, bounded by the Andes and the Pacific, was seen as in a map. The pleasure from the scenery, in itself beautiful, was heightened by the many reflections which arose from the mere view of the Campana range with its lesser parallel ones, and of the broad valley of Quillota directly intersecting them. Who can avoid wondering at the force which has upheaved these mountains, and even more so at the countless ages which it must have required, to have broken through, removed, and levelled whole masses of them? It is well in this case to call to mind the vast shingle and sedimentary beds of Patagonia, which, if heaped on the Cordillera, would increase its height by so many thousand feet. When in that country I wondered how many mountain-chain could have suppplied such masses, and not have been utterly obliterated. We must not now reverse the wonder, and doubt whether all-powerful time can grind down the mountains - even the gigantic Cordillera - into gravel and mud.

Charles Darwin, 1839
"The Voyage of the Beagle"

ACKNOWLEDGEMENTS

I would like to thank Trevor Elliott for his guidance during the preparation of this thesis, for his inspirational field supervision, and for his comments which improved previous versions of this thesis. Financial support was given by: a Natural Environment Research Council grant GT4/84/GS/134; the Department of Earth Sciences, University of Liverpool; my parents and a large overdraft.

I would also like to thank: Jols, Jean, Catherine and Andrew, and Paul; and Rich, Spunky, and Frederique for helping me through crises in research and in life.

Of Spain I have so many fond memories and would like to thank everyone in Graus who made my stay there so enjoyable. In particular the people at Hostal Maella, Aurura, Primi and Jose, and my friends from La Bullanga, Amalia, Pepa, Concha and Caesar among others. Thanks must also be extended to Bill, Rich and Sue who visited me in Graus and with whom I was able to share some of the wonders of Spain.

A special thankyou must be extended to Bill for being himself.

Most of all I have to thank my family whose love and support are the backbone of my life and to whom this thesis is dedicated. Thanks Mother, Big D, Carol, Nick and Dave.

TABLE OF CONTENTS

VOLUME 1 - TEXT

CHAPTER 2

THE STRUCTURAL EVOLUTION OF THE SOUTH CENTRAL PYRENEES

CHAPTER THREE

Diapirism in the South Central Pyrenees

VOLUME 2 - FIGURES

PREFACE

i. Terminology for the classification of fluvial sediment bodies

The terminology for the classification of fluvial sediment bodies follows Friend *et al.* (1979) and Atkinson (1983). The basic unit is the *sediment body*:

> ....the product of a single or series of closely spaced depositional events within the confines of the same river course - Atkinson (1983).

Distinct sediment bodies are recognised by erosive bases, vertical facies sequences and grain size changes. Their classification depends on three features:

(a) Their shape in flow perpendicular sections, especially whether or not they have channelised bases.

(b) Their width (w) to thickness ratio (t) in flow perpendicular sections. Bodies with w/t>15 are termed sheets while bodies with w/t<15 are termed ribbons.

(c) Whether or not the sediment body is the result of a single, or series of filling events - i.e. whether the body is *simple* or *complex* consisting of several *storeys.* Superposed storeys form *multistorey* sediment bodies, while laterally adjacent storeys form *multilateral* bodies.

ii. Imbrication

The following notation has been used to describe pebble imbrication:

a - largest axis of pebble

b - intermediate axis of pebble

c - smallest axis of pebble

p - parallel to flow

t - transverse to flow

i - imbricated, clasts dip upstream.

For example a(t)b(i): longest axis of pebble transverse to flow, intermediate axis parallel to flow and imbricated dipping upstream.

iii. Grain size

The grain size divisions of Wentworth (1922) have been used throughout.

iv. Locations

Locations are given as grid references in square brackets, [8764], they refer to the map enclosures. References are either four figures (grid squares) or six figures (points). Where prefixed by the letters G.R. the reference refers to the Spanish national grid, but the locality is not on either of the two map enclosures.

v. Internal references

References to different parts the thesis are made in italic brackets e.g. {4.2.1(c)}.

vi. Facies

Facies codes are given on Table 4.2 and described in chapter 4.2.

CHAPTER 1

CHAPTER 1

ORIGIN OF THE PYRENEES

1.1 INTRODUCTION

The Pyrenees separate Spain and France. They are a young mountain chain which developed from the late Cretaceous to the Pliocene. Structures related to the chain have a breadth of up to 200km and can be traced laterally for over 1000km, from Asturias in northern Spain eastwards into Provence in southern France (Fig.1.1; Deramond *et al.* 1984).

The mountain chain can be considered in five more or less distinct parts (Fig.1.1; Choukroune and Seguret 1973).

(i) The Aquitaine basin in southern France, which is in part a foreland basin to the Pyrenees.

(ii) A zone of north-vergent thrusts and folds, developed in Alpine basement and cover at the northern margin of the Pyrenees.

(iii) The Axial Zone, bounded in the north by the North Pyrenean Fault. The Axial Zone is composed of Pre-Cambrian (?) to Carboniferous sediments, deformed and metamorphosed during the Hercynian, and gneiss domes and granitoids also of Hercynian age (Zwart 1979).

(iv) A zone of south-vergent thrusts and folds developed in basement and cover at the southern margin of the Pyrenees, including the syn-orogenic South Central Pyrenean basin.

(v) A southern foreland basin composed of two parts: the South Central Pyrenean basin and the Ebro basin in northern Spain.

This chapter introduces the geology of the Pyrenees and outlines their development in three sections. Firstly, palaeomagnetic and regional basin analysis data are considered to establish the microplate and basin-scale evolution of the mountain chain. Secondly, the basins which lie to the north and south of the mountain chain are considered, and a foreland basin model for their Tertiary subsidence is discussed. Thirdly, the structure of the mountain belt is outlined.

The aims of the thesis are set out at the end of the chapter {1.5}.

1.2 ORIGIN OF THE MOUNTAIN CHAIN

The Pyrenees resulted from collision following the closure of a sea which existed between the French and Iberian microplates from the Triassic to the latest Cretaceous. Palaeomagnetic and sea-floor magnetic stripe studies, together with large scale basin analysis, outline three stages in the development of that sea: rifting; strike slip and compression (Grimaud *et al.* 1982, 1983; Boillot 1984; Puigdefabregas and Souquet 1986).

1.2.1. Rifting; Stephanian to late Aptian

The rifting which culminated in the development of a sea between Iberia and France commenced during the Stephanian and Permian with the development of intra-continental fault-bounded basins following the Hercynian orogeny (Fig.1.2; Lucas 1985; Speksnijder 1985). Continued rifting and subsidence in the Triassic and Liassic produced a sequence of clastics (Bunter?) followed by laterally extensive evaporites and limestones, the first marine deposits of the area (Puigdefabregas and Souquet 1986). Rifting at this stage, and tectonism which continued throughout the Lower and Middle Jurassic is considered to have been due to transtension in response to the development of the "Ligurian ocean" (Lemoine 1985; Fig.1.3).

The main rifting phase commenced during the Late Jurassic (Kimmeridgian) and continued into the Early Cretaceous (Late Aptian) heralding the opening of the Bay of Biscay (Pinet *et al.* 1987; Fig.1.4(a)). The upper crust was extended by the development of discrete fault blocks bounded by listric faults. Extension was localised, producing a classic rifted margin at the Biscay margin of the French plate (Fig.1.4(b); Montadert *et al.* 1979) and enhancing a number of discrete sub-basins within the Aquitaine basin (Fig.1.4(c)), the Parentis basin (Fig.1.4(e); Pinet *et al.* 1987) the Ardour basin (Fig.1.4(d); Boillot 1984; Puigdefabregas and Souquet 1986) and the East Central Pyrenean basin (Puigdefabregas and Souquet op.cit.). Some researchers show these basins, as being bounded by E.N.E.-W.S.W.

trending structures (Fig.1.5; Peybernes and Souquet 1984; Curnelle *et al.* 1982; Puigdefabregas and Souquet op.cit.). However, Boillot (1984) argues that the basin bounding faults trended N.E.-S.W. in particular, that the Toulouse fault, a prominent fault active throughout this time in the Aquitaine basin, was a rift-rift transform. This interpretation defines the extension direction *for the whole region* as N.E.-S.W. orientated. Using this direction, Boillot (op.cit) performed an area balance on acoustic basement to estimate 100km of extension for the Armorican shelf, and speculated that, taking in to account extension at the Pyrenean margin, extension may have been as much as 150-200km.

1.2.2. Strike-slip; late Aptian to latest Cretaceous

Northward extension of the Atlantic spreading centre to the west of Spain, and the commencement of sea-floor spreading in the Bay of Biscay in the late Aptian to earliest Albian (J-anomaly, 110m.y.) caused a radical change in microplate motions: Iberia began to move S.E. in a sinistral sense with respect to the French plate (Fig.1.6; Boillot 1984).

This change in microplate motion was almost entirely accommodated in the Central Pyrenean region. Industrial seismic data acquired at the North Pyrenean margin shows the immediate reactivation of normal faults as strike-slip faults, producing flower structures locally (Pinet *et al.* 1987; Hayward pers. comm. 1986). Strike-slip faults also cut across previous faults. For example, the W.N.W.-E.S.E. trending North Pyrenean Fault cuts the Toulouse fault) to break up and displace the rhombic basins formed during the pre-Albian extension (Fig.1.5(b); Boillot 1984; Peybernes and Souquet 1984).

Although the microplate motion was dominated by the Atlantic spreading centre, which produced the strike-slip component, the opening of the Bay of Biscay ensured that the net movement was transtensional (Boillot 1984). This is reflected in the development of local extensional basins right up until the commencement of alpine compression. An Upper Santonian-Campanian extensional phase produced a W.N.W.-E.S.E. elongate turbidite basin, the Vallcarga basin, which is particularly important in the South Central Pyrenees.

1.2.3. Compression; latest Cretaceous to Miocene

Compression commenced during the latest Cretaceous and continued until the Miocene (Boillot 1984; Puigdefabregas and Souquet 1986). It resulted from a change in plate motion caused by the northward extension of the north Atlantic spreading centre, and is dated as post 75 m.y. - the age of the last magnetic anomaly in the Bay of Biscay. The earliest Alpine shortening in the south Pyrenees may have been due to transpressional strike-slip (Puigdefabregas and Souquet 1986; {2.8.3(iv)}).

However, by the Paleocene a convergent margin had developed through the Pyrenean region, extending westwards off the coast of northern Spain and onto the Atlantic sea-floor (Fig.1.7; Grimaud *et al.* 1982, 1983) resulting in limited subduction of the European plate beneath Iberia. Thrusting was confined to a limited area north and south of the mountain chain, but "Pyrenean" deformation extended northwards of the thrust belt into the surrounding basins producing inversion structures (large wavelength anticlines and reverse faults) and initiating diapiric movements (Pinet *et al.* 1987). Curiously, Grimaud *et al.* (op.cit.) suggest that the Iberian plate moved to the N.W. during contraction. This is in marked contrast to the N.E.-S.W./N.N.E.-S.S.W. shortening suggested by a "bow and arrow rule", whereby the thrusting direction is taken as being perpendicular to thrust traces and associated folds (Elliott 1976) and from other structural studies (e.g. Williams 1985). The two data sets are not yet reconciled. The problem is not trivial: balanced sections {1.4.1} are only valid when drawn in the line of tectonic transport.

Estimates of the amount of compression vary widely: Boillot (1984) suggests 120-150km in a N.W. direction from palaeomagnetic stripe studies. Workers using balanced cross-sections estimate north-south shortening of 106km (Williams and Fischer 1984) 115km (Deramond *et al.* 1984) and 55-80km (Seguret and Daignieres 1986) with an increase from west to east.

1.3 ORIGIN OF THE BASINS

The Aquitaine, South Pyrenean and Ebro basins have complex histories. Their Tertiary, synorogenic fill and subsidence history is outlined below and shown to be compatible with a foreland basin model (Allen *et al.* 1986; Brunet 1986) i.e. the basins were caused by the Pyrenean orogeny.

1.3.1 The Southern Foreland Basin

The southern foreland basin is composed of two parts: the South Central Pyrenean basin and the Ebro basin. Overall it is triangular in shape, fringed on each side by mountain chains active during basin fill (Fig.1.8): the Pyrenees to the north; the Catalan coastal range to the S.E. dominated by strike-slip. (Guimera 1984; Anadon *et al.* 1985; Cabrera *et al.* 1985) and the Iberian range to the S.W. dominated by thrust tectonics and reverse faulting (I.G.M.E. sheet 32 1970; Guimera 1984; Julivert 1978).

At the northern margin of the foreland basin a number of thrusts incorporated parts of the basin into the Pyrenean Mountain belt, creating thrust-sheet-top basins (Nichols and Ori 1983) at times, and influencing sedimentation patterns throughout {2.5.3}. The composite effect of thrust sheet ramps, and the increased subsidence towards the mountain chain, was to define an elongate Paleogene basin parallel to the developing Pyrenean chain, the South Central Pyrenean Basin (Fig.1.9).

Through time the depocentre migrated southwestwards from the South Central Pyrenean basin to the Ebro basin (Puigdefabegas *et al.* 1986). The Ebro basin occurs mainly to the south of the Pyrenean thrust sheets (Friend *et al.* 1981; Busquets *et al.* 1985). It has a fill that is Paleocene to Miocene in age, and thickens markedly towards the Pyrenean margin which supplied the majority of its clastic sedimentary fill (Fig.1.10; Friend 1981; I.G.M.E. sheet 23). The fills of the South Central Pyrenean basin and the Ebro basin are discussed below {2.5.3; 4.1.1}.

1.3.2. The Northern basin

The Aquitaine basin has a complex history (Brunet 1984; {1.2}).

In the north extension had ceased by the late Cretaceous, and subsidence was slow - the result of cooling. By contrast subsidence was marked in the south where deep Albian flysch basins coalesced during the Cenomanian, and migrated northwards from Middle Cretaceous to Paleogene times by increased subsidence at their northern edge (Brunet 1984).

Compression caused the inversion of pre-Alpine, listric, normal faults, and the development of a thin-skinned thrust system at high level along the northern Pyrenean margin (James 1984). The Paleocene-Eocene basin fill is composed of flysch facies, passing northwards to a stable carbonate platform (Fig.1.11). The flysch was diachronously replaced by fluvial molasse sediments during the Eocene. Molasse sediment dispersal systems flowed northwards down the dip slope of the rising mountain chain, but turned westwards on reaching the basin which was drained axially (Feuillee *et al.* 1973; Plaziat 1981; Curnelle and Marco 1983)

1.3.3. Geophysical modelling of the basins

Brunet (1986) argues that the Tertiary subsidence of the northern and southern basins was produced by flexure due to Alpine crustal thickening and consequent loading, and so models them as foreland basins.

A foreland basin model (Allen *et al.* 1986) is in closest accord with the southern basin: there is a general southwards (and westwards) migration of depocentres and facies belts with time; basin development and migration are contemporaneous with orogeny; and sediment thickness (subsidence) increases markedly towards the Pyrenean mountain chain. The Catalan Coastal Range and Iberian chain supplied less sediment and, since sediment thickness does not increase markedly towards them, they appear to have been relatively unimportant as loads producing crustal flexure.

At the northern margin the increase in subsidence towards the mountain chain (Brunet 1984) is compatible with a foreland basin setting (Allen *et al.* 1986); as are the northward migration of the basin depocentre and facies belts through time. Brunet (1984) believes that this subsidence had two causes: (i) thermal subsidence following the main rifting, an effect which probably continued until the late Senonian; and (ii) crustal loading by the developing Pyrenean orogen from the late Senonian onwards.

Brunet (1986) considers the Pyrenees to be represented by two semi-infinite elastic plates overlying an inviscid fluid separated by the North Pyrenean Fault (N.P.F.; Fig.1.12(a)). Her model is compatible with the observed structure of the Moho and present-day Bouguer anomalies.

In view of this compatibility Brunet (op.cit.) argues that if at the end of compression in the Miocene, the elevation in the Pyrenees was approaching the present elevation, the topographic load would have been sufficient to explain the formation of the sedimentary basins. However, on the basis of marine Pliocene beds found at elevations of 2000m at the eastern end of the chain (Mattauer and Henry 1974), the Miocene elevation is thought to have been much lower (Fig.1.12(b)). An additional load, probably a cold dense root formed during compression, needs to be postulated. Such a root would take 30-50 m.y. to be removed, either by warming or through detachment and sinking, after which time the influence of the light crustal root would be dominant causing uplift within 10's m.y. Thus, both the delayed uplift of the Pyrenees and the required extra load may be explained by a cold dense root.

1.4 STRUCTURE OF THE MOUNTAIN BELT

1.4.1 Introduction

The structure of mountain belts is best shown on balanced cross-sections, ideally sections balanced in three dimensions (Dahlstrom 1969; Boyer and Elliott 1982; Butler 1985). The power of this methodology is that balanced sections are structural solutions which could be correct: they are "viable" in that they agree with established paradigms, and they are "admissable" in that they agree with collected data (Elliott 1983). In addition, they can be used, firstly, to estimate amounts of crustal shortening during orogeny (Hossack 1979; Elliott and Johnson 1980) and, secondly, to produce sequential restorations of the thrust belt, back to the pre-thrusting configuration.

Balanced sections have been used with much success at high levels in thrust belts, for example: in the Swiss Alps (Boyer and Elliott 1982); in the French Alps (Graham 1985); in the Rocky Mountain foothills (Price 1981); in the "foreland" of the Scottish Caledonides (Elliott and Johnson 1980); in Scandinavia (Townsend *et al.* 1986; Moreley 1987) and in the Appalachians (Kulander and Dean 1986). High structural levels are particularly suitable for section balancing because deformation commonly approximates to plane strain , i.e. material does not move in or out of the plane of section, so that cross-sectional area is preserved between the deformed and restored sections (Dahlstrom 1969). Where beds are of constant thickness and bed lengths are preserved, line balancing techniques can be used (Dahlstrom 1969); however, as is common, where bed thicknesses vary, or thickness is not preserved during deformation area balancing is more appropriate (Hossack 1979).

There have been few attempts at balancing entire orogens; notable exceptions are the work of Coward and Butler (1985) Butler *et al.* (1986) and Butler (1986). The paucity of such studies is explained by two factors: firstly, because deformation at depth is ductile, the cardinal assumption of plane strain may not hold - not even approximately; and, secondly, the quality and quantity of data decreases with depth. Nevertheless, with the proviso that plane strain may not have occurred, so that the section can only represent an approximation of the true structure, a balanced section is the clear aim of all regional structural studies.

The sections which follow outline: firstly, the data set with which a balanced cross-section through the Pyrenees must be compatible; secondly, the postulated nature of the driving forces of orogeny, a section which represents a historical review of models of the Pyrenean orogeny, and, thirdly, the range of structural models presented during the last three years, a range which can be divided by the different structures inferred to exist at depth.

1.4.2 Features constraining balanced sections of the Pyrenean mountain chain

A balanced section should be in accordance with all available geological data. However, at the orogen scale, for the Pyrenees five key features emerge from that data set to constrain the geometry of the balanced sections:

(a) Major shortening occured

Microplate reconstructions show that considerable shortening occurred in the Pyrenean region from the Late Cretaceous to the Eocene (Grimaud *et al.* 1982, 1983). Boillot (1984) estimates 120-150km in a N.W. direction {1.2.3.}.

(b) The North Pyrenean Fault (N.P.F.)

The N.P.F. forms the northern margin of the Axial Zone. It is believed to have been a major strike-slip fault during the Upper Cretaceous. Movement on the fault is dated in two ways: firstly, radiometrically - by associated metamorphism of 90-96m.y. (Albarede and Michard Vitrac 1978) and lhertzolites intruded along the fault (116 +/- 5m.y.; Verschure *et al.* 1967) and, secondly, by tectonic relationships - it cross-cuts the Toulouse fault, which was active during the Late Jurassic-Aptian rifting phase (Boillot 1984), and is in turn truncated in the footwall of the "Alpine" Pic Lakhoura thrust (Deramond *et al.* 1984). Some authors (e.g. Mattauer 1968; Zwart 1979) consider that the N.P.F. is, in part, a reactivated late Hercynian normal fault.

The fault directly overlies a 15km step in the Moho with thicker crust to the south and thinner crust to the north (Gallart *et al.* 1980; Daignieres *et al.* 1982; Fig.1.13).
At the surface it separates the distinct statigraphies of the French and Iberian microplates, and is considered to be an expression of the suture between them (Choukroune and Seguret 1973).

(c) North- and south-vergent structures

North-vergent thrusts and folds occur at the northern margin of the Pyrenees. Similar south-vergent structures occur at the southern margin. There, however, some of the thrusts dip steeply to the south, and some of the associated folds are downward facing (Seguret 1972; Williams 1985) {2.3.2.d}.

(d) Non layer-cake stratigraphy

The deformed Permian to Cretaceous sequences were deposited in a variety of fault bounded basins, so that the stratigraphy is not "layer-cake". Instead thicknesses and facies change rapidly.

(e) No evidence for subduction

There is no evidence for the existence of oceanic crust between the Iberian and French microplates in the Pyrenean region: there are no ophilites and there is a general lack of volcanism. So, it is thought that no subduction occurred, and that crustal area should be conserved on the balanced and restored sections.

1.4.3. The Driving Forces of Orogeny

Much of the early debate on the evolution of the Pyrenees centred on the driving forces of orogeny. The debate was strongly influenced by the evidence of thrusts and folds at the southern margin of the Pyrenees which dip steeply to the south - the Nogueras Zone (Seguret 1972) {2.3.2(d)} - many authors took them as evidence that the deformation was gravity-driven (Van Bemmelen 1955; Seguret 1972; Choukroune and Seguret 1973; Sole-Sugranes 1978).

Indeed, Van Bemmelen (1955) suggested that the Pyrenees were the result of vertical motions only. He believed that a low density "blister" produced uplift in the Axial Zone, and caused gravity-driven thrust sheets, in both the cover and the basement, to be shed to the north and south away from the central high (Fig.1.14(a)). This model can be rejected on two grounds: firstly, because 120-150 km of shortening occurred across the orogen; and secondly, because there is no evidence for major crustal thinning in the Axial Zone - a feature which would be a direct consequence this model.

Choukroune and Seguret (1973) also proposed a two stage model. However, they argued for uplift in the Axial Zone to be the result of compression, with only the high level thrust sheets {2.4} being affected by gravity gliding (Fig.1.14b&c). The first part of their model was supported with extensive microstructural studies, which showed that planes of flattening fan about the vertical across the Pyrenees (Fig.1.15) - a geometry only compatible with an overall shortening perpendicular to the mountain belt. Gravity gliding was favoured for the higher thrust sheets for two reasons: firstly, because of small normal faults and an extensional joint set observed in the higher thrust sheets, and considered incompatible with compressional tectonics; and, secondly, because of the south-dipping thrusts and downward facing folds at the southern margin which implied the correct geometries for sliding. However, as before, there is no evidence for major extensional faults in the Axial Zone - a requirement if the higher nappes were emplaced by by gravity gliding (Cooper 1981) - a feature noted by Choukroune and Seguret (op cit.) themselves.

Williams and Fischer (1984; Fig.1.16) argued that all the structures were produced by <u>compression</u> on a linked thrust system whereby displacement on high level thrusts links downwards into the basement. Their model is in accordance with the absence of extensional faulting in the Axial Zone: it suggests rather than being gravity emplaced, that the south-dipping thrust faults were folded into that attitude by the development of lower thrust sheets. An origin by compression and a linked thrust system model is considered to be the best explanation of the Pyrenean orogen. However, the section which Williams and Fischer (op.cit.) present is not compatible with the strain data; it also fails to accord with a number of other key features of the Pyrenees {1.4.4(a)}.

1.4.4 <u>Recent Models</u>

The section *of Williams* and *Fischer* (1984) was an advance on previous sections in that it could be restored to the undeformed state.

Temporally, their work was closely followed by a number of other sections, all of which are in consensus that (a) the Pyrenees resulted from compression (b) sections drawn across the mountain chain should be balanced, and (c) the structures at high level are best explained by a thin-skinned model. However, there is no consensus on structures at depth. Three distinct models have emerged: the thin-skinned model of Williams and Fischer (op.cit.; Fig.1.16); a thick-skinned model, Deramond et al. (1984; Fig.1.17); and an inhomogenous-strain model, Seguret and

Daignieres (1985, 1986; Fig.1.18). The sections are located on Figure 1.19.

(a) <u>The thin-skinned model</u>

In a true thin-skinned model thrusts flatten at depth, Coward (1983). Williams and Fischer (1984) show a discrete, low-angle sole thrust dipping steadily to the north at an angle of 6° – a geometry which approximates to the thin skinned model in that the faults do not steepen at depth. The majority of the shortening is directed to the south. The north-directed shortening at the northern margin of the Pyrenees is interpreted as a backthrust fan branching from the sole thrust. The North Pyrenean Fault (N.P.F.) is truncated at depth by both the sole thrust and the backthrust fan, so that its extension at depth is inferred to lie somewhere below the Aquitaine basin, and therefore does not correspond to the pronounced step in the Moho observed directly below the present outcrop of the N.P.F. The observed geometry of the Moho is unexplained by this model, so that ~~it~~ the model is not favoured. Three other problems are also apparent with this section: firstly, the section uses a line length balance in the Axial zone and the Nogueras zone where line lengths, the number of thrusts and depth to decollement are all unknown, so that the structure shown is unconstrained; secondly, the Jurassic to Cretaceous sequences on the balanced section have geometries that do not correspond to the basins in which they are inferred to have been deposited, and thirdly, the deformed state section is in conflict with unpublised and subsequent seismic, borehole, and outcrop data obtained in the Spanish Pyrenees {2.7}. In summary the section is viable but not admissable – it is not a balanced section.

(b) <u>The thick-skinned model</u>

In a thick-skinned model thrusts steepen at depth – the change in fault angle implies major strains which are extensional, flat-lying at high level, and extensional, vertical at depth (Coward 1983).

Deramond et al. (1984) show a thin-skinned system at high levels which steepens at depth to cut across the crust-mantle interface and join a decollement between the crust and the mantle north of the N.P.F. (Fig.1.17). Their section is a marked advance on that of Williams and Fischer (1984): the restored section shows the lower crust greatly attenuated, a geometry more compatible with that of a rifted margin. Furthermore, the strains implied by the model appear compatible with those observed by Choukroune and Seguret (1973; Fig.1.15). However, once again, the section does not explain the step in the Moho, and there is no clear relation between the restored section and the fault bounded basins in which the Permian to Cretaceous sequences are believed to have been deposited.

(c) Inhomogenous strain model

In the inhomogenous strain model, brittle displacements along high-level thrusts result from penetrative ductile deformation at depth, probably localised in broad shear zones (Seguret and Daignieres 1986; Fig.1.18). The model is compatible with the Pyrenean strain data which suggests horizontal ductile shortening in the Axial Zone, with vertical extension transferred into nappe displacement near the surface {1.4.3}.

The sections of Seguret and Daignieres (1985, 1986) are essentially drawn in two parts: the cover rocks are restored by line balancing, whereas the basement (down to the geophysical Moho) is restored by area balancing. An important part of the model is that the N.P.F. is taken as a pin line. The interpretation follows Gallart *et al.* (1980) and Daignieres *et al.* (1982) who consider that the Moho step is directly related to the N.P.F., and that it is a long-lived feature inherited from the late Creteaceous when the N.P.F. was active.

The choice of the N.P.F. as a pin line controls the geometry of the section. One consequence is that the restored section shows the lower crust thinned below the Aquitaine basin, but of normal thickness to the south of the fault. The geometry is compatible with the thick Jurassic and Cretaceous sequences preserved in the Aquitaine basin, and the relatively thin sequences in the south. Furthermore, the sections of Seguret and Daignieres (1986) are a marked improvement in that the cover rocks restore to basin geometries with which their sedimentology is compatible. A less fortunate consequence of choosing the N.P.F. as a pin line is that the geometry of the Axial zone is unconstrained - the area of Axial Zone crust which has been eroded from above the present-day topography is unknown. Although Seguret and Daignieres (op.cit.) estimate that the resulting error in the shortening estimate is only +/-10km. An interesting feature of the sections is that they imply that the majority of the thrusting at the western end of the mountain chain was directed to the north, whereas in the east most of the thrusting was south-directed. Seguret and Daignieres (1986) infer 55-80km of shortening increasing towards the east. However, one of their sections (Fig.1.18(b)) does not balance. It underestimates the contraction in the cover south of the N.P.F. by 10km. Unfortunately, the structures on the other figures are too small to accurately check the sections.

In summary, although one of the sections is not viable the models of Seguret and Daignieres (op.cit) are preferred - mainly because they adhere most closely to the geology of the Pyrenees. The estmated values of shortening from these sections are much less than the 120-150km of shortening estimated independently by Boillot (1984). If this higher estimate is correct then major changes in the section can be expected as the data set is widened and refined.

1.5 AIMS OF THESIS

The study area lies in the South Central Pyrenees, between Ainsa and Tremp, and Barbastro and Benasque (Figs.1.1; 1.8 and 1.20). It lies south of the Axial Zone, but includes the south vergent thrust and folded zone, the eastern part of the South Central Pyrenean Basin and the northern margin of the Ebro Basin.

The aims of the thesis are fourfold: firstly, to document new balanced structural cross-sections from the Axial Zone to the Ebro basin along the Esera and Ribagorzana valleys; secondly, to use these sections to outline the three-dimensional structural and sedimentological evolution of the South Central Pyrenees from the Upper Cretaceous to the Miocene; thirdly, to describe some of the geometries and effects of diapirism within the foreland basin, and fourthly, to characterise the nature of a large fluvial system which existed in the study area during the Oligo-Miocene.

Throughout the thesis emphasis is lain upon a careful consideration of the relations between sedimentation and tectonics: for the structural sections syntectonic sediments are used to test structural geometries and infer their timing, while structural studies form the framework for detailed sedimentological work on the Oligo-Miocene fluvial system. A process of constant feedback and refinement occurs between the two disciplines.

CHAPTER 2

The kid on the mountain (trad. Irish).

Jordan (1977)

CHAPTER 2

THE STRUCTURAL EVOLUTION OF THE SOUTH CENTRAL PYRENEES

2.1 INTRODUCTION

This chapter outlines the structural evolution of the eastern part of the South Central Pyrenees. The study area extends from the Axial Zone to the Ebro basin and is bounded to the east by a major oblique ramp, the Segre fault and to the west by a north-south trending anticline, the Boltana anticline (Fig.1.20).

Before addressing Pyrenean problems a brief section discusses the importance of a linked sedimentological and structural approach to section balancing. In this way the philosophical approach to the problems is outlined.

The structure of the study area is introduced in two ways. Firstly, by outlining the structural style seen throughout the South Pyrenees {2.3} and, secondly, through a historical review addressing specific structures seen within the foreland basin {2.4}.

Following the introduction of the structural framework, the pre-orogenic and syn-orogenic stratigraphies are discussed {2.5}. The discussion is not exhaustive, but is sufficiently detailed to allow new balanced sections along the Esera {2.6} and Ribagorzana {2.7} valleys to be outlined. These sections are then linked to define a series of "balanced maps" which outline the structural evolution of the South Central Pyrenees {2.8}.

2.2 BALANCED SECTIONS - THE IMPORTANCE OF SEDIMENTOLOGY AND STRATIGRAPHY

The structural style, of plane strain {2.3.2(b)} and decollement {2.3.2(a)} make the South Central Pyrenees well suited to balanced section construction. The techniques involved are well established in the literature {1.4.1} and will not be discussed here.

However, it should be emphasised that the techniques were developed in areas considered to have a "layer cake" preorogenic sequence, i.e. a sequence constant in thickness and facies. Because of this the importance of variations in stratigraphy and sedimentology has not been fully explored; indeed, in some studies it has been ignored (e.g. Williams 1985; Fillbrandt *et al.* 1985). Marked variations in the pre-orogenic and syn-orogenic sequences of the Pyrenees demand that this factor is addressed in full. Two principle contributions are apparent: the pre-orogenic sequences can help to constrain the restored section, while the syn-orogenic sequences can help constrain the deformed-state section, and intermediate state sections.

2.2.1 The Restored Section - the pre-orogenic stratigraphy

After having plotted all known surface and subsurface geology on a topographic section, the first step in section balancing is to construct a stratigraphic template on which to unravel the deformed sequence. The key question at this stage is: in what basin setting was the deformed sequence deposited? Was it deposited in a basin dominated by regional subsidence, so that the stratigraphy is approximately layer-cake e.g. Rockies (Price 1981)? Or was it deposited in a basin dominated by active faulting so that facies and thicknesses vary markedly e.g. the rift-stage continental margin sequences of the French Alps (Lemoine *et al.* 1986)? With a layer-cake stratigraphy the template can be constructed rapidly, but where thicknesses and facies vary the template should be constructed in a serial fashion, imbricate by imbricate as each successively older thrust is restored, until the basin in which the deformed sequence was deposited is reconstructed.

There are three principal features to consider when restoring the geometry of this basin, marked thickness variations, facies changes and steep ramp angles in the deformed sequence:

(i) Marked thickness changes in the deformed sequence may be due to unrecorded thrust repetition (or thinning; Platt & Leggett 1986) or to original sedimentary thickness variations. Sedimentary facies analysis of the deformed sequence can test these alternative hypotheses as thickness variations are often associated with facies changes.

(ii) Particularly useful are facies that indicate fault activity and variable subsidence rates; for example, thick localised alluvial fans, fan deltas and in some cases submarine fans, or condensed marine sequences (Heward 1981, Surlyk 1978, Bernoulli & Jenkyns 1974).

(iii) Another clue to the geometry of the restored section is the angle of preserved thrust ramps. Fault theory predicts that ramps which develop in flat lying rocks should make an angle of the order of 35° with bedding, (Anderson 1942): although observed angles are commonly lower than this (for examples see Boyer and Elliott 1982). Where ramp angles are steeper, either an early phase of folding or reactivation of earlier higher angle faults can be suspected: these faults represent planes of anisotropy and weakness and may be reused during compression. If these old faults were syn-depositional they may have controlled thickness and facies variations on the restored section. Inversion (that is re-use of faults in a sense opposite to their last movement) of old extensional faults is typical of thrust belts which involve shortening of an old rifted margin (e.g. Lemoine *et al.* 1986).

2.2.2 <u>The Deformed State Section - the synorogenic stratigraphy</u>

In addition to pre-orogenic rocks, the significance of syn-orogenic deposits and their stratigraphic relations must be considered continuously while section balancing. Features such as syn-tectonic unconformities (Riba 1976a&b; Anadon *et al.* 1986), marked changes in palaeogeography and sudden coarse clastic input can assist in defining periods of thrust movement.

As a corollary, the locations of uplift (culminations and large thrust ramps) and the order of their occurrence, as indicated by the thrust sequence shown on the deformed state section, should be compatible with the synorogenic sediments: a section may have to be rejected if it predicts major uplift at a point and place in time where sedimentary facies indicate quiescence.

Bearing these points in mind, the balanced and the restored sections should be constructed simultaneously but iteratively as the stratigraphic template is progressively improved and the likely order of the development of the thrust system emerges. During this stage several sections which are "acceptable" (Elliott 1983) on the basis of structural data may be rejected, or refined, on the basis of incompatibility with either the pre-orogenic or syn-orogenic sedimentary sequences.

The examples presented below {2.6; 2.7} illustrate this methodology. Pre-orogenic and syn-orogenic rocks are treated separately along the Esera valley {2.6}. Although, it is emphasised that the process of section balancing should be iterative and take account of all data at all times.

2.3 THE STRUCTURAL STYLE OF THE SOUTH CENTRAL PYRENEES

The first step in the structural evaluation of an area is to search the available data set and to list the commonly occurring structural styles. The resulting "family of structures" (Dahlstrom 1969) uniquely characterises the study area, and can be used to aid interpretation where data is lacking.

2.3.1 The Data Set

The available data set is wide ranging, but variable in quality. It includes:

(a) Field mapping: around 2,000km^2 have been mapped at a scale of 1:25,000.

(b) Published maps: S.N.P.A. 1972 Carte Geologique des Pyrenees 1:250,000; I.G.M.E. 1:200,000 series, sheets 23 (1970), 24 (1970), 33 (1971) and 34 (1972); Van Hoorn 1970; Seguret 1972; Nijman and Nio 1975; Nijman and Puigdefabregas 1974; Zwart 1979; Van Dijk *et al.* 1985.

(c) Published sections: Seguret 1972; Williams 1985; Camara and Klimowitz 1985; Deramond *et al.* 1985.

(d) Interpreted seismic data - Warburton (pers. comm. 1985)

(e) Borehole data, obtained from B.P. Madrid and ENIEPSA.

2.3.2 The Family of Structures

(a) Decollement

The South Central Pyrenees are characterised by decollement (Seguret 1972). Detachment is localised at Triassic and U. Eocene evaporite, and Paleocene marl horizons which form efficient decollements (Davis and Engelder 1985). These decollements produce laterally extensive thrust sheets in the cover, and form the roof to a culmination in the pre-Triassic basement ((d) below).

(b) Concentric Folding

The major folds are cylindrical. They have straight, subparallel axes (Fig.1.20) and are considered to have developed perpendicular to the compression direction determined by the "bow and arrow" rule (Elliott 1976). The geometry can be interpreted as the product of plane strain. In addition, field mapping and published sections show that bed thickness is maintained around fold hinges, i.e. the folding is concentric. Slickenslides indicate deformation by flexural slip folding (Williams 1985).

Non-cylindrical folds can be explained by features commonly observed in thrust systems: lateral ramps, culminations and tear faults (Butler 1982).

(c) Inversion

The Alpine inversion of extensional faults which bounded an Upper Cretaceous W.N.W.-E.S.E. trending flysch basin, the Vallcarga Basin (van Hoorn 1970) is well documented (Simo *et al.* 1985; Simo 1986).

(d) Antiformal Stack (sensu Boyer and Elliott 1982)

A stack of south-dipping basement imbricates occurs at the northern margin of the study area. This E.S.E.-W.S.W. trending structural unit has been termed the Nogueras Zone (Seguret 1972). The consensus view on its origin follows Williams (1985) that the imbricates form the upper part of an antiformal stack of thrust sheets (Fig.2.1; Camara and Klimowitz 1985; Puigdefabregas *et al.* 1986). However, the author considers that the culmination was a triangle zone (Jones 1982; Washington *et al.* 1987) for much of its history {2.8.3}.

(e) Diapirism

Diapiric salt movements are common within the foreland basin. Two types of salt structures occur: (i) piercement structures where evaporites are intrusive into younger rocks, and (ii) salt-cored folds where stratigraphic relationships are preserved.

2.4 THE STRUCTURE OF THE SOUTH PYRENEAN FORELAND BASIN (S.P.F.B)

2.4.1 Introduction

The S.P.F.B. is dominated by thin, laterally extensive thrust sheets. Locally thrust sheets are well exposed, for example the Cotiella thrust can be seen to repeat stratigraphy over some 20km of ground (Fig.1.20; 2.2(a)). Elsewhere, thrust sheets and thrust geometries have to be inferred by linking isolated thrust fronts and fold zones (e.g. the Montsech thrust; Fig.1.20; 2.2(b)) within the foreland basin. Therein lies the main problem in discussing the structure of the S.P.F.B. - different authors link different overthrusts in different ways.

There are two keys to correctly linking overthrusts: firstly, did the overthrusts develop at the same time and, secondly, do the overthrusts record compatible amounts of shortening at that time? Previous workers have not systematically addressed these problems. However, is only with their groundwork, and the introduction of balanced section techniques, that serious attempts can be made at three-dimensional illustrations of thrust sheet geometries.

Two problems arise with such attempts: the first from the scale of the problem and the resulting necessary incorporation of other workers interpretations in the synthesis, interpretations whose validity is largely unknown. The second problem lies in the exact correlation of events. Because of these difficulties any synthesis can only be an attempted solution, and for it to have any real value the data set on which the interpretations are based must be set out precisely.

At present the terminology of previous workers dominates the literature. For this reason and because it serves to introduce the main structures in the area a short historical review is given below. In section {2.5} the stratigraphy of the pre-orogenic and syn-orogenic sequences are discussed as a building block upon which to develop: firstly, new balanced sections along the Esera {2.6} and Ribagorzana valleys {2.7} and, secondly, an attempted three-dimensional balanced section of the South Central Pyrenean Basin in section {2.8}.

2.4.2 Historical review of thrust sheet terminology in the S.P.F.B

The discussion on south Pyrenean thrust sheet geometry is traced by Sole-Sugranes (1978) back to 1914, but the most influential contributions commenced with the publication of Seguret's thesis in 1972. Seguret (op.cit.) produced numerous unbalanced sections linked to a detailed text and a number of maps which together provide a structural overview of the area (Fig.2.3(a)). Seguret (op.cit) and Choukroune and Seguret (1973) considered that there were two major thrust sheets in the area: the Central Pyrenean unit emplaced in the Lower Eocene/early Cuisian, and the Gavarnie unit emplaced in the middle Lutetian. These two thrust sheets constitute their "higher thrust sheets" which they consider to have been emplaced by gravity gliding {1.4.3}. The Central Pyrenean Unit links the marked overthrust at Cotiella with anothe 120km to the east in the Pedraforca area (Fig.1.20) forming one continuous thrust sheet. Choukroune and Seguret (op.cit) record similar displacement estimates for both overthrusts, which partially justifies this link. The underlying Gavarnie unit derives its name from the major overthrust visible at Cirque de Gavarnie. The southern limit of both thrust sheets is defined by linking the overthrusts seen at the northern margin of the foreland basin to the thrust and folded zone of the External Sierras some 60km to the south. The two thrust sheets are considered to be emplaced intact, although these authors do recognise important hangingwall splays in the thrust sheets, in particular: the Montsech thrust and the Mont Perdu thrust (Fig.1.20). Indeed Seguret (op.cit) shows nearly all the main structural elements of the region (Fig.2.3 (a)iii). However, the geometry is fraught with problems; three problems are immediately apparent: firstly, the link between the thrust at the Cirque de Gavarnie to the deformation in the External Sierras is highly speculative; secondly, the ages of thrust sheet emplacement are problematical in particular, to adhere to Eocene ages of emplacement of these thrust sheets, Seguret (op.cit.) has to consider Oligo-Miocene shortening in the External Sierras to be minor, even though his sections imply several kilometres of shortening, and thirdly, the geometry of the thrust sheets within the foreland basin is not justified - on most of the diagrams their outline is shown dotted.

Seguret (1972) made major contributions to the understanding of the South Central Pyrenees, but in terms of a three-dimensional synthesis only the terms Cotiella nappe, Pedraforca nappe, and Gavarnie nappe can usefully be transferred as they are restricted to the areas of outcrop of overthrusts: the terms Gavarnie unit and Central South Pyrenean unit are not closely defined and should not be used.

Sole-Sugranes (1978) discussed the same structural elements as Seguret (1972) and Choukroune and Seguret (1973), but emphasised them in a different way. He outlined two different thrust geometries, the second being included as a footnote while the text was in proof. The first geometry (Fig.2.3(b)) recognises three major nappes: the Montsech nappe; the Pedraforca nappe; and the "Gavarnie detached mass". He considered the Montsech nappe to be bounded in the west by a hypothetical lateral ramp linking the overthrust at Cotiella to the Montsech thrust fault, and in the east by the Segre fault. The geometry was suggested by Garrido and Rios (1972). The nappe was considered to have been emplaced at the sediment surface by gravity gliding and rapidly buried by continued marl sedimentation during the Cuisian. In this model the Pedraforca nappe is separated from the Montsech nappe by the Segre fault, and thought to have been emplaced during the Biarritzian (Sole-Sugranes 1972). The "Gavarnie detached mass" is similar to, but slightly larger than the "Gavarnie unit" of Seguret (1972); it is considered to be the youngest thrust sheet (late Eocene in age) carrying the complete Montsech thrust sheet in its hangingwall. Again the geometry and timing of the southern parts of the thrust sheets is not closely argued, but recognition of the Segre fault, the Montsech fault and the Coll de Negro fault is important.

The overall geometry is a particularly important one as it formed part of the structural framework for a major sedimentological review of the fill of the Eocene foreland basin (Mutti *et al.* 1975; Nijman and Nio 1975; Puidefabregas *et al.* 1975). The complete framework is illustrated in a block diagram from Seguret (1972; Fig.2.3(a)(iii)). Although Seguret (op.cit) did not articulate it in this way he shows the foreland basin divided into three distinct sub-basins: the Tremp-Graus basin containing the Ainsa basin; the Jaca basin, and the Ager basin {2.5.3}. The first three of these are bounded to the north by the "Central Pyrenees" and are separated from each other by lateral ramp structures. The Tremp-Graus basin is bounded in the east by the Segre fault, and is separated from the Jaca basin by the Boltana anticline (Fig.2.2(c) - a structure considered to correspond to the Mont Perdu thrust at depth. The Ainsa basin is a sub-basin within the Tremp-Graus basin which lies between the Boltana anticline and the western lateral ramp of the Cotiella thrust sheet. The Ager basin developed in the footwall of the Montsech thrust (Fig.2.2(d); it was connected to the Tremp-Graus basin at its western end and closed at its eastern end by the Segre fault.

In his footnote Sole-Sugranes (op.cit) cites "recent" fieldwork and oil well data showing that the Montsech thrust was active largely in the Late Eocene (a date which can be deduced from maps which were published much earlier (SNPA 1972; IGME 1972; Shell 1975)) so that the Montsech and Cotiella thrusts cannot be linked directly during the Cuisian. He suggests that the thrust dies out some km south of Pena Montanesa, and does not link with the Montsech nappe. The Montsech thrust is considered instead to be a hangingwall splay developed in the Gavarnie detached mass of Latc Eocene age. But, assuming that the displacement gradient on the Cotiella thrust is comparable to values obtained from other thrust belts (around 1:10; Elliott 1976), the thrust cannot die out in a tip line as rapidly as Sole-Sugranes (op.cit.) suggests. Furthermore, a balanced section shows that the leading edge of the thrust sheet must extend a considerable distance to the south of the buried thrust {2.6.3(a)}. In summary, the geometry suggested by Sole-Sugranes (op.cit) in his footnote is untenable. The point that the timing of the major movement on the Cotiella and Montsech thrust sheets is not in accordance is, however, an important one and is emphasised in {2.8} where an alternative geometry is suggested.

The work of Williams and Fischer (1984) and Williams (1985) was important not in adding to an understanding of thrust sheets geometries but in introducing new ideas to Pyrenean geology. In particular, the concept of a "linked thrust system" convincingly united basement and cover thrust sheets for the first time, while the introduction of balanced section techniques was the first step towards realistic three-dimensional reconstructions. The suggestion that the detached units should be referred to as thrust sheets rather than as "nappes" a term which has erroneous connotations of major overfolds was also useful. In presenting the first balanced section through the South Central Pyrenees Williams (op.cit.) seems to follow the footnote geometry of Sole-Sugranes (1978) but, confusingly, attributes it to Choukroune and Seguret (1973).

The most complete attempt, to date, at a three-dimensional appraisal of Pyrenean thrust sheet geometries is that produced by Camara and Klimowitz (1985). They propose two distinct, but linked, thrust systems produced by N.E.-S.W. shortening: a system of E.-W. trending faults in the basement (essentially below the Triassic) and a system of N.W.-S.E. trending faults in the cover (Fig.2.3(c)). In the east, to the north of Tremp, the basement system is characterised by an antiformal stack which defines the northern margin of the foreland basin. To the west, in the Jaca region, the basement system is characterised by a hinterland dipping duplex which partly underlies the foreland basin. Imbricates in the corresponding cover system develop first in the east, with movement on the Boixels thrust (the Coll de Nargo fault of Sole-Sugranes 1978) and evolve from north to south and east to west through thrust propagation under frontal and lateral ramps. Camara and Klimowitz (op.cit) consider deformation to commence during the Upper Cretaceous and to continue until the Miocene with the thrust front migrating steadily south without interruption. The model is based upon unpublished borehole and seismic data. Consequently it is difficult to appraise. However, because the model precludes two geometries - a single thrust being active several times and, the switching of the deformation front from external to internal zones - it cannot be accepted without question. The omissions are important as histories of this type can be documented for several major structures: for example, the Montsech thrust was active more than once, while the section lines documented here {2.6; 2.7} indicate switching of the deformation between external and internal zones.

2.5. THE STRATIGRAPHY

2.5.1 Intoduction

The most important conceptual division of stratigraphy in the South Central Pyrenees is the division into: pre-orogenic (pre-Alpine) and syn-orogenic sequences. Over the whole Pyrenees this division is time transgressive since Pyrenean compression commenced slightly earlier in the east than in the west. In the study area Alpine compression commenced during the Maastrichtian. Unfortunately, its effects cannot be correlated throughout the basin.

For correlation purposes a more useful division is the base of the Danian which corresponds to a major transgression that established the fully marine Alveolina Limestone Formation over the fluvial and fluvio-lacustrine Tremp Formation. Both formations are approximately constant in thickness and facies through much of the study area, forming ready marker horizons easily correlated in boreholes and in the field. The constant thickness and facies suggests that they were deposited during a period of tectonic quiescence. Furthermore, they immediately predate the development of the South Pyrenean Foreland Basin {1.3.1} and the main phase of thrusting which created sub-basins within it {2.4.2}. So the geometry of these two formations defines these sub-basins.

2.5.2 The Pre-orogenic stratigraphy

2.5.2(a) Pre-Cambrian to Lower Carboniferous

Sedimentary rocks of Pre-Cambrian to Lower Carboniferous age outcrop in the Axial Zone. They were deformed, metamorphosed and intruded by large granodiorite bodies during the Hercynian Orogeny (Zwart 1979).

Because of this early phase of deformation and because their stratigraphy is poorly known, Alpine structures are difficult to recognise in these rocks, except where a thrust system in the overlying rocks can be linked down into these sequences e.g. Parish (1984). This was not possible in the study area, and the Pre-Cambrian to Lower Carboniferous sequences are shown as undifferentiated basement.

It is felt that a detailed structural study could discern Alpine structures within these sequences, but that the unknown statigraphy would prevent the construction of valid balanced sections.

Tertiary sediments sourced from the Axial Zone are best characterised by unstrained quartz and fresh feldspar derived from the granodiorite bodies, sometimes to the exclusion of other clast types. Cherts and well rounded quartzite pebbles are also likely to have been derived from the Axial Zone, but they may be second cycle detritus reworked from the post-Hercynian molasse sediments.

2.5.2(b) Upper Carboniferous to Triassic

Post-Hercynian molasse sediments unconformably overly Hercynian metasediments at the southern margin of the Axial Zone and form the major component of thrust sheets in the Nogueras Zone. Along the Esera valley these sediments commence with Bunter (Triassic) red beds: to the west and east uppermost Carboniferous and Permian fluvial sediments and volcanics underlie the Bunter (Zwart 1979; I.G.M.E. sheet 23).

Such thickness variations are considered to reflect fault-bounded basins (Fig.1.2; Lucas 1985; Speksnijder 1985). Without excellent exposure and detailed sedimentological work it would be very difficult to accurately restore the Alpine deformation in these sequences. Since exposure is relatively poor such a restoration is not attempted and these red bed sequences are considered, together with the Hercynian metasediments, to be Alpine basement.

The Bunter sandstone is overlain by limestone, marls, gypsum, halite and gabbro (known locally as "ophite"). These lithologies are considered to be Mushelkalk and Keuper in age (I.G.M.E. sheet 23; Zwart 1979). But because of the ubiquitous deformation of these lithologies, through the diapiric movement of the salt and the smearing of evaporites along thrust planes, stratigraphic relationships are rarely preserved. So original thicknesses are unknown and this formation cannot be restored to a pre-Alpine configuration. It too is considered as Alpine basement.

Tertiary sediments sourced from the Upper Carboniferous to

Triassic sequences are best characterised by clasts of Bunter pebble conglomerates which form rounded outsized clasts, and to a lesser extent by brick red siltstones and green volcanics. Where sediment transport paths are short, other parts of the sequence may be distinguished, including reworked gabbro, Muschelkalk limestone and gypsum.

2.5.2(c) Jurassic

Jurassic marine limestones outcrop in the eastern part of the South Central Pyrenees, thinning towards the west and dying out along or to the west of the Esera valley. Following S.N.P.A. (1972) Jurassic rocks are not shown at suface on the Esera valley section line. Other authors suggest that a thin sequence of marine Jurassic may be present in the Rio Esera area (Zwart 1979; I.G.M.E. sheet 23). Further west there are no recorded Jurassic deposits. The author has not studied the Jurassic in detail. However, published maps do not suggest major thickness or facies changes within the Jurassic, suggesting that it was deposited du~~u~~ring a phase of tectonic quiescence.

Clasts sourced from these limestones are difficult to distinguish from limestones of Cretaceous age.

2.5.2(d) Cretaceous

The Cretaceous is characterised by major thickness and facies changes both in a N.-S. sense (Fig.2.4) and in an E.-W. sense (Fig.2.5). The N.-S. changes are fundamental in the development of the restored sections, especially along the Esera valley. So a number of key sections are outlined in detail below. The E.-W. changes are then illustrated by reference to the literature, to allow a regional summary.

The correlations used are those of Souquet (1984) who divides the Cretaceous into a number of depositional cycles: Ca1-6 in the Lower Cretaceous and Cb1-5 in the Upper Cretaceous. He considers the cycles to be controlled by regional tectonics (Fig.2.6). The author is unable to comment on the correlations, having spent, in comparison, only a short time studying a small area. The correlations are useful in the development of the restored section and do not contradict its geometry.

2.5.2(e) North-south changes within the Cretaceous along the Esera valley

(i) Campo Section: the hangingwall to the Cotiella thrust

The Cretaceous sequence north of Campo [8598] is estimated to be 7km thick (Fig.2.4(d)). It commences with a transgressive marine sandstone the San Martin Formation (Zwart 1979) or Turbon Sandstone (Souquet 1984). The sandstone is Albian in age and establishes fully marine conditions which persist until the Maastrichtian.

The San Martin Formation is unconformably overlain by a 1km thick, massive, coarse-grained, bioclastic limestone, interpreted as having been deposited in shallow-marine setting. The limestone is considered by Souquet (op,cit.) to be composed of three formations: the Santa Fe (Cb_1), Reguard and Congost d'Erinya Formations (Cb_2).

The massive limestone passes upwards into thin bedded limestones of the Aguas Salenz Formation (Misch 1934; Cb_3 of Souquet 1984) estimated to be 2.4km thick. Bedding in the limestones is defined by stylolitic horizons, which separate dark and light grey beds around 20cm thick. Some of the beds show infaunal bioturbation. Thin sections reveal: forams, calcispheres and ostracods; a high quartz content (around 15% being angular to subrounded quartz); and a sub-mm sacle lamination primary lamination (Van Hoorn 1970). Van Hoorn (op.cit) interpreted the Aguas Salenz Formation as being pelagic in origin. However, the only feature which the author has seen that bears directly upon depositional depth is the absence of wave ripples. So the deposits are considered to have accumulated below wave base.

Zwart (1979) did not recognise the division between the massive limestone and the thin-bedded limestone, and called the entire sequence the Baciero Limestone Formation, after the Sierra de Baciero G.R.[9140] which forms a steep ridge. Since this ridge is developed in the massive limestone the term Baciero Formation is restricted here to that part of the stratigraphy, and the term Aguas Salenz Formation is retained to describe the thin bedded limestones which overlie it.

The Aguas Salenz Formation is overlain, locally with marked unconformity, by the Vallcarga Group (the Vallcarga Formation of van Hoorn 1970). The Group commences with a stacked sequence of olistostromes known as the Campo Breccia Formation (van Hoorn 1970). The Campo Breccia Fmn. decreases in dip upwards (Fig.2.7(a)) and appears to onlap the unconformity towards the north (the author was unable to reach a position to confirm the second point unequivocally). These relations are interpreted as the product of anticlockwise (when viewed from the east), syn-sedimentary rotation of the unconformity surface about an E-W axis, forming a cumulative wedge system (Riba 1976a,b; Anadon *et al.* 1986). Individual olistoliths are commonly 15m in diameter in the lower part of the sequence, although, exceptionally, they reach 1km in length (Fig.2.7(a); van Hoorn op.cit.).

The Campo Breccia Fmn. is overlain by the Mascarell Formation: together they constitute cycle Cb_4 of Souquet (1985). The Mascarell Formation is characterised by a thinning upwards sequence of turbidites and olistostromes interbedded with marls, and by a number of intraformational slump sheets (Fig.2.7(b); van Hoorn 1970). Palaeocurrents in the Campo Breccia and the Mascarell Formations indicate flow to the east, parallel to the inferred axis of rotation during development of the cumulative wedge system (Fig.2.8). Resedimented rock fragments in the Vallcarga Group are dominantly composed of Upper Cretaceous limestones. But, Triassic sandstone, gabbro and marl clasts are also recorded - they may constitue up to 3% of the rock volume (van Hoorn 1970). A varied ichnofauna occurs throughout the Vallcarga Group (Fig.2.7(c)). Van Hoorn (1970) considers the traces to be representative of a Nereites community (Seilacher 1967) i.e. representative of fully marine deep water conditions.

The Mascarell Formation is overlain by blue-grey marls, the Salas Marls Formation (Cb_5).

The Vallcarga Basin is filled by the westerly progradation of shoreline and fluvio-lacustrine facies, the Aren and Tremp Formations respectively (Fig.2.7(d) Mutti *et al.* 1975; Nagtegaal *et al.* 1983; Sgavetti *et al.* 1984).

2.5.2(e)(ii) The Seira section
the footwall to the Cotiella thrust

The sequence at Seira [8906] lies in the footwall of the Cotiella thrust. It is composed of two parts, a limestone at the base, the Ventamillo Limestone Fmn (Souquet 1967) passing upwards into grey-brown mudstones (Fig.2.4.(c)).

Misch (1934) and Souquet (1967) date this transition as being stratigraphically equivalent to the transition of the Aguas Salenz formation into the Vallcarga Formation (van Hoorn 1970). Souquet (1984) however, dates the transition as equivalent to the base of the Salas Marl Formation, with the Campo Breccia and Mascarell Formations being time equivalents to the Ventamillo Formation, and calls the overlying sequence the Barburens Formation. The sequence underlying the Ventamillo Formation has not been studied by the author, but Souquet (1984) recognises the same lower three formations he records in the hangingwall of the Cotiella thrust i.e. the Santa Fe Cb_1, Reguard and Congost d'Erinya Fm. limestones (Cb_2). In addition, he records bioclastic limestones (Espierba Limestone Fmn.) and sandy calcarenites (Cotiella Macingo Limestone Fmn.) deposited above wave-base of cycle Cb_3 age.

A transition from biomicrites to clean washed bioclasts, including Miliolids, occurs in the top 30m of the Ventamillo Limestone (van Hoorn op.cit). The bioclastic limestone is overlain by sharp-based, quartz-sandstone sheets characterised by hummocky cross-stratification (Harms *et al.* 1975) which mark the base of the Barburens Formation and herald the input of siliciclastic and detrital clay material.

The first sandstones of the Barburens Formation fine and then coarsen upwards. They are overlain by a channelised sandy, bioclastic limestone. The channel fill fines upwards and is overlain by sharp-based, bioturbated, hummocky cross-bedded, fine sandstone beds. The sandstones give way to coarse silts that commence with starved wave ripples but become increasingly monotonous. Upwards they are commonly slumped, and occasionally punctuated by 4cm thick turbidite(?) units.

The change to clean bioclastic limestones and increased importance of Miliolidae at the top of the Ventamillo Limestone Fmn. reflect progressive shallowing (van Hoorn 1970). The overlying sharp based H.C.S. beds and F.U. channel sequence are interpreted respectively as storm deposits in a shoreface setting (for a review see Elliott 1986) and a tidal channel. Together they represent a continued shallowing of the shelf, the product of shoreline progradation, probably linked to an increase in siliciclastic sediment supply. However, the return to sharp-based H.C.S. beds and the gradual transition upwards into wave-rippled silts and monotonous silts punctuated by turbidites is interpreted as the product of shelf deepening, coupled with a steady siliciclastic silt and clay input. The interpretation is supported by the loss of an abundant shallow marine fauna and the increased importance of "pelagic" faunal elements (van Hoorn 1970). The slumps indicate an unstable shelf.

2.5.2.(e)(iii) The Salinas section

A section through the U. Cretaceous 1km north of Salinas on the Rio Cinca, has been documented by van Hoorn (1970; Fig.2.4(c)). It is laterally equivalent to the section at Seira, 22km to the S.E. It also lies in the F.W. of the Cotiella thrust and is particularly useful in that it records the transition to the overlying Upper Maastrictian Aren Sandstone Formation.

The sequence commences with a transgressive lag which reworks the underlying Triassic. The lag is overlain by 100m of limestones and dolomites of Santonian age, the Larra Limestone Formation (Cb_4 of Souquet 1984) interpreted as having been deposited in a high-energy shallow warm sea (van Hoorn op.cit.); it is a time equivalent of the Ventamillo Limestone Formation (Fig.2.6.). The Larra Limestone Fmn. is overlain by 320m of cross-laminated biosparrudites - the Marbore Sandstone Formation (Cb_5 Souquet 1984) - which are 40% angular quartz. Van Hoorn (op cit.) interpreted these sandstones as having been deposited in a shallow tidal sea, and sourced from Hercynian granites.

The sandstones are overlain by 80m of calcareous mudstones which underly the Aren sandstone. The fossils and low energy facies in the mudstones indicate deposition in a shallow non-agitated sea.

2.5.2.(e)(iv) Aguinaliu

A section at Aguinaliu [8064] (van Hoorn op.cit.; Fig.2.4(a)) is representative of the Upper Cretaceous in the External Sierras between the Esera and the Ribagorzana rivers (Selzer 1934). It commences with a transgressive basal lag which reworks the underlying Triassic. The bulk of the sequence is composed of rudist limestones (185m) dated as Campanian in age and interpreted as shoal reefs, deposited in clear, shallow (30-60m) warm, open-marine conditions.

The limestones are directly overlain by red marls of the Tremp Formation: there is no Aren Sandstone equivalent in the External Sierras.

Work by the author in the Esera gorge [7667] shows that the Tremp Formation is dominated by thick lacustrine limestones. The limestones exhibit well developed "crumb textures" indicating periodic desiccation (Freytet 1973) and are interbedded with silty palaeosol horizons. The overlying Alveolina Limestone Formation is characterised by large scale cross-bedding >3m.

2.5.2(f) East-West variations in the Hangingwall of the Cotiella thrust

Van Hoorn (1970) and Souquet (1984) have attempted to synthesise east-west stratigraphic variations in the hangingwall of the Cotiella thrust. Both authors show marked thickness and facies variations along strike, especially within the Vallcarga basin, and consider that the changes are due to faulting (Fig.2.5)

However, there is no agreement either on where thickness changes occur, or on fault orientation. Souquet (op.cit.) postulates that the faults were orientated E.N.E.-W.S.W. having the same orientation as the faults which he believes developed during the Late Jurassic-Aptian extensional phase {1.2.1}. But the author knows of no evidence for this fault orientation during the Upper Cretaceous in the South Central Pyrenees. By contrast, van Hoorn (op.cit.) emphasises facies changes which occur to the east of the Esera across a N.E.-S.W. trending line, not only in the Cretaceous but also in the Devonian and Jurassic, and he inteprets them as having been controlled by faults localised above a long-lived basement lineament. It may be possible to extend this line of reasoning along the length of the basin, but at the moment the orientation of other faults must remain speculative {but see 2.8.3(iii)}.

2.5.3 The Foreland Basin

The post-Alveolina limestone stratigraphy represents the fill of the South Pyrenean foreland basin {1.3.1.}

The stratigraphic nomenclature and ages adopted here are those of Mutti *et al.* (1985a) (Fig.2.9) who follow the Palaeogene Numerical Time Scale of Hardenbol and Berggren (1978). However, some of the formation names used by Nijman and Nio (1975; Fig.2.10) are retained since they have special significance with respect to the NNE-SSW structural sections presented here.

The most complete review of the basin fill is given in an excursion guidebook edited by Rosell and Puigdefabregas (1975). They subdivided the foreland basin into three parts {2.4.2}: (1) the Tremp-Graus basin in the east, containing the Ainsa sub-basin, documented by Nijman and Nio (1975); (2) the Jaca basin in the west, documented by Puigdefabregas *et al.* (1975) and (3) the Ager basin south of the Tremp-Graus basin, described by Mutti *et al.* (1975). More recent but less comprehensive reviews are given by Friend *et al.* (1981); Atkinson (1983); Mutti *et al.* (1985a,b); and Puigdefabregas and Souquet (1986).

The fill of the Tremp-Graus basin is important in constraining the structural sections presented here. So it is discussed in some detail below. The fill of the Ager basin is important only in constraining the Ribagorzana valley section and so is discussed elsewhere {2.7}.

Nijman and Nio (1975) concentrated on the Paleocene, Lower and Middle Eocene parts of the Tremp-Graus basin. They document an axial sediment dispersal system that passes from fluvial sediments in the east via shallow marine and slope facies to deep water turbidites in the west (Figs.2.10; 2.11). Throughout the Palaeogene the position of the slope facies remained approximately constant, despite changes in the rate of sediment supply and in sea-level. The fixed nature of this facies belt led Nijman and Nio (op.cit.) to postulate that its position was tectonically controlled by the lateral ramp of the Cotiella thrust sheet which Garrido and Rios (1972) had shown as passing through this area {2.4.2}. Nijman and Nio (op.cit.) also recognised that the deep-water facies west of the slope thinned and passed laterally into shallow water carbonates towards the present day Boltana anticline. They considered this structure to have been present from the Cuisian onwards partially separating the Tremp-Graus basin from the Jaca basin further to the west. The Boltana anticline also defined the western margin of the Ainsa sub-basin, whose eastern margin was the lateral ramp of the Cotiella thrust sheet.

Mutti *et al* (1985a) also use these divisions, recording channel-fill turbidite bodies and thick mudstone units to the east of the Boltana, and non-channelised sandstone lobes to the west of the Boltana. In addition, they recognise two sectors within the Jaca basin characterised by a down system change to basin plain turbidites west of Jaca. Mutti *et al* (op.cit) date the main folding of the Boltana anticline as being late in the fill of the S.C.P.B. - after the deposition of the Banaston sequence (Fig.2.9). However, this scheme is in a state of flux; Mutti (pers.comm. 1987) currently recognises 5km wide channels in the Jaca basin.

The fill of the Tremp-Graus basin is complicated by two factors: localised fanglomerate bodies sourced directly from the Pyrenean margin and sea level changes. The fanglomerate bodies are of particular importance as they can be used to infer periods of uplift along NNE-SSW balanced sections.

A number of discrete fanglomerate bodies have been recognised at the northern margin of the basin (Nijman and Nio 1975; Atkinson 1983; Cuevas *et al.* 1985). The largest of these, the Campanue Conglomerate Formation, is centred on the Esera valley and transected by the first balanced section (2.6). The Campanue Conglomerate is modelled as a fan-delta (Nijman and Nio 1975; Crumeyrolle and Mutti 1986). 600m of stacked conglomerates occur in the most proximal part of the fan (Fig.2.12(a)). They overlie the Castisent depositional sequence with a low-angle unconformity that dies out towards the basin axis. The conglomerates pass laterally into an extensive sandy haloe composed of granitic sands, reworked by basin processes into high energy shoreline and tidal channel deposits (Crumeyrolle and Mutti 1986) which form part of the Perarrua Formation (Fig.2.12(b); Nijman and Nio 1975). Conglomerate clasts indicate a source area containing U.

Cretaceous limestones, the Mascarell Formation, the Alveolina Limestone Formation, and Hercynian Granitoids. Hercynian metasediments and Permo-Triassic clasts are absent or very rare, while palaeocurrents indicate flow to the S.E. (Crumeyrolle and Mutti 1986).

The fill of the Tremp-Graus basin along the Ribagorzana valley is not well documented in the literature. Parts of the fill have been studied in detail (Puigdefabregas and Van Vliet 1978; Friend *et al.* 1981); some palaeogeographic maps have also been published, showing sediment input from the north-east and south-east during the Eocene (Nijman and Nio 1975; Fig.2.11) and Atkinson (1983; Fig.2.13). But the author could not find a sedimentological section along the valley from which to infer basin margin uplift on the basis of fan progradation and/or thickness variations. The only sections which show the Ribagorzana valley run east-west along the basin axis (e.g. Nijman and Nio 1975; Fig.2.10). These sections suggest that there was no major fan body at the northern margin of the basin along the Ribagorzana valley. However, the absence of such a body is thought to reflect a long-lived sediment input point immediately to the west of the Ribagorzana valley at the Sierra del Sis, rather than the absence of uplift at the northern basin margin along this section line. The input point is located in a syncline at the basin margin below the Sierra del Sis (Fig.1.20; {4.4.1}). It produced coarse sediment bodies throughout the Tertiary (Fig.2.10(a)) incuding the Roda sandstone during the Figols depositional sequence and the San Esteban Conglomerate during the Figols and Montanana depositional sequences (Puigdefabregas 1975; Nio and Siegenthaler 1978; Atkinson 1983; Cuevas *et al.* 1985; Puigdefabregas *et al.* 1985). A north-south section through the input point shows a marked increase in fan activity at the end of the Castisent depositional sequence (Fig.2.14). The fans extend up to 15km into the basin and would have required a catchment area of greater radius (Bull 1964a) easily encompassing the Ribagorzana section. Therefore, variations in sediment input from the Sierra del Sis would, in part, record uplift along the Ribagorzana valley. The timing of the increase in sediment input is closely comparable to that recorded by the Campanue

Conglomerate along the Esera valley. Indeed, Crumeyrolle and Mutti (1985) suggest that the Sierra del Sis input point was an important contributor to the Campanue Conglomerate.

In summary, there is evidence for a marked increase in uplift at the northern margin of the basin at the end of the Castisent depositional sequence along both the Esera and Ribagorzana valleys.

The northern fans formed only one component of the sediment dispersal system. The majority of the palaeocurrents suggest flow along the basin axis to the west: a third component are the N.W. directed palaeocurrents in the Montllobat and Castisent depositional sequences recorded at the southern margin of the basin (Figs.2.11; 2.13). The Pyrenean provenance of these sandstones (Atkinson 1983) suggests that thrust-generated relief to the south of the preserved basin margin must have existed to turn south flowing fluvial systems northwards.

Mutti *et al.* (1985a) consider the major sequences in the South Central Pyrenean basin to be the product of sea-level changes. Using aerial photographs they trace unconformity and disconformity surfaces from the shelf into the basin. They define two distinct types of <u>depositional sequence</u> (Fig.2.9.): (i) sequences characterised by shelf sedimentation, and channel-levee deposits in the deep basin; and (ii) sequences characterised by shelf edge entrenchment, producing submarine canyons and large scale slump scars. Entrenchment supplied large volumes of coarse sediment to the deep basin where it accumulated as thick sandstone lobes.

The first type of sequence is considered to be the product of a high sea-level stand: coarse sediment is held on the drowned shelf in a deltaic setting. The second type of sequence is thought to be the product of a low sea level stand: lowering of sea level is considered to initiate subaerial and/or submarine erosion.

The depositional sequence stratigraphy can be closely married to the lithostratigraphy developed by Nijman and Nio (1975; Fig.2.15). But without aerial photographs it is not possible to give an accurate assessment of this work: no comment can be made on the vital correlations between the shelf and the Ainsa basin. But, some reservations can stated. Firstly, because of erosion over the crest of the Boltana anticline the Jaca and Ainsa sub-basins are not now physically connected, so that the correlation between these basins suggested by Mutti *et al.* (1985a) must be treated with caution. Secondly, it seems unlikely that tectonics played no role in the development of these sequence boundaries. In particular, the change in source and rapid progradation associated with the Castisent sandstone is thought to have been tectonically triggered.

2.6 A BALANCED SECTION ALONG THE ESERA VALLEY

2.6.1 Introduction

The section extends from the Axial Zone above Villanova G.R. [9114] southwards along the Esera valley to the northern margin of the Ebro basin S.E. of Barbastro [6453] i.e. from the most northern exposure of Alpine cover rocks to the southern limit of Alpine deformation, including the entire deformed part of the South Pyrenean foreland basin. The section is orientated N.N.E.-S.S.W. parallel to the direction of tectonic transport {1.2.3}. It was chosen as a line to which balanced section techniques had not previously been applied: the production of a balanced section being considered a significant step in the the explanation of the development of the area. Previous workers laid the foundations for the balanced section in providing important regional tectonic summaries (SNPA 1972; Seguret 1972; IGME sheet 23) and sedimentological and stratigraphic data (Van Hoorn 1970; Nijman and Nio 1975; Mutti *et al.* 1985; Souquet 1985). In addition to the family of structures which can be deduced from this literature {2.2.2} a number of features specific to the section are apparent:

(i) The Cotiella thrust. This thrust repeats the Upper Cretaceous and Paleocene stratigraphy to the west of the section line, where the map distance from the Alveolina Limestone Fmn. hangingwall cut-off below Pena Montanesa 10km N.E. of Ainsa to its equivalent footwall cut-off N.W. of Plan in the Cinca valley, suggests a minimum displacement of 20 km (Fig.1.21). Since displacement gradients on thrusts are low - Elliott (1976) suggests 1:10 - and because there are no intervening structures the Cotiella thrust must be present along the section line with at least 18km of displacement. Since it is unknown whether displacement increases or decreases from Pena Montanesa towards the Esera valley, the section assumes that the leading edge of the Cotiella thrust sheet is translated 20km to the south.

(ii) A marked decrease in the thickness of the Cretaceous and associated marked facies changes from north to south {2.5.2}

(iii) A broad area of gently folded Paleocene to Middle Eocene, siliciclastic, shallow-marine sediments in the central part of the section.

(iv) A zone of folded and thrust Triassic to Miocene sequences and a major W.N.W.-E.S.E. trending evaporite cored fold at the southern end of the section, the Barbastro anticline.

2.6.2. The Development of the Stratigraphic template: the restored section.

The first question to ask in the development of the restored section is: in what basin setting were the pre-orogenic sediments deposited? {2.3}. Along the section line, pre-orogenic sediments are Upper Cretaceous in age. Lower Cretaceous and Jurassic rocks are absent, while older rocks are considered to be Alpine basement {2.4.1.}. A review of Pyrenean literature suggests that Upper Cretaceous basins were dominated by transtensional strike-slip {1.2.2.}. So, rapid thickness and facies changes due to faulting are to be expected on the restored section.

Construction of the stratigraphic template commences by drawing the topography prior to contraction. In some studies this may be complex, as with the fault scarps shown by Simo *et al.* (1985 Fig.2.23): along this line of section it is simple, the Alveolina Limestone and Tremp Formations are constant in thickness and facies and so restore to the horizontal, where they can be used as a "regional" or reference level on which to develop the template.

There are two main points which assist in the further development of a stratigraphic template (1) the identification of a steep, long hangingwall ramp within Upper Cretaceous rocks above the Cotiella thrust and (2) the recognition of a marked change in the thickness of the Upper Cretaceous from the hangingwall (the Campo section {2.5.1.(e)(i)}) to the footwall (the Seira section {2.5.1.(e)(ii)}) of the Cotiella thrust.

The steep hangingwall ramp was first shown on a series of unbalanced sections by Seguret (1972; Fig.2.16). The ramp extends eastwards from Pena Montanesa and must occur on the section line at depth {2.4} where its geometry can be constrained in three ways: firstly, by the low angle of the Cotiella thrust (Seguret 1972; S.N.P.A. 1972; I.G.M.E. sheet 23); secondly, by the surface dips of units in the hangingwall and footwall of the Cotiella thrust, and thirdly, by the requirement that the section balances.

The steep hangingwall ramp, and the marked increase in thickness from footwall to hangingwall, suggest (1) the inversion of an old steep fault to explain the steep ramp angle and (2) active faulting during the Upper Cretaceous to explain the observed thickness variations. A model along these lines can be developed by attempting to join the hangingwall cut-off in the Santa Fe, Reguard and Congost d'Erinya Formations. (Cb_{1-2}) south of Seira G.R.[8705] with their equivalent footwall cutoffs north of Chia G.R.[9213] (for locations see Figure 1.20): the Cotiella thrust sheet has to be pulled back 15km but this still leaves 5km of U. Cretaceous rocks above the regional. An extensional syn-sedimentary fault is postulated to explain this (Fig.2.17.(a)). The model makes use of the convention of drawing sections that show the minimum values of shortening: an alternative is to have an Upper Cretaceous basin which thickens northwards gradually and is then thrust through at a high angle some distance to the north of the preserved footwall cutoff, which, in fact, is a cut off against a backthrust. The models can be tested by examining the hangingwall (Fig.2.4(d)) and the footwall (Fig.2.4(c)) stratigraphies of the Cotiella thrust above the Santa Fe, Reguard, and Congost Formations which are considered to be equivalent in time and facies across the thrust.

The hangingwall sequence commences with the 2.5km thick Aguas de Salenz Formation Cb_3, interpreted as having been deposited below wave base. It contrasts with bioclastic limestones and sandy calcarenites in the footwall deposited above wave base (Souquet 1984). Since the top of the Congost Formation Cb_2 and the base of the Ventamillo Limestone Fmn. Cb_4 were not picked in the field the thickness of the bioclastic limestone is not accurately known: it is estimated to be <500m thick, one fifth, or less, of the thickness of the Aguas Salenz Formation. The thickness and facies changes from hangingwall to footwall are compatible with a half-graben geometry, but the general absence of fault-scarp related features suggests that a marked scarp did not develop, perhaps because sedimentation kept pace with subsidence. Souquet (1984) shows a similar transition from the

Aguas Salenz Formation eastwards to the Turbon and also postulates a syn-sedimentary fault to explain the facies and thickness differences (Fig.2.5(a)).

The overlying cycle Cb_4 is composed of the Campo Breccia and the Mascarell Formations in the hangingwall and the Ventamillo Limestone Fmn. in the footwall. The Campo Breccia and the Mascarell Formations are the cornerstone of the half-graben model.

The size, composition and angularity of the clasts in the Campo Breccia suggest that it is the product of fault breakup of a carbonate platform. Van Hoorn (1970) suggested that faulting was the result of diapirism, with diapiric rise of Keuper evaporites taking other Triassic lithologies to the surface (Fig.2.18). There are structures in the area similar to this "diapir", for example the Turbon anticline {2.8.3(a)(i)} but there is no direct evidence for the model, and van Hoorn (op.cit) is unclear about the location and orientation of the structure which he postulates. Furthermore, his model does not explain the marked difference in thickness between the hangingwall and the footwall of the Cotiella thrust.

The sequences are compatible with the half-graben model. The graben bounding fault could have triggered olistoliths and would have needed a throw of only around 500m to expose Triassic lithologies and so explain the observed clast types. Decreasing fault activity would have produced the gradual decrease in the supply of olistostromes. The fault is postulated to have been orientated W.N.W.-E.S.E. parallel to palaeocurrents in the Campo Breccia Fmn and in the overlying Mascarell Formation.

The Mascarell Formation is also compatible with the half-graben model: slumps and turbidites indicate a tectonically active region.

The Ventamillo Limestone Fmn. contrasts markedly with the Campo Breccia and the Mascarell Formations It is interpreted as having been deposited at shallow water depths on a stable shelf and is, at the most, 500m thick - some four times thinner than the hangingwall sequences.

So, as during cycle Cb_3, the half-graben model is strongly supported by facies and thickness differences between the footwall and hangingwall of the Cotiella thrust during cycle

Cb_4.

The final Cretaceous cycle Cb_5 also shows a number of changes between the hangingwall and footwall of the Cotiella thrust. But they cannot be related to the half-graben model.

In the hangingwall the cycle is represented by the 1km thick Salas Marl Formation and the Aren and Tremp Formations. The absence of slumps and turbidites in the Salas member (along the section line) is interpreted to record a period of quiesence, and low rates of clastic sediment supply. The regressive Aren and Tremp formations which complete the cycle are the final fill of the Vallcarga basin, with the Tremp Formation recording subaerial exposure.

The footwall is around 250m thinner, indicating that subsidence was still greater in the hangingwall but that the difference in subsidence rates had greatly reduced. The footwall commences with disconformities and a shelf deepening sequence overlain by slumps and thin bedded turbidites of Campanian age. Unfortunately the cycle is incompletely preserved: its top has been removed in the hangingwall of the Cotiella thrust. However, the Cotiella thrust cuts up section laterally - 10km along strike to the north above Barburens G.R.[8609] (see Figure 1.20 for location) - so that the complete cycle is preserved, allowing an estimate for the thickness of the removed sequence to be made (500m to the base of the Alveolina Limestone Fmn.). The complete footwall sequence has been described by van Hoorn (1970) north of Salinas {2.4.1.(e)(i)} some 22km N.W. of Seira. The sequence near Salinas is interpreted as shallow marine thoughout. In this respect it differs from the Seira sequence. However, it is approximately the same thickness, and the difference in facies is considered to be the result of high rates of sediment supply in the Salinas area. So shelf sedimentation is considered to have persisted in the Seira area in the Campanian. Both the Salinas section and the section 10km north of Barburens are capped by a regressive upper Maastrichtian sandstone (van Hoorn 1970; I.G.M.E. sheet 23) an Aren Sandstone Fmn. equivalent, in turn overlain by the Alveolina Limestone Fmn. Both of these formations are considered to occur on the section line.

Thus the footwall and the hangingwall sequences are similar during Cb5, differing only in the enhanced subsidence of the hangingwall and in the shelf deepening, instability and higher rates of sediment supply in the footwall. The model of a half-graben is not applicable at this stage - differential regional subsidence is preferred.

In summary, the sequences in the hangingwall and footwall are compatible with the restored section, the facies indicate: firstly, a long-lived depositional depth contrast between the hangingwall and footwall sequence during Cb_3 and Cb_4, and secondly, periodic fault movement during Cb_3 and Cb_4 especially at the start of Cb_4 with the initiation of the Campo Breccia Fmn. after which time faulting decreased gradually in intensity. Outside of cycles Cb_3 and Cb_4 Upper Cretaceous sequences in the hangingwall and footwall of the Cotiella thrust are similar, indicating that the postualted half-graben had not developed by Cb_1 and Cb_2 times, and that it had become inactive by Cb_5.

During compression part or all of the syn-sedimentary fault was reactivated as a reverse fault i.e. it was inverted.

2.6.3 The deformed state section

The deformed state section (Fig.2.17(e)) has three key features (1) the Cotiella thrust sheet emplaced into the foreland basin during the Late Paleocene and Early Eocene; (2) the development of a triangle zone at the northern margin of the foreland basin from the Middle Eocene to the latest Oligocene; and (3) a switch of the thrust front back to within the foreland basin during the latest Oligocene with the emplacement of a second major thrust sheet.

2.6.3 (a) The Cotiella thrust

From regional constraints the leading edge of the Cotiella thrust sheet must exist along the section line {2.4.2; 2.6.1}. The leading edge is considered to be represented by a buried high in the Palaeogene foreland basin. The reasoning is complex, there are two factors.

The first factor is the absence of evidence for exposure of the Cotiella thrust front along the section line south of Seira. Where the thrust is seen at Seira the footwall decollement is at the level of the Salas Marl member (Cb_5), the hangingwall decollement is at the base of the Cretaceous but is considered to be about to rapidly climb stratigraphy on a steep ramp (2.4.2). So, without recourse to thrusts cutting down section in the transport direction, or early deformation, the geometry of the Cotiella thrust front would be latest Cretaceous (or younger) thrust over latest Cretaceous (or younger). No such thrust front is seen within the basin. Thus, the leading edge of the Cotiella thrust sheet must be buried. The first thrust front to the south brings Triassic and Jurassic rocks to the surface, indicating that a second major thrust underlies the Cotiella thrust.

The second factor is derived from balanced section constraints. A long Upper Cretaceous footwall flat occurs in the Ventamillo Limestone (Cb_4) and Barburens Formations (Cb_5) north of Seira - a section of equivalent length of the Aren Sandstone and Alveolina Limestone Formations must have been removed in the hangingwall of the Cotiella thrust. So the leading edge of the Cotiella thrust sheet must have extended southwards of its surface outcrop at Seira into the Tremp-Graus basin (Fig.2.17(e)). The Centenera borehole in the Tremp-Graus basin penetrates to the Triassic but does not record a repetition of the Tremp and Alveolina Limestone Formations. So a ramp in these formations by which the Cotiella thrust could have broken to the palaeo surface must be to the south of this borehole. Following the assumption of 20km of displacemnent on the Cotiella thrust, the leading edge (the equivalent hangingwall ramp) of the thrust sheet must extend 20km to the south of the Centenera borehole. Unfortunately the Graus borehole is not deep enough to test this model. The Benabarre 3 borehole is - like the Centenera borehole it shows no repetition of the Tremp and Alveolina Limestone Formatins, and also penetrates to the Triassic. So the Cotiella thrust does not extend as far south as this borehole - a minimum of 4km of erosion of the leading edge of the the thrust sheet must be postulated for the presented model to hold (the section shows 9km of erosion).

The high itself is represented by a fold in the Alveolina and Tremp Formations. It has been identified by picking these formations in the Graus, Campanue and Centenera boreholes (located on Fig.1.20). The effect of the fold cannot be seen at surface in the Middle Eocene Perarrua Formation or in its lateral equivalent, the Campanue Formation, giving an upper age limit for its formation. The Alveolina Limestone Formation gives a lower limit for its development. The Alveolina Limestone Formation hangingwall is approximately constant in thickness and facies along the section line, indicating that it was deposited during a period of tectonic quiesence: after the Upper Cretaceous extensional phase and prior to Alpine compression. Paleocene-Early Eocene sediments thin on to the high. Unfortunately, their exact relations with the high are unknown, so that the date of emplacement cannot be refined beyond this. The initiation of inversion may be subjectively related to an olistostrome of Late Paleocene age, composed of resedimented Alveolina Limestone blocks that occurs at the base of the Figols depositional sequence.

The geometry, that of a buried thrust front, is compatible with the published maps (SNPA 1972; Seguret 1972; IGME sheet 23) which show the Cotiella thrust being buried by Cuisian (Lower Eocene) sediments south of Pena Montanesa (Figs.1.20; 2.2(a)). The recently developed correlations of Mutti *et al.* (1985a) which trace depositional sequences across the trend of the lateral ramp of the Cotiella thrust, suggest that movement on the ramp had ceased by the time of deposition of the Castisent depositional cycle, refining the age of thrust sheet emplacement to the Figols and Montanana sequences (Thanetian and earliest Ypresian).

Incorporation of the Upper Cretaceous graben-fill into the hangingwall of the Cotiella thrust suggests either, that the graben fill was plucked from the graben with the basement retaining an extensional geometry, or that the graben was inverted and the Triassic in the hangingwall was elevated to the level of the Salas Marl Formation in the footwall prior to detachment in the cover. Both suggested geometries appear unlikely - major deformation may be expected as the graben fill passes over the apex of the Upper Cretaceous extensional fault. However, two points can be brought to the support of the second geometry which is shown in Figure 2.17(a,b,c)): firstly, the graben is not undeformed, faults thought tohave originated as synthetic normal faults during extension are reactivated as thrusts during compression, and secondly, the angularity of the apex reflects a geometry which gives the minimum amount of shortening. If, as suggested elsewhere {2.8.3(b)} there was some distance between the backthrust cutoff and the hangingwall cutoff of the Cb_{1-2} limestones in the Upper Cretaceous extensional fault, then a more gentle profile could be accommodated, albeit with a slightly more complex geometry.

Figure 2.17(b) is schematic - it shows the preferred model of graben inversion but does not take the foreland basin fill into consideration. The geometry of Figure 2.17(c) and other partially restored sections (Figs.2.17(d); 2.20(b-e)) is controlled by taking the contemporaneous basin fill as a regional, or horizontal reference line.

2.6.3.(b) <u>The development of the triangle zone</u>

The trailing edge of the Cotiella thrust sheet is folded to dip towards the south. In addition, it is truncated by a major thrust which also cuts earlier southward directed thrusts within the basement and its Permo-Triassic cover (Fig.2.17(d)). A duplex is postulated to explain these features, with the shortening being taken up by a backthrust - a triangle zone in the sense of Jones (1982). The backthrust is not a simple passive roof thrust: it is considered to have been folded during the development of the basement imbricates and to have been simplified during continued thrusting, so that the backthrust developed an extensional geometry. The triangle zone developed later than the Cotiella thrust as it folds the thrust and causes its trailing edge to be truncated.

The emergent thrust front is now the backthrust to the triangle zone: it has switched from the foreland basin to the northern margin of the basin, and the Cotiella thrust has become inactive.

The Campanue Conglomerate Fmn. {2.5.3} is thought to record this switch; it is considered to be the product of the development of the triangle zone: its scale, northern provenance, and the unconformity which underlies it are all indicative of major uplift at the northern margin of the foreland basin. Furthermore, the Campanue Formation and its lateral equivalent, the Perarrua Formation, are the first formations that are clearly not folded by the emplacement of the Cotiella thrust, so that the uplift inferred by the conglomerate occurred after the emplacement of the thrust sheet.

A further virtue of explaining the Campanue Conglomerate in this way is that the backthrust continuously uplifts Paleocene and Cretaceous rocks in the source area of the fanglomerate body: although older rocks are deformed at depth they are not thrust directly to the surface, so helping to explain the dominance of Cretaceous and Paleocene clasts within the conglomerate body.

A marked thinning of the Paleocene-Eocene sequence occurs at the southern margin of the Tremp-Graus basin, from surface outcrops along the Esera valley to the Benabarre 3 borehole. The thinning is interpreted as the product of the pillow stage of the La Puebla diapir. Diapirism is considered to have been triggered by the loading effect of the emplacement of the Cotiella thrust sheet (c.f. {3.8.1}) so that the Perarrua and Capella Formations unconformably overlie the Montanana and Figols depositional sequences and thin towards the diapir. The geometry cannot be demonstrated directly, but is supported by thinning of the overlying Pano and San Martin sequences towards the La Puebla diapir {3.6.4} and by the Triassic core of the La Puebla diapir{4.4.3}.

2.6.3.(c) <u>Emplacement of thrust sheet 2</u>

The W.N.W.-E.S.E. band of folded and thrust Triassic to Miocene sequences which comprise the External Sierras at the northern margin of the Ebro basin, is considered to be the product of a second major thrust. The thrust underlies the Cotiella thrust and contains the Cotiella thrust entirely in its hangingwall {2.6.3.(a)}. This second major thrust is termed thrust 2. Its extent in the foreland basin is comparable to the Gavarnie unit of Seguret (1972). However, the author does not link the thrust in the foreland basin to the thrust at the Cirque de Gavarnie (as Seguret (op.cit) does) and, therefore, considers the term "Gavarnie unit" inappropriate. Futhermore, the timing of thrust movement and the detailed geometry of thrust sheet 2 is different to that suggested by Seguret (1972).

The deformed state section shows the External Sierras allochthonous upon an Upper Eocene evaporite horizon (Fig.2.17(e)). The geometry is driven by the need for compatibility with sections along strike: the seismically controlled sections of both Camara and Klimowitz (1985; Fig.2.19) and Hale (Warburton pers.comm. 1985) show such a geometry. In addition to this allochthonous geometry, Camara and Klimowitz (op.cit) imply a diapiric origin for the Barbastro anticline, with the Upper Eocene evaporites thickening markedly towards the anticline core. A similar geometry is shown by Riba *et al.* (in prep., Fig.1.10(b)).

There is no surface evidence for the allochthonous nature of the External Sierras along the section line. However, small scale structural features do indicate that the Barbastro anticline is a type of diapiric structure and an attractive model whereby differential loading following emplacement of thrust sheet 2 initiated diapirism can be developed {3.8.1(b)}.

Timing the emplacement of thrust sheet 2 is problematical. There is some evidence for Upper Eocene-Oligocene deformation, but the majority of the evidence points to an Uppermost Oligocene to Miocene age for thrust sheet emplacement.

Evidence for an Upper Eocene-Oligocene age is two-fold. The first point comes from the balanced section: the absence of Oligocene and Miocene hangingwall sequences which correspond to the Upper Eocene evaporite footwall flat shown on the deformed state section, suggests that thrust sheet 2 was emplaced on an Upper Eocene land suface. Alternatively, these sequences could have been eroded. The second line of evidence is derived from thick conglomerate sequences seen to the west of Naval at Cerro Comun [6975] (100m+ thick) and at Mipanas [6976] (30m+ thick; Fig.2.21). Both conglomerate sequences underlie Escanilla Formation sediments of Upper Eocene age (Fig.2.15). They are similar in lithostratigraphy and provenance - both sequences are dominated by unsorted debris flows, composed of large (up to 3m in diameter) subangular clasts derived from locally outcropping lithologies - so that they are considered to be part of the same conglomerate body, interpreted as the proximal part of an alluvial fan of Upper Eocene age. The thickness and inferred extent of the conglomerate body implies major uplift, as do the W.N.W.-E.S.E. folds which the Upper Eocene sediments overlie. Uplift could be the product of diapirism or thrusting.

Evidence for uppermost Oligocene to Miocene deformation is exhaustively documented in chapter 4. Marked changes in paleogeography, local generation of sediment sources and stratal wedging relationships occur during the Oligo-Miocene throughout the External Sierras: features which are all considered to be the product of emplacement of thrust sheet 2. At some localities angular relationships indicate that all the observed deformation is Oligo-Miocene in age {4.4.3(b)}.

In summary, the second thrust sheet is considered to have been emplaced during the latest Oligocene to Miocene. Although some movement could have occured during the Upper Eocene. The Barbastro anticline is thought to be a post-Miocene feature: it deforms Oligocene fluvial sediments, but has no noticeable effect on palaeocurrents or architecture in Miocene sediments on either side of the structure (Hirst 1983) {4.5}.

2.6.4 <u>Summary</u>

The Esera valley section line records: (i) development and fill of a north-facing half-graben during the Upper Cretaceous; (ii) quiesence during deposition of the Alveolina Limestone and Tremp Formations; (iii) inversion of the Upper Cretaceous half-graben and emplacement of the Cotiella thrust sheet into the foreland basin during Figols/Montanana (Paleocene/Lower Eocene) depositional sequences; (iv) a switch of the thrust front from within the foreland basin to the northern margin of the basin during the Middle Eocene, the thrust front becomes a roof thrust to a triangle zone which causes uplift at the northern margin of the basin and produces the Campanue fanglomerate body. The triangle zone continued to develop until the Oligo-Miocene, when (v) the thrust front switched back to within the foreland basin, with the emplacement of a second major thrust sheet.

2.7 A BALANCED SECTION ALONG THE RIBAGORZANA VALLEY

2.7.1 Introduction

The section extends from the Axial Zone north of Pont de Suert southwards along the Ribagorzana valley to the northern margin of the Ebro basin. As with the Esera section {2.6} it traverses the entire deformed part of the Tertiary foreland basin.

Several workers have produced sections along the valley, but they have tackled the problem from a structural viewpoint (Seguret 1972, Fig.2.3(a); Williams 1985, Fig.2.22; Camara and Klimowitz 1985, Fig.2.19; Hale (pers.comm. J. Warburton 1985). The linked sedimentological and structural approach adopted here {2.2} has allowed a new section to be constructed from the literature.

The section is discussed firstly by highlighting those parts of the data set which constrain it most closely and, secondly, by describing five sequentially restored sections.

2.7.2 The key features of the section

Five key points constrain the section: seismic data; boreholes; published sections in the Nogueras Zone; the sedimentological fills of the Ager and Tremp-Graus basins; and the varied nature of the Upper Cretaceous. The fill of the Tremp-Graus basin is addressed in {2.5.3} above, the other points are discussed below.

(a) Seismic data

The author was unable to obtain uninterpreted seismic sections through the area. However, B.P. Madrid were able to supply interpreted sections produced by N.Hale (1985) based on seismic data. J. Warburton (pers. comm.) emphasised that the seismic data was of poor quality, with the exception of two major reflectors beneath the foreland basin extending southwards from the Axial Zone to the Barbastro anticline. Hale (1985) interpreted these reflectors as salt horizons. The lower reflector was considered to be Triassic salt overlying undeformed foreland. The upper reflector was interpreted in two parts: below the External Sierras it was interpreted as an Upper Eocene salt horizon; below the Tremp-Graus basin it was intepreted either as Triassic salt overlying a thin basement slice, or as a layer of Triassic salt at the base of a thick, cover thrust-sheet overlying a long footwall flat in Eocene rocks.

The second alternative is preferred because it is compatible with the Esera valley section {2.6} and the family of structures for the area {2.3.2} which does not suggest thin basement slices within the foreland basin.

This geometry is also preferred by Camara and Klimowitz (1985) (Fig.2.19) whose section is also based on seismic data. Their section differs in showing a single, continuous, low-angle Tertiary footwall ramp. In this repect the geometry shown by Hale (op.cit.) is preferred, as it is more detailed and in closer accordance with the dataset, in particular with the Tolva borehole which suggests a steep ramp below the Sierra del Montsech.

(b) Boreholes

Boreholes are particularly important in constraining thicknesses and timing deformation within the Cretaceous in the Central part of the section.

In the latter respect the Cajigar borehole (Fig.1.20) is the most important. The Cajigar borehole records a repetition of the Jurassic and Liassic, and, by comparison with the sequence preserved at the surface north of Aren, provides evidence of a marked reduction in the thickness of the Cretaceous. The repetition of the Jurassic gives a lower bracket on the age of deformation. The Aren Sandstone Fmn. gives an upper age: it is approximately constant in thickness across the Tremp-Graus basin and undeformed by the repetition seen in the Cajigar borehole.

Thus, the repetition in the Cajigar borehole is considered to be Cretaceous in age, and to have caused the marked thinning in the Cretaceous sequence. From the data on the section line this date cannot be further refined, but along strike information suggests that the repetition occurred at the end of the Upper Cretaceous during cycle Cb5 {2.8.2}.

(c) Published sections in the Nogueras Zone

Williams (1985) published two short sections through the

Nogueras Zone {2.3.2(d)} to the west of the Ribagorzana valley (Fig.2.1) and one long balanced section along the Ribagorzana valley extending from the Axial Zone to the Ebro basin (Fig.2.22). On each section he interpreted the Nogueras Zone as an antiformal stack. However, his balanced section shows that the culmination was roofed by a major backthrust suggesting that it developed as a triangle zone in a fashion similar to its suggested origin along the Esera valley {2.6.3(b)}.

Williams' balanced section shows the correct geometry for the wrong reasons. He had to include the backthrust because the shortening he implied in the development of the antiformal stack was much larger than the shortening recorded further to the south, both in an absolute sense and in a relative sense.

The shortening in the antiformal stack was large in an absolute sense for two reasons. Firstly, because Williams (op.cit) chose to produce the southwards dip of the trailing edge of the Cotiella thrust sheet by a duplex of thin basement slices - there is no reason why the slices should be thin and, furthermore, the Cajigar borehole shows that this duplex does not exist, at least not at the level suggested by Williams (op.cit.). Secondly, because the geometry chosen for the Permo-Triassic in the Nogueras Zone does not give a minimum shortening estimate.

The shortening in the antiformal stack is large in a relative sense because the section estimates only a very small amount of shortening further to the south: Williams (op.cit.) takes no account of the seismic data suggesting extensive footwall flats beneath the Tremp-Graus basin and the External Sierras {2.7.2(a)}. In addition, the section does not explain the Barbastro anticline, and shows a marked down-to-the-south monocline immediately south of the pinline. In summary, the southern part of the section is almost entirely speculative and constrained only by map data.

A triangle zone roofed by the trailing edge of the Cotiella thrust sheet is the preferred geometry for two reasons, neither suggested by Williams (op.cit.): firstly, it is compatible with the surface geology; and secondly, there is no thrust front within the foreland basin that is time equivalent with the development of the culmination {2.8.3}.

Because the Permian and Triassic sequences are considered to be basement {2.5.2(b)} and because thrust geometries are unknown at depth, in contrast to Williams (op.cit.) no attempt is made to restore the triangle zone.

(d) The Ager basin

The Ager basin is located in the footwall of the Montsech thrust. It has a fill of Tertiary age which records two phases of alpine compression: the first phase during the Paleocene created the basin, the second phase of Upper Eocene/Oligocene age terminated sedimentation in the basin.

The Paleocene history of the Ager basin has been documented by Mutti *et al.* (1985b). They show a basin floored by the Tremp Formation, with a fill folded into an asymmetric south-vergent syncline (Fig.2.24). Thinning of the Alveolina Limestone Formation, and thinning and onlapping of the Figols sequence towards the northern limb of the syncline, suggests that the syncline developed during sedimentation. Clasts of Cretaceous limestone in beach-related facies within the Alveolina Limestone Formation at the northern margin of the basin, are interpreted to indicate the uplift and unroofing of Cretaceous rocks in the hangingwall of the Montsech thrust. Mutti *et al.* (op.cit) go on to imply that movement on the Montsech thrust created and folded the basin during the Paleocene, a view with which the author is in broad accordance. However, several features indicate that the Paleocene basin margin was fixed in position, and is currently exposed close to the Montsech thrust, so that thrusting at this stage must have been relatively minor: major shortening would have overthrust the basin and terminated sedimentation within it. The key features are the persistent thinning and onlapping of formations towards the northern limb of the syncline and the presence of basin margin facies within the Alveolina Limestone Formation on the the northern limb of the syncline.

The preferred interpretation is the inversion of a down-to-the-north extensional fault. The structural style is seen elsewhere in the area {2.3.2(c); 2.6.2} and is closely compatible with the fixed position of the basin margin and the relatively minor displacement. Furthermore, there is a marked decrease in the thickness of the Upper Cretaceous from the hangingwall to the footwall of the Montsech thrust.

In addition to the onlap and thinning relations at the northern margin of the basin, the Figols depositional sequence also onlaps the Alveolina Limestone Formation at the southern margin of the basin. The anticline shows that Alpine deformation extended south of the Montsech during the Paleogene, although the amount of deformation is unknown.

The timing of the second phase of deformation is derived solely from published maps, and based on relationships in the Ager basin immediately west of the Ribagorzana valley. The key points are : firstly, that Middle Eocene (IGME sheet 23 1970; Shell 1975) and Upper Eocene (SNPA 1972) sediments are overthrust by the Montsech, and, secondly, that the Collegats Group of Oligo-Miocene age unconformably overlies the Montsech thrust. The author knows of no studies which describe these relationsips in detail.

(e) Variations within the Upper Cretaceous

Two important aspects of the Upper Cretaceous have been discussed above: the Cretaceous deformation recorded in the Cajigar borehole and the increase in thickness within the Upper Cretaceous from the footwall to the hangingwall of the Montsech thrust, interpreted as the product of an Upper Cretaceous down to the north extensional fault.

In addition to these two points, marked changes in thickness and facies, from shallow marine platform limestones and shales at the Sierra del Montsech to a thicker sequence of deep water turbidites, breccias and olistostromes north of Aren, also help constrain the balanced section (Fig.2.25; van Hoorn 1970; Simo *et al.* 1985). The changes are closely comparable to those seen along the Esera valley from the hangingwall to the footwall of the Cotiella thrust {2.6.2} and directly comparable to the changes seen along the Rio Pallasera to the east (Simo *et al.* 1985; Simo 1986; {2.8.2}). In both examples the thickness and facies changes are considered to reflect Upper Cretaceous extensional faults which defined the southern margin of the Vallcarga flysch basin, the variations along the Ribagorzana valley are interpreted similarly.

The extensional faults along both the Rio Pallasera and the Rio Esera were inverted at a later stage, during the uppermost Cretaceous (Cb5) and the Paleocene respectively. The observed stratigraphic repetition in the Cajigar borehole is interpreted similarly and considered to be the product of inversion along the Rio Ribagorzana, so that it marks the margin of the Vallcarga basin.

2.7.3 <u>The sequentially restored section</u>

(a) <u>The restored section</u>

The restored section is a complete restoration of Tertiary shortening along the Ribagorzana valley (Fig.2.20(a)). However, Tertiary thrust-induced subsidence {1.3.3} is not restored so that the Garumnian dips steadily to the north. Because the Upper Cretaceous deformation in the Cajigar borehole is considered to have been produced during strike-slip when movement was out of the plane of section {2.8.3(a)(iv)} no attempt is made to restore beyond the Tertiary.

By comparison with the deformed state section 64km of Tertiary shortening is implied. The estimate is a minimum because shortening involved in the development of the triangle zone (which is too poorly constrained to be balanced realistically) is not included. The Tertiary shortening can be broken down into four phases: inversion at the Montsech; emplacement of the Cotiella thrust sheet, 21km of shortening; Upper Eocene thrusting at the Montsech, 3km of shortening; second phase of movement on the major thrust, 40 km of shortening.

2.7.3(b) <u>Post inversion at the Montsech</u>

Inversion of an Upper Cretaceous extensional fault during deposition of the Alveolina Limestone Formation and the Figols depositional sequence divided the foreland basin into two discrete basins: the Tremp-Graus and the Ager basins (Fig.2.20(b)). In addition, the Triassic in the hangingwall of the extensional fault was returned to the Triassic regional (albeit a tilted regional). In this way, decollement along the Triassic was facilitated without involving the basement

(Fig.2.20(c)).

The section is drawn with the top of the Figols depositional sequence at sea level. It implies that sedimentation also occurred to the south of the Ager basin, a feature required by section (c) (Fig.2.20(c)) which shows the Cotiella thrust sheet emplaced on a thick sequence of Tertiary sediments. The position of the top of the Montanana depositional sequence is constrained by two features: (i) the preserved thickness of the Montanana depositional sequence in the Tremp-Graus basin; and (ii) the seismically defined thickness of the Tertiary footwall ramp below the Tremp-Graus basin. The second constraint is derived from the need for compatible amounts of shortening along the Esera and Ribagorzana valleys during the Figols and Montanana depositional sequences. Since compatibility can only be achieved by considering that the displacement seen on the Cotiella thrust along the Esera valley is represented along the Ribagorzana valley by the repetition on the lower of the two Tertiary footwall ramps, the thickness of this ramp must be equal to that of the Figols and Montanana sequences combined.

Uplift must also have occurred at the northern end of the section line at this time, since sediments within the Figols depositional sequence record a northern basin margin and sediment source below the Sierra del Sis (Puigdefabregas *et al.* 1985). The uplift is shown schematically, inferring a northern pinchout of the Figols Formation just north of the preserved basin margin.

2.7.3(c) Emplacement of the Cotiella thrust sheet

The requirements of compatibility with the Esera valley section {2.6} suggest that major displacement occurred along the Ribagorzana valley during the Figols and Montanana sequences {2.6.3(a); 2.7.3(b); 2.8.3} corresponding to the emplacement of the Cotiella thrust sheet.

Section (c) (Fig.2.20(c)) shows the leading edge of the thrust sheet displaced 21km southwards at the land surface, to a position equivalent to the base of the second ramp in the Tertiary basin fill (the sediments which form the second ramp had yet to be deposited). The top of the Montanana sequence must be horizontal at the end of the emplacement of the Cotiella thrust sheet when allowance is made for regional subsidence. The basin fill is only shown schematically, as the thrust sheet is believed to have moved <u>during</u> the Figols/Montanana depositional sequence, and not immediately after the deposition of the Montanana depositional sequence.

Two features are implied by section (c): (1) that relief created by inversion at the Montsech thrust front was overtopped during deposition of the Montanana sequence so that the Ager and Tremp-Graus basins were no longer distinct; and (2) that a southern source area developed during emplacement of the thrust sheet but that it was the product of a ramp in Tertiary sediments not the Montsech thrust.

Mutti *et al.* (1985b) record Montanana deposits in the Ager basin but do not document their nature. So the first point cannot be tested. The southern source area/high is in accordance with the northwest directed palaeocurrents recorded in fluvial sediments at the southern margin of the Tremp-Graus basin by Nijman and Nio (1975; Fig.2.11) and Atkinson (1983; Fig.2.13; {2.7.2(d)(ii)}).

(d) <u>Development of the triangle zone</u>

The basement and cover culmination at the northern margin of the basin is considered to be a triangle zone for two reasons: firstly the surface geology has the correct geometry, an antiformal stack of thrust sheets roofed by backthrusts and, secondly, it is the only geometry compatible with uplift at the northern margin of the basin to produce the post-Castisent increase in sediment input {2.5.3} without a time equivalent thrust front in the foreland basin.

The development of the triangle zone immediately after emplacement of the Cotiella thrust sheet is directly comparable to the history seen along the Esera valley.

(e) <u>Renewed movement of the thrust sheet</u>

The deformed state section shows that a further phase of thrust sheet movement occurred, during which the leading edge of the thrust sheet climbs the second Tertiary ramp to decolle along an Upper Eocene evaporite horizon. The resulting variable load on the salt horizon is considered to mobilize the salt and cause it to flow to lower lithostatic pressure at the margins of the thrust sheet, where it forms a salt pillow, the Barbastro anticline {3.8.1(b)}.

The fill of the Ager basin dates the second phase of movement on the Montsech as syn- or post-Upper Eocene {2.7.2(d)}. However, map relations suggest that folds in the External Sierras above the Upper Eocene decollement, are largely syn- or post-Oligocene in age (IGME sheet 33 1970) while deformation along strike is clearly Oligo-Miocene in age {4.5}. Because of the dominance of Oligocene and Miocene structures the renewed movement of the major thrust sheet is considered to be Oligocene or Miocene in age, and the Upper Eocene movement on the Montsech thrust is considered to be local in nature {2.8.3(b)(iii)}.

2.8. BALANCED, SEQUENTIALLY RESTORED MAPS OF THE SOUTH CENTRAL PYRENEES

2.8.1. Introduction

A three-dimensionally balanced cross-section would offer the most complete illustration of the structural evolution of an area, defining thrust sequences and thrust sheet geometries. But they are a new innovation: no such sections have been published. A methodology for their construction involving interpolation between two-dimensional sections is developed below.

The two two-dimensional balanced sections presented above {2.6; 2.7} are shown to be compatible and, therefore, to have the potential to produce a three-dimensional section. However, the lack of data which could be used to enhance these sections makes an attempt at the production of a three-dimensional section fruitless. Instead a sequence of balanced maps were produced. The maps have the same advantages as three-dimensional sections in restoring basin geometries and showing points of uplift.

The maps describe the Upper Cretaceous to Miocene evolution of the South Central Pyrenees.

2.8.2 Three dimensional section balancing, and balanced maps

Accurately balancing volumes would be an extremely complex and time consuming task - requiring the mathematical integration of the area of an infinite number of infinitesimally thin cross-sections throughout the study area. But the method can be approximated by linking closely spaced two-dimensional sections. In this way only the two-dimensional sections are truely "balanced": the intervening volumes are compatible with the sections which bound them and the data recorded within them.

The absence of three-dimensional sections in the literature reflects two exacting requirements: firstly, that deformation approximates to plane strain over a large volume of crust, and secondly, that the dataset is complete.

Deformation can be considered to be plane strain when thrusting directions are parallel and constant over a large area. This requirement is satisfied in many mountain belts at the intermediate scale (10's km along strike) and at the large scale where mountain belts are straight (Fig.2.26(a)). But, many mountain belts are arcuate with divergent thrusting directions that imply strike parallel extension and non-plane strain, for example the Alpine arc swinging from S.W. directed thrusting in southern France (Graham 1985) to N.W. directed thrusting in northern France and Switzerland (Butler *et al.* 1986; Fig.2.26(b)). In such cases the technique is inappropriate.

A complete data set is required to constrain the geometries of the two-dimensional sections, and the geometries of thrust sheets between sections. With a poor data set the number of possible geometries rapidly multiplies and it is increasingly difficult to reconcile adjacent two-dimensional sections.

The two sections presented above are parallel to each other and to the thrusting direction {1.2.3}. So the volume of rock between them can be considered to have undergone strain only in the plane of the section line. In this way the first requirement is satisfied.

However, two points show that the data set is incomplete: firstly, the sections do not imply the same amount of shortening in the foreland basin (the Esera section implies 51km of shortening while the Ribagorzana section implies 64km) and, secondly, the thickness of Tertiary sequences preserved below the External Sierras is incompatible between sections.

Both points demonstrate that either or both balanced sections must be changed to make them compatible. But the changes required are small, and within the bounds of interpretational licence. So two compatible balanced sections can be constructed which would allow a three-dimensional balanced section to be drawn.

Such a section has extra value over a map which links balanced sections only when extra information is included. Because the section lines discussed here are close and because data has been projected on to both section lines, there is little information to add. Consequently no attempt was made to construct a balanced volume. Instead a series of "balanced maps" was produced linking sequentially restored sections. The maps have the merits of a three-dimensionally balanced section in showing the traces of thrust ramps which are the key features in controlling the basin geometry and sediment dispersal patterns.

2.8.3. Balanced sequentially restored maps of the eastern part of the South Central Pyrenees

The evolution of the eastern South Central Pyrenees can be divided into two parts by the base Danian Transgression {2.5.1}. Prior to this transgression the Upper Cretaceous was characterised by a north-facing slope, followed by marked extensional faulting and latest Cretaceous transpression producing local inversion. Alpine compression developed after the Danian transgression and was characterised by the emplacement of large thrust sheets within the foreland basin. Between periods of thrust sheet emplacement the active thrust front switched from the foreland basin to become emergent as a roof thrust to a triangle zone at the northern margin of the basin.

The two phases, pre- and post-Danian transgression, are considered separately below. For each the relevant data from the section lines and key localities not on the section lines is summarised, and the evolution is then inferred. Particular attention is paid to how thrust sheet geometries relate to the fill of the foreland basin.

The extra shortening implied by the Ribagorzana section requires two changes to the Esera section, both involving the addition of shortening to the Esera section, 1km during the emplacement of the Cotiella thrust sheet and 9km during the emplacement of the the second major thrust sheet. The changes are explained during synthesis of the section lines. They are not shown on the balanced section of the Esera valley, but they are taken into account on the "balanced maps".

2.8.3(a) The Upper Cretaceous - pre-Danian transgression

For ease of correlation the depositional cycles of Souquet (1984) Cb_{1-5} are used throughout (Fig.2.6).

2.8.3(a)(i) The Esera valley

Upper Cretaceous deposition commenced at the top of Cb_1 and continued into Cb_2 with deposition of massive marine limestones which have not been studied in detail by the author.

Cycle Cb_3 is considered to have been associated with two extensional, down-to-the-north, normal faults to the N.W. of the present Turbon structure, one orientated W.N.W.-E.S.E. and another N.E.-S.W. Shallow water facies developed in the footwall, while deeper water facies and thicker sequences developed in the hangingwall block.

Cycle Cb_4 was associated with increased activity along the faults seen in cycle Cb_3, especially the W.N.W.-E.S.E. fault. This had two effects: firstly, the formation of a W.N.W.-E.S.E. fault scarp, and secondly, the drowning of the Turbon high to the east of the section line, which had previously been a carbonate platform. The fault scarp defined the southern margin of the Vallcarga flysch basin and produced the Campo breccia Formation as a base of slope deposit.

At the Turbon cycles Cb_{1-4} are deformed by N.-S. trending folds and reverse thrusts which are onlapped by the Salas Marls Formation of cycle Cb_5. The deformation is interpreted as base Cb_5 in age, and to have been produced by E.-W. compression. There is no evidence for deformation of this age along the Esera valley.

2.8.3(a)(ii) Ribagorzana

The Upper Cretaceous along the Ribagorzana valley has a feather edge below the Tremp-Graus basin. From there it thickens steadily northwards to an Upper Cretaceous, down-to-the-north, extensional fault, which became inverted during compression to form the footwall ramp to the Montsech thrust. Northwards of that fault the Upper Cretaceous changes in facies and increases in thickness from platform limestones at the Sierra del Montsech to deep water sediments north of Aren. The changes are considered to be the product of an Upper Cretaceous extensional fault inverted during the uppermost Cretaceous to produce the repetition of the Jurassic and thinning of the Cretaceous at the Cajigar borehole.

2.8.3(a)(iii) Boixels thrust and San Corneli anticline

The Upper Cretaceous from the Ribagorzana valley to the Segre fault has been reviewed by Simo (1986). He recognises the same five cycles as Souquet (1984) and documents the development of carbonate platform, slope and deep basin facies. The author follows Simo's interpretation that the changes which produce the cycles are controlled by:

> (1) tectonism in terms of abrupt subsidence events, flexure, and extensional and compressional movements; (2) relative abrupt sea-level rises in accordance with subsidence events; and (3) inherited depositional profile. Growth potential of platform and rim, siliclastic progradation and sea-level fluctuations of each sequence were less important factors.
> - Simo (1986)

During cycle Cb_1 a north-facing slope developed (Simo op.cit.). The slope was composed of a carbonate platform with its edge defined by a Lower Cenomanian listric normal fault, passing northwards into slope and basin facies (Fig.2.23).

Cycles Cb_2 and Cb_3 both commence with relative sea level rise, so that the carbonate platform is initially drowned, only to be re-established in time. Throughout these two cycles the north facing slope inherited from Cb_1 times is maintained, but the platform edge is defined sedimentologically.

Cycle Cb_4 commences with a third relative sea level rise and the development of an extensional, down-to-the-north, normal fault. The fault divided two distinct depositional areas:

> (a) on the south side (footwall block) a skeletal homoclinal ramp prograding northward; and (b) on the North (hangingwall block) a turbiditic basin [The Vallcarga Basin] developed over irregular faulted and tilted blocks. - Simo (1986)

Cycle Cb_5 commences with the development of major contractional structures, e.g. the San Corneli anticline, whose axis rotated from N.E.-S.W. to E.-W. orientated during the cycle (Simo *et al.* 1985) and the Boixels thrust considered to be an inverted extensional fault which previously bounded the margin of the Vallcarga basin. The first effect of these structures was to trigger a major olistostrome, the Pumanyons olistostrome, but they were active throughout Cb_5, controlling sediment thickness and facies patterns in the Aren Sandstone and Tremp Formations (Nagtegaal *et al.* 1983; Sgavetti *et al.* 1984; Simo *et al.* 1985

2.8.3(a)(iv) Summary of the Upper Cretaceous

The different localities have closely comparable Upper Cretaceous histories in terms of depositional facies and tectonic phases. Consequently a generalised evolution can be established.

During depositional sequence Cb1 a north-facing slope developed with a slope break defined by listric normal faults of Lower Cenomanian age. Three distinct facies belts were established that were to persist until Cb_4: carbonate platform; slope and deep basin.

Cycles Cb_2 and Cb_3 each commence with a sharp sea-level rise. Sea level rise drowned the southern carbonate platform, but subsequent still stands allowed re-establishment and progradation of the platform environment. In addition, local faulting, along E.S.E.-W.N.W. and N.N.E.-S.S.W. trends, controlled facies and sediment thicknesses during Cb_3 to the N.W. of the Turbon.

Cycle Cb_4 commenced with two significant events: a phase of marked extension dominated by down to the north W.N.W.-E.S.E. trending faults and a marked sea-level rise. Sea level rise drowned the carbonate platform in the south while faulting produced a scarp which separated the platform from a deep basin, the Vallcarga basin, in the north. Thickness variations, onlaps and palaeocurrents in the Mascarell Formation (Fig.2.8) suggest a number of fault bounded N.N.E.-S.S.W. trending highs within the Vallcarga basin notably: (i) a high at the present position of the Turbon anticline; (ii) a down to the N.W. fault at the site of the present Sierra del Sis, to the east of the Rio Isabena, and (iii) a down to the N.W. fault north of San Corneli. The thickness and facies changes which suggest these faults are well illustrated by Van Hoorn (1970) and SNPA (1972) and were first interpreted in this way by Van Hoorn (op.cit.). Simo (1986) produced the first map linking faults inferred from W.N.W.-E.S.E. trends to those on N.N.E.-S.S.W. trends (Fig.2.27). He depicts faults which are scallop-shaped in plan, with the N.N.E.-S.S.W.

faults being transfer faults that define the extension direction and the W.N.W.-E.S.E. faults being oblique dip-slip faults. His model is closely followed here, with added emphasis on the N.N.E.-S.S.W. trending faults as having controlled sediment thickness and, in the case of the Turbon, facies changes. The balanced map shows the situation at the end of this extensional phase, with the sequences restored to their pre-contraction positions (Fig.2.28(a)).

Cycle Cb_5 is characterised by compressional structures: in particular the Turbon anticline, stratigraphic repetitions in the Cajigar borehole, the San Corneli anticline and the associated Boixels thrust. But the interpretation of these structures is problematical. Previous authors (Camara and Klimowitz 1985; Simo 1986) have linked the Boixels thrust to the stratigraphic repetitions in the Cajigar borehole, implying a single laterally extensive thrust surface produced by N.E-S.W. orientated compression. The thrust surface is considered to be a reactivated extensional fault which formed during Cb_4. Although this interpretation - a direct change from N.N.E.-S.S.W. extension (Cb_4) to N.N.E.-S.S.W. (Cb_5) compression signalling the onset of Alpine compression - is appealing, it is difficult to reconcile with the structures seen elsewhere. At the Turbon N.N.E.-S.S.W trending folds and reverse thrusts imply W.N.W.-E.S.E. compression, while there is no evidence for deformation of this age along the Rio Esera. An alternative interpretation is required.

The variable compression directions could be the product of dextral strike-slip movement, with the hangingwall blocks formed during the extensional phase moving E.S.E. with respect to their footwalls. In such a situation, uplift, strike-slip or extension would result depending on the orientation of the pre-existing fault structures with respect to the direction of strike-slip. The model can explain all the observed structures since the sites of compression during the uppermost Cretaceous are coincident with extensional faults orientated oblique to the inferred W.N.W.-E.S.E direction of strike-slip. The Boixels thrust and repetition in the Cajigar borehole are interpreted as flower structures (Harding and Lowell 1979); the folds and reverse faults seen at the Turbon are considered to be the product of inversion and cover shortening over a basement high orientated perpendicular to the direction of strike-slip. By contrast, the absence of compressional structures of this age along the Esera valley is thought to reflect a basin bounding fault parallel to the strike-slip direction.

Although the explanation appears to fit the data set it can only be considered speculative at present. Particularly because the dextral sense of strike-slip required for the model is opposite to the regional sinistral strike-slip between Spain and France during the Upper Cretaceous {1.2.2}. However, this problem is tempereu by the fact that the relatively well constrained N.N.W.-S.S.E extensional phase also has to be accommodated in the regional setting of W.N.W.-E.S.E sinistral strike-slip.

2.8.3(b) The Tertiary structural evolution of the South Central Pyrenees

For ease of correlation the depositional sequence terminology of Mutti *et al.* (1985a) is used in this section. These and other stratigraphic terms are described in {2.5.3} and summarised on Figure 2.15.

2.8.3(b)(i) The Esera Valley

The first post-Danian event along the Esera valley is the emplacement of the Cotiella thrust sheet. The thrust sheet is known from an exposed overthrust below the mountain of Cotiella where a minimum dispacement of 20km can be recorded (Fig.1.20). The thrust sheet continues into the foreland basin where it is buried by Cuisian (Lower Eocene) sediments, at La Atiart G.R.[7401] so that it cannot be traced southwards from this point (SNPA 1972; Nijman and Nio 1975). However, because there are no intervening accommodating structures and because displacement gradients on thrusts are low, the thrust must be present along the Esera valley section line within the foreland basin. Balancing constraints suggest that the leading edge of the thrust sheet has been considerably eroded but that it extends well into the Tremp-Graus basin {2.6.3}. The emplacement of the thrust sheet is bracketed to have occurred after deposition of the

Alveolina Limestone Formation and before the Castisent Formation i.e. during the Figols and Montanana depositional sequences.

Thrust sheet emplacement was followed by the development of a triangle zone at the northern margin of the foreland basin. The roof thrust to this culmination became the active thrust front i.e. the thrust front switched from within the foreland basin to its northern margin. The uplift of the culmination resulted in an unconformity at the basin margin overlain by a major fanglomerate body, the Campanue Conglomerate within the Santa Liestra depositional sequence.

Two features along the Esera indicate that the thrust front may have switched from the triangle zone to the foreland basin during the Upper Eocene: firstly, the presence of W.N.W.-E.S.E. trending folds overlain by stacked conglomerate fan sequences of Priabonian age to the east of Naval; and secondly, a phase of uplift at the Mediano during the Biaritzian. Alternatively both deformation events could be interpreted as the product of halokinesis.

The final stages of Alpine deformation are recorded in the Collegats Group, a fluvial basin-fill sequence (over 500m thick in places) of uppermost Oligocene to Miocene age {4}. Provenance arguments, onlapped folds and the scale of the sequence suggest that Collegats Group sedimentation was initiated by the continued development of the triangle zone. However, marked changes in palaeogeography, the generation of local sediment sources, and evidence for syntectonic sed~~i~~imentation, suggest that the active thrust front switched to become emergent in the foreland basin during deposition of the lower part of the Collegats Group.

2.8.3(b)(ii) <u>The Ribagorzana Valley</u>

The first Tertiary deformation along the Ribagorzana section is recorded in the Ager basin, in the footwall of the Montsech thrust. The generation of a local sediment source in the area of the present Montsech, onlapped folds and stratal wedging, suggest N.N.E.-S.S.W. compression during the deposition of the Alveolina Limestone Formation and Figols sequence. However, any movement on the Montsech thrust is thought to have been minor at this stage: large movement would have overthrust and terminated sedimentation in the Ager basin. Given the major thickness contrast between hangingwall and footwall in the Montsech thrust, a model of inversion of an Upper Cretaceous down to the north extensional fault is preferred.

A marked increase in sediment input at the northern margin of the basin at the end of the Castisent depositional sequence is considered to record the development of a triangle zone in that area.

The main phase of thrusting at the Montsech is bracketed as syn- or post-Escanilla Formation (Priabonian) and prior to the deposition of the Collegats Group (uppermost Oligocene to Miocene; SNPA 1972).

2.8.3(b)(iii) Synthesis

The first Tertiary event along the Esera valley is the Figols-Montanana emplacement of the Cotiella thrust. The geometry of the thrust sheet is well known in the west, but difficult to trace eastwards towards the Ribagorzana valley. The fill of the Ager basin records approximately contemporaneous movement along the Montsech thrust - Alveolina Limestone Formation-Figols in age - but the shortening on the Montsech thrust is considered to have been minor at this time - insufficient to explain all the contraction implied by the Cotiella thrust.

The shortening is considered to have been accommodated in the footwall of the Montsech thrust for two reasons: firstly, there is no evidence for deformation of this age to the north of the Montsech thrust, and, secondly, the Ribagorzana section implies a large amount of shortening in the footwall of the Montsech thrust which can only be compatible with the Esera valley section if some of that shortening occurred at this stage.

So the leading edge of the Cotiella thrust is considered to extend to the south of the Montsech thrust into the External Sierras: the Montsech thrust was only a minor splay in the hangingwall of the Cotiella thrust sheet at this stage. There is some evidence for deformation of this age in the External Sierras; the southern margin of the Ager basin, in the footwall of the Montsech is defined not by a feather edge but by an anticline, which formed contemporaneously with the Ager basin.

The exact geometry of the thrust sheet is only vaguely defined. Paleocene and Eocene sequences preserved within the External Sierras (SNPA 1972) offer the opportunity to test and refine the model, but they have not been studied.

A consideration of the thickness of the Tertiary sediments that underlie the lower of the two Tertiary footwall flats suggests that this flat represents the contemporary land surface, and that the leading edge of the Cotiella thrust sheet was emplaced to a position equivalent to the base of the second Tertiary footwall ramp {2.7.3(b)(c)}. A 21 km displacement of the leading edge of the Cotiella thrust sheet is implied, 1km greater than the minimum displacement recorded along the Esera valley. For compatibility, the displacement on the Esera section must be increased by 1km. This can be achieved by lengthening the Esera section immediately to the south of the postulated Upper Cretaceous extensional fault (position A on Figure 2.17(a)). With this geometry the changes which must be made to the deformed state section along the Esera valley are minimal: the extra length in the footwall would become buried below the backthrust, and the extra length in the hangingwall would simply increase the postulated amount of erosion of the leading edge of the Cotiella thrust (the slight change in cross-sectional area of the Salas Marl Formation in the hangingwall of the Cotiella thrust could easily be accommodated on the deformed state section by subtle changes of fault attitudes).

The reconstruction at the end of the emplacement of the Cotiella thrust sheet (Fig.2.28(b)) corresponds to the basin geometry during the Castisent depositional sequence. The basin geometry is controlled by inverted Upper Cretaceous extensional faults, and by ramps in the footwall and hangingwall of the Cotiella thrust. The folds and thrusts produced by inversion control the northern margin of the basin and are shown schematically along the Esera and Pallasera valleys. A hangingwall fold above a footwall ramp corresponding to the Segre fault forms the eastern margin of the basin, while a similar structure defines the southern margin of the basin on frontal/oblique ramps between the Pallasera and Esera valleys. The inferred southern and eastern margins constitute an explanation for N.W. directed palaeocurrents seen within the Castisent depositional sequence - the "Segre hangingwall fold" would have guided distributary systems southeastwards to be reflected northwestwards at the E.-W. trending hangingwall fold. Since the basin configuration shown in the diagram was established during the emplacement of the Cotiella thrust sheet, and persisted until the Upper Eocene, it also offers an explanation for similar north-directed palaeocurrents in the S.E part of the Tremp-Graus basin in the Montllobat and Capella sequences.

The amount of erosion required from the leading edge of the Cotiella thrust sheet is problematical, particularly along the Esera valley. However, two points can be brought to the support of the geometry shown: firstly, the geometry of the Cotiella thrust sheet is well defined along the Esera valley, and secondly, it is at its thinnest where the most erosion is considered to have occurred.

The development of a triangle zone at the northern margin of the foreland basin following the emplacement of the Cotiella thrust sheet has been inferred along both the Esera and Ribagorzana valleys. An equivalent culmination occurs all along the northern margin of the foreland basin, but it has not previously been interpreted as a triangle zone. Williams (1985) considered it to be an antiformal stack along the Ribagorzana valley (Fig.2.22). Camara and Klimowitz (1985) consider the culmination to be a hinterland dipping duplex in the west and an antiformal stack in the east (Fig.2.19).

There are two key points in generalising this structure. Firstly, that with an antiformal stack, or a hinterland dipping duplex, shortening within the culmination is transferred southwards so that a time equivalent thrust front should be present within the foreland basin, and, secondly, although the actual geometry of the culmination at depth is only poorly constrained, and cannot be balanced with much certainty, large amounts of shortening are implied. Thus, although the development of a triangle zone during the Santa Liestra sequence (Ypresian-Lutetian) is well documented only along the Esera valley, the absence of a Middle Eocene thrust front within the foreland basin, suggests that the structure occurred throughout the study area.

The Upper Eocene saw a number of important deformation events within the foreland basin: th main folding of the N.-S. trending Boltana Anticline occurred · mediately after the deposition of the Banaston sequence (Mutti t *al.* 1J85a); the main phase of d'apiric uplift at the Med a o oc u red duting the Biarritzian and lower Priabonian; p r Co ene fold'ng and alluvial fan sedimentation is seen to the east of Naval; the first deformation n the External Sierras s uth f the Jaca bas'n occurred during the Priabonian (Arguis arls F mation); and the main phase of thrusting at the Montse h is bracketed as syn- or post-Escanilla (Priabonian) to pre-uppermost Oligocene to Miocene.

It might be felt that there is sufficient evidence to infer that the thrust front was active w'thin the foreland basin during the Upper Eocene However, the outcrops are scattered and it is xtremely difficult to meaningfully link the structures to produce well defined thrust s eets.

Following Deramond *et l.* (1984) the Boltana anticline and the first phase of def rma ' n in the External Sierras south of the Jaca basin are interpreted as the product of westwards collapse of a series of oblique lateral ramps. But it is difficult to link this deformation eastwards as a coherent thrust front. One point is clear though, activity on the Montsech thrust demonstrates that the progressive westwards migration of the thrust front suggested by Camara and Klimowitz (1985) did not occur.

The author considers that the Upper Eocene movement on the Montsech thrust is only local and that it was not present along the Esera valley. The Mediano diapir and the deformation at Naval are thought to be the product of halokinesis, which may have been triggered by movement of a larger thrust sheet whose front is not clearly defined eastwards of the Boltana anticline. If this is correct then the balanced section implies that the shortening was accommodated in the footwall of a second major thrust sheet, termed thrust sheet 2. Upper Eocene shortening in the forland basin would help to explain the regional unconformity at the base of the Collegats Group. However, the idea is speculative, and the amount of shortening is unknown so that no balanced map has been produced. The proposed geometry is problematical on one account: it is unclear how the displa ement on the Montsech thrust dies out or is accommmodated to the st. A N N.E.-S.S.W. trending fault zone near the Village of Luz s, b ack t d in age as syn/post-Escanilla Formation and pre-C ll g s Gr p, could have helped to accommodate the d's l nt. Ind ed Cuevas *et al.* 1985 consider these faults to l a e b en dominated by strike-slip. But this observation shou d be treat d w h caution: firstly, Cuevas *et al.* (op.cit) cons'd r t e mai strike slip phase to have occurred during the y Eocene, and, secondly, they do not report the sense of m ve nt on the faults.

Syntectonic sedime tati n, locally generated sediment sources and marked changes in pala ogeography document that the thrust front was located in the External Sie ras at the northern margin of the Ebro basin d r' g the uppermost Oligocene and Miocene {4; Nichols 1987}. Howev r, t p ints suggest that immediately prior to this the thrust ro was lo ated at the northern margin of the foreland basin, nd hat a second switch in the position of the deformation front ccu red from the foreland basin to the Axial and Nogueras Zon s: firstly, E. Jolley pers. comm. (1987) documents the southward m'gration of a thrust front from the Axial Zone to the External Sierras during the Oligocene (Stampian and Chattian) folding the foreland basin fill and controlling sediment thickness and facies patterns in contemporaneous sediments now preserved in E.-W. trending synclines north of the External Sierras and, secondly, provenance arguments and onlapped folds which suggest thrusting in Axial and Nogueras Zones immediately prior to the development of the Collegats Group {4}.

CHAPTER 3

An Awhesyth - the lark (trad. Cornish)

Coleman and Spring (1987)

CHAPTER THREE

DIAPIRISM IN THE SOUTH CENTRAL PYRENEES

3.1 INTRODUCTION

Diapirs are an important part of the geology of the South Central Pyrenees. Locally they dominate the structure, and control sediment thickness and facies patterns. This chapter discusses five distinct diapirs which illustrate a range of structural styles and sedimentary responses to diapirism.

Prior to detailing these examples factors which weigh upon the initiation and recognition of diapirism are discussed. This step is important as although diapirs are a widespread phenomenon with great economic significance (Taylor 1984) they have a low profile in recent academic literature, and their mechanics and geometries are not widely known.

3.2 THE MECHANICS OF DIAPIRISM

3.2.1 Diapiric forces: controls on dipirism

The forces and controls of diapirism are simple. Consider two horizontal layers of fluid of infinite width: the upper layer being of density p_1 and thickness h, and the lower layer of density p_2 where $p_1 > p_2$ (Fig.3.1(a)). The situation is stable, since the flat surface which separates the fluids is a plane of constant pressure. The fluids will not move until there is a perturbation, or tilt on the surface which separates them.

pressure at any point a, on the surface = $h.p_1.g$ -(1)
where g = acceleration due to gravity.

The introduction of a tilt or a perturbation sets up a pressure gradient in the fluid. Consider a point c at the base of a perturbation extending a distance h' above the lower layer (Fig.3.1(b)).

pressure at c = $(h-h').p_1.g + h'.p_2.g$ -(2)

The pressure difference between points a and c =(1)-(2)

$$= h.p_1.g - [(h-h'.p_1.g + h'.p_2.g]$$
$$= h'.g.(p_1-p_2)$$

This equation expresses the pressure differential created by diapirism. Two key factors are apparent.
(1) The top surface of the diapiric material cannot be an equipotential surface: it must be tilted, have a perturbation or be differentially loaded for diapirism to initiate. The pressure gradient increases with the value of this perturbation (h')
(2) For the lower layer to rise diapirically $p_2 < p_1$.
Furthermore, the pressure gradient increases with the value of (p_1-p_2) i.e with the density contrast.
To these a third can be added
(3) For a material to be diapiric it must be able to flow at pressures and temperatures seen within the crust.

3.2.1(a) <u>The development of a non-equipotential surface</u>

Equipotential surfaces are geologically unlikely because of the wide range of ways by which salt can be differentially loaded.

At the basin scale most sedimentary layers are tilted during burial due to variable subsidence and compaction rates. Consequently a salt layer of constant thickness would be under greater load at the basin centre than at the margins. So salt flows to the margins and may be completely withdrawn from the centre of the basin (Fig.3.2). Trusheim (1960) considers that the North Sea Zechstein generally accords with this model. Because the highest pressure gradient is set up between the basin centre and the basin margin the first diapirs are likely to develop at the margin; once they have reached an equilibrium position, younger diapirs may develop closer to the basin core.

Within basins, faulting may produce a non-equipotential surface. Consider an extensional fault which produces a step in a salt layer (Fig.3.3). Subsequent sedimentation on the hangingwall block sets up a pressure differential across the fault, and causes salt to flow from the hangingwall to the footwall. The close parallelism between salt structures and tectonic trends in the Southern North Sea suggests that salt movement there may have been triggered in this way (Taylor 1984; Fig.3.4).

Differential loading within basins may also be generated by facies or depositional sedimentary thickness variations. Jenyon (1985) considers that deltas may be important in this respect (Fig.3.5). In foreland basins thrust sheets could have a similar effect - the author considers that the Barbastro anticline may have been initiated in this way {3.8}.

3.2.1(b) Relative densities

For diapiric movement to initiate a layer must have a lower density than its overburden. This requirement divides diapirs into two types: those that are detached from their "mother" or source beds, and those that maintain a link.

Diapiric bodies not attached to their mother beds rise to a level in the crust where the overburden has the same average density as the diapiric material (Fig.3.6(a)). The average salt dome is 90% halite and has a density of 2.16g/cm^3 (Gussow 1968; Jenyon 1985). Salt domes may rise to the surface in consolidated sequences, but stop short in newly deposited sequences. For example, compaction curves show that shale has the same density as halite at a depth of 3,000 feet (Gussow op.cit.; Fig.3.7).

This contrasts with diapiric bodies attached to deep mother beds. As a result of the differential hydrostatic pressure between the base and the top of the diapir, attached diapirs can rise to the surface (Fig.3.6(b)). The key feature is that the average density of the overburden above the mother bed is greater than that of the salt column. The pressure difference between the top and the base of the diapir increases until it reaches sediment of the same density, and then decreases but remains positive (Fig.3.7).

Density variations are also important within diapirs, producing variable rates of diapiric rise. Halite dominates most salt domes, but impurities with lower density and lower yield stress, such as the bittern salts (e.g. carnallite 1.602 g/cm^3 and sylvite 1.992 g/cm^3) rise faster than halite (2.2 g/cm^3) while high density materials such as anhydrite (2.98 g/cm3) and limestone are negatively bouyant. The structures produced by these impurities depend upon their initial distribution and on the degree of separation that operates during movement (Jenyon 1985).

3.2.1(c) The ability to flow

For a material to behave diapirically it must be able to flow at pressures and temperatures seen within the crust. Coal, for example, has a low density but does not form diapirs because it cannot flow.

The capacity of halite to flow is controlled by two factors: temperature, and the amount of water included within the halite lattice.

Laboratory experiments show that plastic deformation of pure halite will commence at temperatures above 200°C at stresses of 44kg/cm³ (Gussow 1968; Fig.3.8). However, glaciers of salt issue from piercement structures in Iran. Talbot (1979) documents contemporary, wet-season movement rates of up to 50cm per day in these glaciers, and the active development of mylonitic fabrics, suggesting that pla tic flow occurs despite the low temperatures (max'mum 46 C (K nt 1979)) and low surface gradients. The observat'ons are critical. They demonstrate that the activity of re l alt domes is not limited to temperatures greater than 200°C and stresses in excess of 44kg/cm³ as suggested by laboratory work. Jenyon (1985) suggests that the presence of even minute quantities of free water in the halite lattice markedly lowers the elastic limit below the strength, so that flow occurs at low temperature and low stress.

3.3 RECOGNITION OF DIAPIRISM

Features characteristic of diapirism occur at a range of scales. Large scale features re the diapir c bodies themselves and a suite of related struc es in the d pir carapace, and medium and small scale structu es are prod ed by plastic flow within the salt body.

3.3.1 Diapir structures

3.3.1(a) Diapiric bodies

There are two types of diapiric bod s: pillows and piercement structures (Fig.3.9). In p'llows the stratigraphic relation between the salt and overlying sequences is maintained. By contrast piercement structures are intrusive into overlying sequences: they usually evolve from salt pillows.

The plan form of diapirs is variable. Pillows are likely to follow the geometry of whatever mechanism triggered salt movement. For example the elongate salt pillows seen in the southern North Sea reflect fault trends (Fig.3.4). Piercement structures may follow pillow trends. Alternatively, a perturbation in the pillow crest or a weakness in the overburden may favour a localised piercement. Such a piercement is increasingly favoured as it develops in height above the pillow so that an elongate piercement may never develop {3.2.1(b)}. Where the regional stress field is neutral and the overburden isotropic, piercements of this type are circular in shape (Fig.3.9).

3.3.1(b) Salt withdrawal basins

As diapirs grow they draw salt from surrounding areas, causing subsidence in regions overlying areas of withdrawl. Basins formed in this way show a systematic three stage pattern of salt withdrawal (Trusheim (1960); Fig.3.10).

The first stage corresponds to the development of a salt pillow. Salt is withdrawn from a relatively large area producing a wide basin known as the "primary rim syncline". Sediments in the basin thin gradually towards the pillow crest.

The second stage corresponds to the development of a piercement structure. The crest of the pillow is breached and salt is withdrawn from the pillow, producing rapid subsidence close to the growing salt column; subsidence which forms the "secondary rim syncline". The secondary rim syncline is characterised by a sedimentary fill that thickens towards the salt column. The subsidence also causes the inner margin of the primary rim syncline to dip towards the piercement structure. Where two adjacent piercement structures occur the primary rim syncline may be inverted to produce a "turtle-back" structure (Fig.3.10).

As the top of the diapir reaches the surface, subsidence and uplift produced by the domed top of the diapir, may combine to produce a Tertiary rim-syncline.

Salt withdrawal basins control sediment thicknesses; the changes in subsidence rates which they imply can be expected to produce facies variations. In addition, their fills demonstrate that diapirs are long lived features (e.g. Taylor 1984) and not catastrophic as some early authors thought (Gussow 1968).

3.3.1(c) Faults in the diapir carapace

(i) Ballooning stage

As a diapir grows two types of fault develop in its carapace: extensional faults due to diapir inflation and extensional faults due to variable compaction.

Extensional faults are produced during diapir inflation because the carapace is effectively pinned below rim synclines: as the diapir inflates the sequence between pinning points must be extended. The resulting fold is a bending fold, characterised by strike-parallel extension; it contrasts with buckle folds which are characterised by strike-perpendicular extension (Hancock 1985; Fig.3.11). Where the salt structure is circular and the regional stress field is neutral a radial pattern of extension can be expected (Gussow 1968; Fig.3.12). A circular crestal graben also develops (Fig.3.13). However, the extensional faults which develop as a result of ballooning have relatively small displacements (Jenyon 1985). They serve only to allow uplift of the overlying sequences and do not drop sediments below their depositional regional.

Because sequences over diapirs are relatively thin and salt is effectively incompressible, the difference in compaction between those sequences and sequences deposited in a rim syncline may be large, resulting in major extensional faults. The faults are concentric to the salt structure, but contrast with those produced by doming in that they dip away from the salt dome, and cause sediments to drop below their depositional regional.

Although carapace extension and compaction produce extensional faults with distinctly different features, it may not be possible to determine two fault systems above real diapirs. The two systems are likely to be linked with the top of the diapir acting as a decollement surface (Fig.3.14).

3.3.1(c)(ii) Dissolution collapse stage

When a diapir impinges on an aquifer containing fluids undersaturated with respect to salt, dissolution will commence. If the resulting brines are removed and replaced by fresh fluids dissolution will continue, and collapse of the carapace may result. Jenyon (1985) emphasises that the rate of dissolution and the geometries of resulting collapse structures depend very closely on the "plumbing system" which develops to cause the dissolution. Such systems are finely balanced and may be switched on and off. So the collapse tends to be episodic (Jenyon op.cit.). Two common features are: (i) that faults produced by dissolution collapse down-throw towards the salt body and (ii) displacements on these faults can be large - larger than the down to the crest extensional faults formed during the balooning stage (Jenyon 1985) so that sediments may be downfaulted below their depositional regional (Fig.3.15; compare with Fig.3.13).

Dissolution of salt at the top of a diapir leads to the concentration of impurities and the formation of a "cap rock". The thickness of cap rocks depends on the rate of salt solution, time, and the amount of impurities. Gussow (1968) reports cap rocks up to 300m thick. Their composition reflects the impurities in the salt column, usually gypsum, and the effects of fluid circulation. For example, Thode et.al. (1951, 1953) showed that elemental sulphur and associated limestone, in the cap rocks of oil and gas prone Gulf Coast diapirs were the product of bacterial reduction of sulphates in the presence of hydrocarbons.

3.3.1(d) Structures within diapirs

Structures within diapirs record the mechanisms of salt deformation. Surface work is often limited because of salt dissolution and cap rock formation. The best results have been produced from subsurface studies.

From mines in Gulf coast salt domes Kupfer (1968) records:

> ...isoclinal, attenuated, vertically plunging, refolded and faulted folds resembling those in a handkerchief drawn vertically from the centre through a small ring.

The fold structures which he and Muehlberger and Clabaugh (1968) describe would be termed folded sheath folds in modern parlance. Muehlberger and Clabaugh (op.cit) consider that the sheath folds, and their refolding, are the product of differential salt movement, with the salt moving upwards as a series of localised spines.

Two aspects of microstructural work confirm that deformation is ductile (Muehlberger and Clabaugh op.cit.): firstly, thin sections reveal mylonitic textures, and secondly, plots of salt crystal axes show that a large number of salt crystals are orientated similarly suggesting that movement occurred along glide planes.

For field analysis, the key features of structures within diapirs are that folds and fabrics are expected to be steeply dipping but randomly orientated in diapir cores.

3.4 DIAPIRISM IN THE SOUTH CENTRAL PYRENEES

3.4.1 Introduction

There are two evaporite horizons in the South Central Pyrenees, both of which form diapiric structures: (i) Triassic salt which underlies the entire area (Fig.3.16(a); Riba *et al.* 1986; {2.5.2(b)}); and (ii) Upper Eocene salt, the extent of which is less well known. Busquets *et al.* (1985) show the extent of Upper Eocene evaporite formations in the Ebro basin and state that one formation, the Barbastro Gypsum Formation (B.G.F) is underlain by important masses of salt rock (Fig.3.16(b)). The basin shown by Busquets *et al.* (op.cit.) and the outcrop of the B.G.F., suggest that the B.G.F. and, therefore, the rock salt, underlie the External Sierras in the study area as far west as the Boltana anticline, and occur at depth in the northern part of the Ebro basin. This is in accordance with balanced sections presented above {2.6; 2.7} and sections produced by other workers which show pinch out of the Upper Eocene towards the south (Riba *et al.* in prep.; Fig.1.10(b)).

Four diapir case studies are considered below: three cored by Triassic salt, the Mediano, La Puebla de Castro and Estada diapirs and one cored by Upper Eocene salt, the Barbastro Anticline. In each case study an attempt is made to justify a diapiric interpretation using the diagnostic criteria outlined above {3.3}. Features which were found to be particularly useful were: the presence of salt; evidence of piercement; steep internal fabrics; the development of cap rocks, and large scale geometries compatible with diapirism, in particular rim synclines and extensional faults.

In addition, each diapir exhibits a number of unique features: the Mediano provides an example of dissolution collapse; the La Puebla de Castro diapir is considered to be allochthonous; the Estada diapir {3.4.4} is a salt spine whose cap of gypsum has been reworked into small alluvial fans; and the Barbastro anticline is a linear salt pillow produced by thrust loading.

3.5 THE MEDIANO

3.5.1 Intoduction

South-east of the Embalse de Mediano [6989] the Triassic crops out extensively ($14km^2$). It has extensional fault contacts with the surrounding Jurassic to Miocene sequences and is considered by the author to represent the top of a diapiric piercement structure which has undergone dissolution collapse: it will be referred to below as "the Mediano".

A linked sedimentological and structural evolution has not previously been outlined for the Mediano. Many regional studies consider that the Mediano is an expression of the western lateral ramp of the Cotiella thrust sheet (e.g. Seguret (1972); Garrido and Rios 1972; Deramond *et.al.*. 1984). However, a balanced map (2.8) suggests that the Cotiella thrust sheet does not extend as far south as the Mediano, and that if it did it would be as a thrust with Upper Cretaceous, or younger, rocks in both its footwall and hangingwall. No thrust with this geometry is seen at the Mediano and, furthermore, such a geometry does not involve the Triassic. The Cotiella thrust sheet may have triggered diapiric movement by differentially loading salt below the Tremp-Graus basin but it is not considered to be exposed at the surface in the Mediano area. Camara and Klimowitz (1985) imply this geometry (Fig.2.19). Nijman and Nio (1975) refer to the Mediano as a doubly-plunging anticline, and cautiously make an analogy between facies patterns around the Mediano and modern salt dome islands in the Persian Gulf. Cuevas *et al.* (1985) are more forthright; they consider that the Medinao is a diapiric piercement structure, initiated from the crest of a NNW-SSE trending fold during the Lutetian.

A number of features, in addition to extensional fault contacts at the margins of the Triassic, support a diapiric origin. They are listed briefly below. However, the majority of the following discussion concentrates on documenting a two-stage history for the diapir: a prolonged period of ballooning during the Eocene followed by collapse as a result of salt dissolution during the Oligo-Miocene.

3.5.2 Evidence of diapirism

Diapirism can be demonstrated by the geometry of the carapace and by structures in the Triassic at the core of the Mediano. In outlining a two stage history for the Mediano the carapace is discussed in detail below (see {3.5.3}). So only evidence from within the diapir is discussed here. There are three key features: the presence of salt at depth; the development of a cap rock and steeply dipping folds and fabrics in gypsum.

Although no rock salt was seen at the surface the salt springs which dot the area indicate the presence of salt at depth, e.g. Salinas de Trillo [7587]. The observation is critical as a body will only move diapirically when its density is lower than that of the surrounding sequences, and in sedimentary rocks this requires salt {3.2.1(b)}.

The exposed Triassic is composed of marls, gypsum, limestone, and dolerite, and is closely associated with travertine. The assemblage is considered to represent a cap rock - an accumulation of the insoluble parts of the Triassic sequence following dissolution of salt. The travertine deposits, which include limestones deposited around vegetation, are considered to record spring activity during cap-rock formation and to reflect $CaCO_3$ rich groundwaters.

The Triassic is poorly exposed. So systematic mapping of structures in the manner of Kupfer (1968) {3.3.1(d)} is not possible. However, the gypsum which is exposed is dominated by steeply dipping planar fabrics and steeply plunging folds, compatible with a diapiric emplacement mechanism (Fig.3.17(b)). The retention of steeply dipping fabrics is curious given the evidence for dissolution, which may be expected to produce rotation of gypsum as it moves to accommodate salt withdrawl.

3.5.3 Evidence for a two stage history

3.5.3(a) Ballooning

Several features in the diapir carapace record the ballooning stage: (i) facies changes and unconformities associated with the development of localised carbonate buildups; (ii) slumps which indicate that the area was a growing high; and (iii) the tentative recognition of a rim syncline.

(i) Localised carbonate buildups

Localised carbonate buildups are thought to have developed over the Mediano twice during the Eocene: the Guara Formation (Puigdefabregas 1975) during the lower part of the Cuisian, and the San Martin Formation (Cuevas *et al.* 1985) during the Biarritzian (Fig.2.10(a)). The carbonate buildups are discussed individually below.

Nijman and Nio (1975) suggest that the Alveolina Limestone Formation is also a local buildup over the Mediano, passing laterally into Ilerdian marls (Fig.2.10(a)) but there is no evidence for this as the Alveolina Limestone Formation is present throughout the basin.

3.5.3(a)(i) cont. The Guara Formation

The main outcrop of the Guara Formation is in the External Sierras south of the Jaca basin and at the Boltana Anticline, where it persists until the Priabonian (Puigdefabregas 1975). Nijman and Nio (1975; Fig.2.10(a)) suggest that the Guara Formation at the Mediano is isolated from these sequences by a facies change across the Santa Maria del Buil syncline which separates the Boltana and Mediano structures. However, this cannot be demonstrated as the Guara Formation at the Mediano is unconformably overlain by the San Martin Formation and cannot be traced westwards. The outcrops only constrain north-south variations.

At the large-scale, time-equivalent sediments to the north are slope deposits of the Montanana delta, so the Guara Formation does have limited extent.

At the Mediano three outcrops suggest that the Guara Formation thins and onlaps towards the south. Firstly, using field glasses at the north-south trending Santa Barbara anticline [7189] the Guara Formation can be seen to onlap the Alveolina Limestone Formation towards the southeast (Fig.3.18). The Guara Formation is at its thickest at the axis of this anticline, commencing with grey glauconitic marls, which pass upwards into well-bedded limestones and are capped by a distinctive brown-weathering bioclastic limestone. Secondly, the Guara limestone thins towards the south on the western limb of the Santa Barbara anticline.

Thirdly, on the eastern side of the Tozar de Salinas anticline [7588] a brown limestone similar to that seen at the top of the Santa Barbara anticline, directly overlies the Alveolina Limestone Formation suggesting that the grey marls were not deposited there.

In summary, the evidence is not incompatible with the model suggested by Nijman and Nio (1975; Fig.2.11) that the Guara Formation was deposited some distance offshore from the Montanana delta around a high produced by growth of the Mediano diapir. However, because east-west facies changes cannot be determined, the alternative possibility of a carbonate shelf extending from the External Sierras in the Jaca basin to the Mediano cannot be excluded.

In detail, the field evidence presented above differs from that of Nijman and Nio (1975) - they show no unconformity between the Alveolina Limestone and Guara Formations, but do recognise an unconformity at the top of the Guara Formation which they consider to be overlain by a localised sandy deposit. The author did not recognise the higher unconformity, and Nijman and Nio (op.cit.) do not show it on their map (Fig.2.10(b)). It is in a position on their section which appears to correspond with the brown bioclastic limestone taken here as being the top part of the Guara Formation.

3.5.3(a)(i) cont. The San Martin Formation

A number of formations, including the San Martin Formation, developed in response to a transgression of Biarritzian age. Cuevas *et al.* (1985) recognise four formations (Fig.3.19): the Capella Formation, equivalent to the top part of the Santa Liestra depositional sequence of Mutti *et.al* (1985a) {2.5.3}; the Pano Formation, a sandstone-marl sequence interpreted as having been deposited during transgression by a shoreline reworking the Capella Formation (Donselaar and Nio 1984); the San Martin Formation, a fully marine shallow water limestone deposited at the height of transgression, and the Escanilla Formation produced by the westwards progradation of coastal and continental systems sourced from the Pyrenean chain.

The formation names of Cuevas *et al.* (1985; Fig.3.19) are followed here with two exceptions: firstly, for consistency (see {2.5.3}), the term Santa Liestra depositional sequence is preferred to Capella Formation, and secondly, the Escanilla Formation is redefined to describe only the continental deposits overlying the San Martin Formation (Fig.2.15). The term "Belsue Formation" (Fig.2.15; Puigdefabregas 1975; Nijman and Nio 1975) is used for the shallow-marine siliciclastic sequences which immediately underlie the Escanilla Formation.

The San Martin Formation outcrops on both sides of the Mediano. At the N.W. margin, just south of the Embalse de Mediano, the syntectonic localised nature of the carbonate buildup can be clearly demonstrated. From conformably overlying marls at the top of the Ainsa basin at Barrio Nuevo de Mediano [6888] the limestone marking the base of the San Martin Formation becomes unconformable towards the south, overstepping the Guara, Alveolina Limestone and Tremp Formations to rest on Cretaceous limestones dipping at 46° to the N.W. above Samitier [700874]. At this locality the San Martin Formation is composed of two limestone horizons each around 30m thick, separated by a grey marl of similar thickness (Fig.3.20(a)). Both limestones contain a diverse fully marine fauna and are massive, except that the base of the lower limestone is composed of boulders up to 1.4m in diameter (Fig.3.20(b)). Towards the west the marl between the limestones thickens dramatically, and, furthermore, both limestones pass laterally into different facies: the lower limestones pass into marls at the top of the Ainsa turbidite basin and the upper limestones pass into shallow marine shoreline sequences of the Belsue Formation.

The basal unconformity is interpreted as representing a phase of uplift which increased in magnitude towards the Mediano and occurred immediately prior to deposition of the San Martin Formation. The stratal wedging seen in the marl between the two limestone horizons suggests that the uplift continued during deposition of the San Martin Formation. This is compatible with the large boulders seen at the base of the first limestone horizon, which are interpreted as olistoliths shed from San Martin limestone deposited at the core of the diapir to the south-east.

At the eastern margin of the diapir exposure is less complete, and relationships are complicated by extensional faulting and by interdigitation of facies characteristic of the Pano and San Martin Formations. Nevertheless, an unconformity can be seen all along the eastern margin of the diapir between the gently dipping Pano Formation and folded Upper Cretaceous limestones (e.g. [7482]). In addition, the Pano Formation thickens away from the centre of the Mediano, from 160m below Misuales [7481] to 360m below San Martin [7580]. The thickness change and the gently dipping nature of the Pano Formation together suggest that the unconformity surface dips to the east.

Thus both the western and eastern parts of the Mediano are characterised by marginal unconformities and thinning of sequences towards the centre of the structure suggesting uplift during the Biarritzian. A similar unconformity can be inferred from the S.W. margin of the the Mediano, to the east of Escanilla, i.e. the unconformity appears to almost encircle the Mediano - a geometry which strongly suggest diapir ballooning.

It would be wrong, however, to suggest that the San Martin Formation is restricted to the Mediano. Although its extent in the N.W. is delimited by an ancestral Mediano high, in the south it extends considerably beyond that structure, and has a constant thickness in an east-west sense (i.e. it does not thin towards the Mediano) but thins together with the Pano Formation steadily towards the La Puebla de Castro diapir {3.6}.

3.5.3(a)(ii) Slumps

Federico (1981) considers that more than half of the Ainsa basin is composed of slumps. On a series of palaeogeographic maps, from the Lower Ilerdian to the Upper Lutetian, he shows that the slumps were predominantly derived from the slopes of the Mediano and the Boltana anticline, and, furthermore, that slumping commenced during the upper part of the Lower Lutetian - i.e. coincident with the main phase of ballooning of the Mediano, suggesting that they were diapirically triggered(Fig.3.21; Fig.2.10).

3.5.3(a)(iii) Rim synclines

There are two sequences which may have been deposited in rim synclines: the Pano Formation on the eastern side of the Mediano; and the Belsue and Escanilla Formations in the Santa Maria del Buil syncline on the western side.

Cuevas *et al.* (1985) consider that salt withdrawal enhanced the Biarritzian transgression in the Pano area; implying that the Pano Formation was deposited in a rim syncline. Thickness variations in the Pano Formation appear to support this: it is thickest at San Martin (360m) and thins both towards (160m at Misuales) and away from the Mediano (160m at Panillo [7680]).

However, Donselaar and Nio (1984) and Cuevas *et al.* (1985) consider that the Pano Formation is equivalent in time with the top part of the Santa Liestra depositional sequence. Thus, although the Pano Formation thins away from San Martin, it is not a time bound unit, so the thinning may be more apparent than real. Furthermore, the exposed beds in the Pano Formation, and Santa Liestra depositional sequence do not have a synclinal geometry: they are flat lying and support an alternative interpretation, that the Pano Formation onlaps the folded margin of the diapir.

Sediment thicknesses and bed dips in the Santa Maria del Buil syncline are compatible with deposition in a rim syncline. In particular, the Belsue Formation thins gradually towards the Mediano indicating increased subsidence at the syncline core (Nijman and Nio 1975; Fig.2.10(a)). The Escanilla Formation has a similar geometry, although this may be the result of preferential preservation in the core of the syncline below the unconformity at the base of the overlying Collegats Group - the author knows of no systematic study of sediment thickness across the syncline.

Independent evidence shows that the main folding of the Boltana anticline occurred at this time (Puigdefabregas 1975). So although salt withdrawl at the Mediano may have contributed to the development of the Santa Maria del Buil syncline it was not the sole cause.

The inferred rim syncline deposits are too poorly constrained to be used to define stages of pillow formation and piercement. However, the marked overstep at Samitier [6987] and the unconformity which almost encircles the Mediano, suggests that

Biarritzian sediments unconformably overlay Triassic rocks in the diapir core. So the Mediano was a piercement structure at this stage, and furthermore, the Triassic had reached the surface. In doing so it laid itself prone to dissolution and collapse.

3.5.3(b) Dissolution Collapse

Salt springs indicate contemporary salt dissolution at the Mediano. Since the diapir is thought to have reached the surface and come into contact with seawater during the Biarritzian {3.5.3(a)} dissolution is likely to have been in progress since the Upper Eocene.

The author considers that the major faults which cross-cut the Mediano are the result of dissolution induced collapse, and furthermore, that they controlled sediment thicknesses and facies patterns over the diapir from the Priabonian through to the Miocene.

The discussion which follows is in two parts: firstly, it argues that the faults are compatible with an origin by dissolution collapse, and secondly, it documents phases of active faulting by discussing variations in the Biarritzian and later sediments.

3.5.3(b)(i) Fault Geometries

The faults which cross-cut the Mediano define four distinct grabens: the Palo graben in the north [7388]; the Escanilla graben in the west [8469]; and the Trillo graben in the N.E. [8775] linked via the Caneto fault system [7485] to the Fuente Lobata graben in the south [7477].

The graben-bounding faults downthrow towards the core of the diapir and displacement on them is large - sufficient to downfault the Pano and younger Formations to positions below their depositional regional. For example, along easting 79 the base of the Oligo-Miocene is downthrown 500m to the west across the Caneto fault system. Of the three ways in which diapirs may form faults (see {3.3.1(c)}) the large displacement and the geometry, downthrowing to the core of the diapir, is only compatible with an orgin by dissolution collapse.

Mesoscale (m's and cm's) faults from the diapir carapace support this interpretation: all the measured faults are steeply-dipping (Fig.3.17(a)(i)); and slickenslides indicate that the majority of them have a large dip-slip component (Fig.3.17(a)(ii)). Some slickenslides indicate strike-slip movement. However, two points indicate that this should not be used to imply a transtensional origin for the large-scale faults: (i) the maximum displacement on the large faults occurs at the core of the Mediano, they are not systematically orientated, and do not link into a strike-slip zone; and (ii) none of the faults, not even at the mesoscale, show any sign of compression - given the range of observed fault orientations, and the sensitivity of strike-slip systems to slight deviations in fault orientation, major compressional structures would be expected at obliquely orientated faults (S. Knott pers. comm. 1987).

3.5.3(b)(ii) The timing of faulting and patterns of sedimentation

As argued above relationships at the N.W. corner of the diapir demonstrate that the San Martin Formation and the laterally equivalent, lower part of the Belsue Formation were deposited during the ballooning stage. At the eastern margin of the Mediano the Belsue Formation thickens steadily from south to north and is interpreted as having been deposited between the ballooning and collapse phases. The Escanilla Formation is the first sequence to be affected by dissolution collapse, with the Palo graben and the northern part of the Caneto fault system developing to control sediment dispersal patterns. By the Oligo-Miocene the dissolution collapse faults were well developed controlling sediment thickness and facies patterns in the Collegats Group. The following discussion documents the evidence for this history, detailing the Belsue, Escanilla and Collegats sequences in turn.

3.5.3(b)(ii) cont. The Belsue Formation

The Belsue Formation decreases in thickness from north to south across the Mediano (Fig.3.22) from 100m in the Trillo graben to 16m below Tozal Panchado [765783]. This thickness change may be influenced by the same factors which produce southwards thinning in the Pano and San Martin Formations {3.6.}. It could also be due to synsedimentary movement on faults produced by dissolution collapse. This alternative hypothesis can be tested by examining east-west thickness changes across the Mediano.

Graphic logs through the Trillo and Palo grabens suggest a marked thickness reduction in the Belsue Formation from Trillo to Palo (Fig.3.22): the section through the Palo Graben ([733893 - 736885]) shows a thin wave rippled sequence and no San Martin Formation. However, the section is cut by an extensional fault, significant displacement on which could have produced thickness variations. A steeply dipping fault block just west of Ermita de Santa Brigada [727883] supports the second of these possibilities. The block is based by the San Martin Limestone and overlain by 100m of sandstone and marls of the Belsue Formation. Since neither the San Martin Formation nor the coarse sandstone are seen in the log directly south of Palo the aforementioned extensional fault is considered to have had a large throw cutting out these formations. So the wave rippled silts at Palo stratigraphically underlie the San Martin Formation, and the measured thickness variation between Trillo and Palo is apparent and not real. The few palaeocurrents which were obtained from the measured sections through the Belsue Formation suggest that sediment dispersal systems were west-directed, axial to the basin, and not affected by north-south faults (Fig.3.22).

Two conclusions can be drawn from these observations: firstly, since the Belsue formation is of constant thickness in an east-west sense over the Mediano, there is no evidence for extensional faulting during deposition, and secondly, the bounding fault to the Palo graben has a bucket shape, closed to the north.

3.5.3(b)(ii) cont. The Escanilla Formation

Thickness variations in the overlying Escanilla Formation are difficult to assess because the formation is unconformably overlain by the Oligo-Miocene Collegats Group.

Published maps (IGME sheet 23; SNPA 1972) suggest that the Escanilla Formation thins towards the Mediano across the Santa Maria del Buil syncline. Nijman and Nio (1975; Fig.2.10) suggest that the thickness decreases steadily to zero at the eastern side of the Mediano. The author is unsure to what degree this thinning is depositional or a consequence of the unconformity at the base of the overlying Collegats Group {3.5.3(a)(iii)}. Van Buchem *et al.* (1985) suggest that the decrease in thickness is depositional: they show the Mediano as a high defining the eastern margin of a basin, the "Sobrarbe basin", which essentially corresponds to the Santa Maria del Buil syncline (Fig.3.23). However, significant thicknesses of Escanilla Formation occur in fault blocks within the Mediano: the author records 246m in the Pano graben, and 233m in the Trillo graben (Fig.3.2?(a)&(b)), and a map estimate suggests that 750m of section is preserved in the Bodiello area west of the Caneto fault system. These thicknesses preclude the interpretations of Nijman and Nio (op.cit) and van Buchem (op.cit) that the Mediano was a simple high during the deposition of the Escanilla Formation.

The key thickness variation in the Escanilla Formation is that between the hangingwall and the footwall of the Caneto fault system: 19m in the footwall at Tozal Panchado [764783] and perhaps as much 750m within the diapir core. Is the thickness difference the result of syn-sedimentary faulting, or is it the result of preferential preservation of the Escanilla Formation in grabens formed immediately prior to deposition of the Collegats Group? Sedimentological analysis gives some support for syn-sedimentary faulting.

Palaeocurrents in the Escanilla Formation are shown on Figures 3.22 and 3.24, and on the map (sheet 1). Palaeocurrents from the Trillo and Palo grabens are north directed, a 90° switch from the west directed palaeocurrents in the underlying Belsue Formation, suggesting that dissolution collapse had commenced and that sediment dispersal was controlled by the resulting faults: the Caneto fault system and the Palo graben bounding faults respectively. By contrast, palaeocurrents at Bodiello have a radial pattern ranging clockwise from the S.E. to the N.E., but directed dominantly towards the west at 90° to the Caneto fault system. These palaeocurrents, and other west directed palaeocurrents ([6879]; Fig.3.24) are interpreted to reflect the main direction of sediment dispersal during the Escanilla

Formation (syn "Campodarbe Group") which Puigdefabregas (1975) and Atkinson (1983) consider to be axial to the basin towards the west or W.N.W. (Fig.2.13(6)). The lack of an obvious effect of the Caneto fault system may be explained in two ways: firstly, by invoking that the Bodiello area was the location of a major sediment transport path, and that sediment dispersal was radial from this point once the Caneto fault system had been traversed, and secondly, by considering that, at this time, displacement on the Caneto fault system diminished southwards, so that it did not form surface features and, therefore, did not affect sediment dispersal. The second interpretation is preferred as it also explains the west directed palaeocurrents at La Corona [6985] at the N.W. margin of the diapir.

The stacked sediment gravity flow units composed of Alveolina Limestone and Tremp Formation clasts which occur within the Escanilla Formation and outcrop at [696754] are interpreted as locally sourced alluvial fans {2.6.3(c)}. They indicate uplift at the margins of the Mediano during deposition of the Escanilla Formation. Uplift may be the product of a thrust front or diapirism {2.8.3(b)(iii)} i.e. diapiric uplift may have continued in the S.E. while dissolution collapse was initiated in the north.

3.5.3(b)(ii)cont. <u>The Collegats Group</u>

The outcrops of the Collegats Group which surround the Mediano are discussed in detail below {4.4.7}. This account concentrates on features which: (i) help to date phases of faulting, and (ii) demonstrate major fault-induced changes in palaeogeography.

The most dramatic of these features are seen on the Caneto fault system in the Fuente Lobata graben [7476] where two wedge-shaped fault blocks record three phases of faulting during deposition of the Collegats Group:

The first two phases are recorded in a wedge-shaped block below Tozal Panchado [750775]. The block is composed of Escanilla sediments unconformably overlain by the Collegats Group. It is bounded to the west and N.E. by faults, and is overlain by an intra-Collegats Group unconformity in the S.E. (Fig.3.25). There is no equivalent footwall sequence for the Collegats Group and the Escanilla sequence is much reduced in thickness (19m at Tozal Panchado [764783]). The unconformity at the base of the Collegats group and the increased thickness of the Escanilla Formation from the footwall to the hangingwall across the N.W.-S.E. orientated fault, is considered to indicate a pre-Collegats Group phase of faulting. The intra-Collegats unconformity is thought to be caused by a second phase of fault movement during the Oligo-Miocene.

Palaeocurrents and clast provenance considerations can be used to refine the interpretation of the second phase of faulting {4.4.7(c)}. A section measured through the Collegats Group within the block shows that the basal 87m is dominated by fluvial sediments sourced from the Axial Zone with palaeocurrents indicating flow to the south (Fig.4.74). Above this level palaeocurrents are dominantly towards the west, and sediments are increasingly derived from local sources: tufa deposits and the San Martin Formation, exclusively so above the intra-Collegats unconformity. The changes are interpreted as the product of a phase of fault activity on the N.W.-S.E. trending fault, which increased in intensity up to the development of the unconformity. The south-directed palaeocurrents in the lower part of the sequence are thought to be related to relief generated by the Caneto fault system. The increased percentage of locally sourced material is thought to be the product of rejuvenation of that relief with tufaceous sediments developing at this time as fault-scarp spring deposits. The change in palaeocurrent direction is similarly related to increased faulting and thought to be the result of capture of part of the Graus transfer system.

The second v-shaped fault block, the Secastilla graben [748766], records the third phase of faulting. Coarse sediment bodies within the graben thicken towards its western margin, and fault displacement appears to die out upwards in time suggesting that movement on the faults was syn-depositional {4.4.7(d)}.

In addition to these well documented examples, syn-depositional faulting of Collegats Group age can be tentatively inferred from the Palo Graben {4.4.7(b)}. The thick sequence of extremely coarse sediments which is preserved there is only compatible with deposition in a confined fluvial transfer system early in the

Collegats Group. A detailed assessment of the palaeogeography shows firstly that the footwall of the Caneto Fault system acted as a major interfluve at this time, so that the fluvial transfer system probably developed entirely to the west of the fault system. Because of these factors and because the Palo-Graben is known to have already existed at this time the fluvial transfer system is thought to have been confined by syn-sedimentary movement on the Palo Graben faults.

3.6 THE LA PUEBLA DE CASTRO DIAPIR

3.6.1 Introduction

The La Puebla de Castro diapir is located southwest of Graus between the Rio Cinca and Rio Esera and named after the village of La Puebla de Castro [7569]. Evaporites at the diapir core have an outcrop area of $16km^2$ and are considered to be Triassic in age.

The discussion which follows is in five parts. The first part documents the range of evidence for diapirism, the second considers the age of evaporites at the core of the diapir. The third part discusses the sedimentary sequences around the diapir and how they can be related to phases of diapirism. The fourth part discusses a complex three-stage origin fo the diapir: (i) initiation as a pillow, or as a buckle fold, cored by Triassic salt during the Middle to Upper Eocene; (ii) emplacement of the structure on an Upper Eocene decollement and development of a recumbent buckle fold, probably during the Oligo-Miocene, and (iii) development into a piercement structure.

The final part discusses the deformation of Triassic limestone and gabbro bodies that occur within the diapir.

3.6.2 Evidence for diapirism

Three features characteristic of the ballooning stage of diapirism {3.3.1(c)(i)} are recognised in the La Puebla de Castro diapir: firstly, active salt springs, from which salt can be inferred to exist at depth [738737] (Fig.3.26(a)); secondly, intrusive contacts between Triassic evaporite and Oligo-Miocene Collegats Group at the margins of the diapir [734733] (Fig.3.26(b)); and thirdly, the fact that the evaporites are dominated by steeply dipping fabrics and by steeply plunging fold and lineations (Fig.3.27(a)).

By contrast with the Mediano there is no evidence for dissolution collapse of the diapir. This is true despite evidence for (i) contemporary salt dissolution, and (ii) dominance of the surface outcrop by marls, limestone and gabbro which, using the argument presented in {3.5.2}, are interpreted as a cap rock.

3.6.3 Age of evaporites

A Triassic age for the evaporites at the surface of the La Puebla diapir is indicated by three features: firstly, by the inclusion of Triassic limestone and gabbro bodies within the evaporites, e.g. at the village of La Puebla de Castro and at the deserted hamlet of Bolturina [730717] respectively; secondly, by Cretaceous limestones which overlie the evaporites at the Sierra de Urbiego [7370]; and thirdly, by their distinctive red-mauve olour - the only other gypsum body in the area, the Barbastro Gypsum Formation {3.8} is grey-white.

However, there is also some evidence to suggest that the diapir may be cored at depth by Upper Eocene salt (see below {3.6.5}).

3.6.4 Sedimentary sequences

Two sequences are considered to have been affected by the La Puebla diapir: the Pano and San Martin Formations of Upper Eocene age, and the Collegats Group of Oligo-Miocene age.

The Pano and San Martin Formations {3.5.3(a)(i)} thin towards the La Puebla diapir, from San Martin in the north [7580] to near Graus in the south [7975] (Fig.2.17(e)). The thinning geometry is closely comparable to that shown by sedimentary sequences deposited in primary rim synclines {3.3.1(b)}, suggesting that the diapir was a salt pillow during the Upper Eocene. As noted above {2.6.3(b)} a pillow geometry is also compatible with observed thickness variations in the Capella Formation which directly underlies the Pano Formation (Fig.2.17(d))

The major thickness changes in the Collegats Group relate to the Secastilla graben and are the result of syn-sedimentary movement on the Caneto fault system {3.5.3(b)(i)}. This fault system extends almost to the northern margin of the La Puebla diapir. The faults are extensional, downthrown towards the diapir, their displacement is significant but dies out towards the diapir. Beds within the graben dip towards the south.

The geometry of the graben faults and graben fill does not accord with any of the extensional fault systems produced by diapirism {3.3.1(c)}, suggesting they were not produced by the La Puebla diapir: (a) faults produced by ballooning have small displacements and maximum displacement over the diapir crest; (b) faults produced by compaction are downthrown away from the diapir core; while (c) faults produced by dissolution collapse have aximum displacement over the diapiric body, causing rotation of beds towards the diapir core. The geometry of the faults and ,raben fill is more compatible with an origin due to dissolution collapse at the Mediano {3.5.3(b)}.

Palaeocurrents derived from the Secastilla graben are persistently directed towards the south, suggesting that the La Puebla diapir had little effect on sediment dispersal patterns in he area (Fig.4.76). However, outcrops adjacent to the diapir ield palaeocurrents which suggest that flow was directed around the diapir towards the west, at least in the very youngest part of the Collegats Group. It is possible that the outcrops from which most of the palaeocurrents are taken were too far from the margins of the diapir to feel its effects.

Recent(?) fluvial sediments outcrop to the west of the S castilla [7373]. They are erosive into the underlying gypsum nd Collegats Group, and have a geometry which suggests a palaeovalley fill. The valley is orientated east-west parallel to the margins of the La Puebla diapir, and has been partially exhumed by the modern Barranco de Selva - the position of both valleys is considered to have been structurally controlled by the diapir margin. The recent sediments are flat lying and undeformed (Fig.3.26(b)). So there is no evidence for contemporary diapirism in this area, and the diapir must have been emplaced prior to the development of the palaeovalley.

3.6.5 Origin of the diapir

A number of key points can be used to develop a complex three stage history for the diapir. A balanced section through the area {2.6} and the inferences about timing which have been made in the preceeding section are particularly important. However, the evidence on which the history is based is largely interpretational. So it must be considered speculative.

The balanced section shows that the diapir forms part of the hangingwall of a large thrust sheet (thrust sheet (2)) and furthermore, that it is allochthonous resting upon a decollement within an Upper Eocene salt horizon, thought to underlie the

Barbastro Gypsum Formation (Fig.2.17(e); {3.8; 2.6.3(c)}). A regional consideration of syntectonic sedimentation and balanced s ction constraints {4.5; 2.6; 2.8} suggests that the diapir was e iplaced on the Upper Eocene decollement early in the deposition of the Collegats Group, prior to the deposition of the upper C llegats sediments into which the diapir is intrusive. So the pir would have been detached from its Triassic mother bed b fore it had finished acting in a diapiric fashion. Such a eparation is likely to have terminated salt movement (Jenyon 1985). So the continued diapirism may have been produced by robilization of the Upper Eocene salt i.e. the diapir could be cored by salt of Upper Eocene age at depth.

Since the evaporites, limestone and gabbro at the surface are Triassic in age, it is likely that the diapir developed prior to emplacement on the Upper Eocene decollement. The possibility that the Eocene Capella, Pano and San Martin Formations were deposited 'n a primary rim syncline to the La Puebla diapir supports this interpretation. The structure at this stage may have been either a salt pillow or a salt cored fold.

The major south vergent recumbent fold (Fig.2.17(e)) is considered to have developed during emplacment of the diapir on to the Upper Eocene decollement. An origin by buckling is supported by the fold vergence and by the joint pattern in the overturned limb at Urbiego [7370] which suggests strike perpendicular compression.

The development of the diapir as a piercement structure is likely to have occurred during emplacement. So the palaeocurrents in the Collegats Group are expected to have been affected by the structure. There is some evidence to support this, but the effects appear to have been strongly localised, confined to the margins of the diapir. Piercement of the upper part of the Collegats Group predates the recent sediments seen to the west of Secastilla. There is no evidence for contemporary diapirism.

3.6.6 Limestone and gabbro bodies

Several limestone and gabbro bodies occur within the diapir core, forming low isolated hills 100-400m long and 50-300m wide. The margins of the bodies are poorly exposed.

The limestones are well bedded on a dm scale, dark grey in c lour and micritic in composition. Individual limestone bodies are coherent slabs, which either have c nstant d p or are gently f lded. Block orientation and dip e n m, but folia ions in a jacent gypsum parallel block m rgi s. I nally, the only idence of deformation are small- le faul s nd well developed joint sets which show no co t nt o ta o but tend to be eply dipping (Fig.3.27(b)). H w v , h ple set is small.

The gabbroic bodies are ma sive d h e no obvious shapes. Internally, they are cross-cut by a pletho a of randomly orientated joints and faults which ha e bec me s tes of eferential weathering (Fig.3.27(c)).

The limestone and gabbro bodies re considered to have been formed during diapirism by the breakup of layers of limestone and gabbro interbedded with the Keuper and Mushelkalk evaporites. Fabrics within the gypsum indicate that they deformed in a ductile fashion {3.6.2; 3.3.1(a)}. The interbedded limestones and gabbro responded to the same stress regime by deforming in a brittle fashion, in a manner comparable to fractured grains in more conventional mylonites. The approximately constant size of the gabbro and limestone bodies, and the lack of evidence of breakup of the bodies suggests that their dimensions were reduced to an equilibrium size, governed by the flow stress in the ductile matrix and the fracture strength of the gabbro and limestone masses.

The lack of consistent fracture patterns in both the limestones and the gabbros suggests that they were rotated several times during deformation, with new fractures forming in each position.

3.7 THE ESTADILLA DIAPIR

3.7.1 Introduction

The Estadilla diapir is located along the Rio Cinca just to the south of the External Sierras [7160] map 1. Evaporites outcrop in two areas, to the west of Estadilla and at Estada; each outcrop is surrounded by red beds of Oligo-Miocene age {4.4.4},

The following discussion addresses: firstly, the evidence for diapirism; secondly the age of the evaporites, which are considered to be Triassic at the surface but may be Upper Eocene in age at depth, and thirdly, evidence at the northern margin of the Estada outcrop for sedimentation during diapir intrusion in the form of stratal wedging and discrete fans of detrital gypsum.

3.7.2 Evidence for diapirism

The evidence for diapirism {3.3.1} is three-fold: (i) the structure is cored by Triassic evaporites which are intrusive on nearly all margins into Oligo-Miocene red beds; (ii) the evaporites are dominated by steep fabrics in the core of the structure and at vertical margins but exhibit low-angle flattening fabrics near the top of the diapir (Fig.3.28(a)&(b)); (iii) the diapir carapace is dominated by extensional normal faults, both strike-parallel and strike-perpendicular, consistent with diapir ballooning (Fig.3.28(b)). However, no systematic microtectonic study has been caried out in this area.

Unfortunately, neither salt nor salt springs were seen within the Estada diapir. However, salt can be inferred to exist at depth given the strong evidence for diapirism, and the subcrop maps of the Triassic and Upper Eocene salt horizons.

3.7.3 Age of evaporites

As with the La Puebla and Mediano diapirs three features indicate a Triassic age for the evaporites seen at the surface within the Estadilla diapir: their distinctive colour; the inclusion of Triassic limestones and gabbros, and the fact that the evaporites underlie Cretaceous limestones.

Following the same arguments advanced for the evolution of the La Puebla diapir {3.6.5}, the Estadilla diapir may also be cored by Upper Eocene salt at depth.

Γhe initiation of diapirism at Estada is difficult to assess. ınce the Triassic is considered to be allochthonous the ⁴ructure is thought to have predated emplacement on the Upper Eocene decollement. At this st it ould ha e been a salt-cored f d comparable to those seen thoughout the External Sierras, ıly developing as a diapir o e emplaced upon the decollement nd buried by Miocene sequences

7.4 The northern margin of th t d'lla d'ap'r

The northern margin of the E tad l a d' p'r is well exposed at Estada in two road cuts, the old ro d from [715617] to [718617] and a new road, not shown on the map from [717614] to [718617]. The margin is defined by vertically dipping Oligo-Miocene sediments intruded by Triassic gypsum. The sediments form the uppermost part, formation (d) of the Estada Group {4.4.4}. Away from the contact their dip decreases steadily as a result of stratal wedging, and at least ne discrete unconformity (Fig.3.29). The geometry is interpreted as the product of diapiric intrusion during sedimentation, producing uplift and tilting towards the north through time. The unconformity suggests that uplift rates varied through time.

The sediments are of two types: a fine grained sand and siltstone facies assemblage and a conglomerate facies composed of gypsum and limestone clasts.

The fine-grained sandstones occur in sharp-based lithosomes up to 4m thick, separated by siltstones. The sandstones are well sorted and exhibit trough and ripple cross-lamination. Palaeocurrents indicate flow parallel to the diapir margin dominantly to the west.

The sandstones are considered to be in-channel fluvial deposits, and the siltstones are thought to be their overbank equivalents. The regional palaeogeography suggests that the rivers flowed southwards {4.5}. The palaeocurrents are considered to reflect the presence of the Estadilla diapir within the flood plain causing flow to divert towards the east and west around the structure.

The gypsum and limestone conglomerate facies assemblage occurs a two discrete bodies that punctuate the sandstone/siltstone facies assemblage [717616; 715618]. Each of these bodies is sharply based and around 20m thick. The older of the two bodies s the best exposed (Fig.3.29). It is mposed of 1 to 24cm thick e sively based beds. The beds are dominated by 1-2cm diameter c' sts of gypsum, but 0-10/ are angular clasts of Triassic l' estone. In many cases the gypsum cl sts a e recrystallised. So conglomerate textures are d' f'cult to as ss, ex ept where there 's a high percentage of limeston c sts. Beds of that type show ndomly orientated, bed-p r l l and imbr cated a(t)b(i) clast f rics. There is no tendancy for mes one clasts to be at the top or base of beds. The imbri ation suggests flow towards the N.E.

The conglomerate facies are interpreted as the product of a variety of sediment gravity flow p ocesses. The erosive bases and the a(t)b(i) imbrication are ind'cative of fluidal sediment gravity flows {4.2.1(d)}. However, not all the beds have these characteristics, those with random and bed-parallel internal fabrics could have been emplaced as debris flows {4.2.1(b)}. The facies assemblage is compatible with deposition on the proximal part of a small alluvial fan {4.3 2}.

Provenance and palaeocurrents considerations suggest that the fan was sourced from the adjacent Estadilla diapir. The sharp bases of the two fan bodies are thought to reflect tectonic triggering of fan activity: discrete phases of diapiric uplift each producing discrete pulses of sediment.

In summary, stratal wedging in sediments at the northern margin of the Estadilla diapir records northwards tilting of the margin during sedimentation. An unconformity within the sequence and two discrete fan bodies sourced from the diapir suggest that the rate of tilting periodically increased. The uplift controlled the local pattern of sedimentation, generating the two fan bodies and diverting the main south flowing, fluvial distributary system to the west and east around the diapir.

3.8. THE BARBASTRO ANTICLINE

3 1 Intoduction

he Barbastro anticline is a major F W. trending anticline at t e northern margin of the b o b in. I s st westerly point is s me 10km west of Barbastro. F m h re to the Segre fault, with e exception of a bifurcati n o M n, the anticline urs as a single fold at the t e n arg'n f he External S erras. Eastwards of the Seg e l the t re turns s ıtheastwards and bifurcat s thr e me at Vil nova de la A uda, Cardona and Suria (ig . 0) At the surface the anticline i cored by well bedded gyps m f the Barba tro Gypsum Formation a d flanked by Oligocene fluv al sediments (SNPA 1972).

The discussion which follows addresses two points: firstly, the evidence for diapirism, which suggests that the structure is a salt pillow, and secondly, a model whi h invokes thrust loading as a trigger for initiation of diapirism.

3 8.2 Evidence for diapirism

There are four points which demonstrate a diapiric origin for the Barbastro anticline.

Firstly, two structural sections controlled by borehole and seismic data suggest that the evaporites thicken markedly towards fold cores: a section passing through Binefar (Riba *et al.* in prep.; Fig.1.10(b)) and a section at the eastern end of the anticline which traverses the Cordona and Suria anticlines (Busquets *et al.* 1985; Fig.3.30). The geometry is strongly indicative of diapirism.

Secondly, Busquets *et al.* (op.cit.) suggest that the Barbastro Gypsum Formation is underlain by important masses of rock salt. This is crucial as diapirism in compacted sedimentary sequences will not occur without salt.

Thirdly, isoclinal medium-scale folds occur in the core of the diapir with vertical axial planes and apparently, no consistent trends (although no systematic study was carried out on this; Fig.3.31). Their geometry is closely comparable to folds described from the cores of other diapirs (3.3).

Fourthly, joints in the Oligocene sediments which overlie the evaporites are dominated by a bed normal, strike-parallel extensional set. Such joint sets are best explained by bending olds of the type produced by diapirism and cannot be formed by c mpression perpendicular to f d xes {3.3.1(c)}. However, this extensional set is predat d by a which indic tes early strike perpendicular compresson {3.8 3}.

The diapir is interprete a a s lt p llow rather than a ercement structure beca e l l'go ene fluvial sediments are onformable with the Barb tro Gy um Forma 'on. The contact is not intrusive. Along the R' E ra the gypsum and fluvial facies are interbedded.

3.8.3 A model for the initi 'on of diapirism

Balanced sections across the Barbastro anticline show the External Sierras to be allochthonous upon an Upper Eocene evaporite horizon (Fig.2.17(e); 2.20(e)} - the salt horizon considered to underlie the Barbastro Gypsum Formation.

Emplacement of the External Sierras as thrust sheet (2) during the latest Oligocene and Miocene is considered to have triggered salt movement as a result of differential loading ({3.2.1(a)}; Fig.3.5). The salt flowed to the edge of the thrust sheet and accumulated to form a salt pillow seen today as the Barbastro anticline. The early strike-perpendicular joint set inducates that the structure may have initiated as a salt cored buckle-fold; alternatively, this joint set could represent the regional stress field in front of the alpine thrust front (c.f. Bevan and Hancock 1986).

The model suggests that the structure developed at or after the end of the Miocene. This is in accordance with the observations of Hirst (1983) who records no effects of the Barbastro anticline on a large alluvial fan, the Huesca fan, which was deposited in the Miocene {4.3.2; 4.5}.

3 9 CONCLUSIONS

This study of selected d'apirs in the South Central Pyrenees allows comment on four features of diapirism: (a) how diapirism an be recognised; (b) the effect of diapirs on their carapace nd syn-diapiric sediments; (c) mechanisms of diapir initiation; d (d) the level in the crust to whı h diapirs intrude.

) The recognition of d ap' m. 'x f at res were found to be seful: (i) evidence for h l 'n e re f the structure, 'ther in terms of salt spri gs or g log al maps of salt asins; (ii) thickening of ev porit s towards the core of a given structure; (iii) intrusive conta ts be ween evaporites and younger rocks; (iv) assemblages indi a ıve of cap rocks - an accumulation of impurit'es with'n the iapiric sequence and tufaceous deposits; (v) structures within evaporite bodies (steeply dipping fabrics and folds) are supportive of diapiric activity, but exposure is oft n too poor to study them systematically; and (vi) feat res in the carapace of the evaporite bodies.

(b) The effects of diapirs on their carapace and syndiapiric sediments. Five main effects have b en recognised: (i) unconformities may devel p over the crest of the diapir indicating localised uplift; (ii) syn-diapiric sequences thin towards diapir crests forming cumulative wedge systems (Riba 1976a&b; Anadon *et al.* 1986) which indicate prolonged diapir growth; (iii) facies changes (e.g. localised carbonate buildups) and changes in sediment dispersal patterns (e.g. rivers flow around diapirs) may occur towards diapir crests; (iv) diapiric highs may become an important sediment source - by inducing resedimentation of sediments deposited on the diapir, or by the reworking of evaporites which reach the surface; (v) faults produced by dissolution collapse, can have large throws and control variations in sediment thickness and sediment dispersal.

(c) Initiation of diapirism. All the studied diapirs are considered to have been initiated by alpine compression; either by thrust induced loading, or by the development of a pillow or piercement structure from a salt cored fold.

(d) Level of intrusion. Two of the studied diapirs are considered t have reached the sediment surface while they were diapirically a tive.

CHAPTER 4

Ryb an Avon - the source of the river (trad. Cornish)

Coleman and Spring (1987)

4.1 OLIGO-MIOCENE FLUVIAL MOLASSE SEDIMENTATION

4 1.1 Introduction, aims and structure of the chapter

Fluvial molasse sediments of Oligo-Miocene age outcrop e ensively within the Ebro b sin (Fig 4.1). At the northern, Pyrenean margin of the ba 'n they unconformably overlie Eocene d posits of the South Pyrenean basin. The e marginal deposits are o great interest as they record how sed' ent was supplied to the m in Ebro basin. Well xp ed seq enc s depos'ted in this m rginal position, in the ar a b twe n N val and Viacamp (west to e t) and Barbastro and Pont Suert (s uth to north) form the s bject of this chapter.

The work represents the first detailed study of these s diments. Nijman and Nio (1975) coined an informal Formation name "the Collegats Formation" for the sequence. The name is r tained here but at the Group level. Nijman and Nio (op. cit.) c nsidered the Group to be "post-tectonic". Seguret (1972) h wever, recorded uplift and the production of localised breccias al ng the entire length of the External Sierras, during the d position of the Collegats Group.

The primary aim of the chapter is to characterise the nature of the Collegats Group: to discuss the factors which controlled the system, how they produced change and how they inter-related.

The goal will be acheived progressively: firstly, by describing the key facies recognised within the Collegats Group; secondly, by discussing established models and facies associations of relevant fluvial environments; thirdly, by outlining the features of key environments specific to the Collegats Group; fourthly, by detailing the key localities; fifthly, by linking the key environments and localities into a synthesis which characterises the fluvial system and outlines the palaeogeographic evolution of the area, and sixthly, by way of summary, the controls on the fluvial system are outlined.

The description of facies and model environments allows a facies code to be erected (c.f. Miall 1978) and the principles that underpin environmental interpretations to be discussed. Both steps save on descriptive repetition. The distillation of six key environments moulds the data into a form by which it can be readily compared with other fluvial sequences, while the description of key localities is a concession to the data set - certain areas being particularly well exposed. It has the advantage of allowing particular parts of the system to be shown in great detail.

4.1.2 Ages of the sequences

At present there are no published palaeontological dates for any part of the study area. The nearest published dates come from Santa Cilia at the northwestern end of the Barbastro anticline, where vertebrate material suggests an Aquitanian (early Early Miocene) age for the Collegats Group (Crusafont *et al.* 1966). The Instituo Geologico y Minero de Espana 1:200,000 maps, sheets 32 and 33 suggest that the Collegats Group in the study area is of the same age. However, the date must be treated with caution as it rests on lithostratigraphical correlation over 26km of ground with variable exposure. Indeed unpublished work within the study area, based on mammal teeth collected near Olvena in the Esera gorge (Van Damms pers. comm 1987) suggests that part of the Collegats Group is uppermost Oligocene in age.

In summary, biostratigraphic studies within the Collegats Group are at an early stage. Throughout this thesis the Group is considered to be Oligo-Miocene in age.

Despite the current problems of biostratigraphy it has been possible to correlate within the study area. Locally exposure is complete. So unconformities and beds can be traced laterally for several kilometres. Correlation between such well exposed areas was achieved by conventional mapping and cross-section construction. In this way the relative ages of different localities were determined, so that the entire study area could be viewed as a coherent system - even though the exact age of the deposits and the precise relations to sequences outside the study area are unknown.

4.1.3. General setting - the fluvial system

The palaeogeographic setting at the start of the deposition of the Collegats Group can be gleaned from a balanced map (Fig.2.28(c)) and an isopach map (Fig.1.10(a)). Key elements to the landscape were the Pyrenean mountain chain to the north, and the Ebro basin to the south. The mountain chain at this stage had a two-fold origin: firstly, remnant relief from the inversion and southwards translation of the Vallcarga flysch basin, and secondly, active development of a triangle zone in Alpine basement {2.8}. The fill of the Ebro basin thickens markedly towards the Pyrenean mountain chain and the basin is considered to have acted largely as a foreland basin to the Pyrenees {1.3}.

The drainage pattern resulting from this palaeogeography can be deduced by applying a model of the "fluvial system" developed by Schumm (1977). Schumm's concept of the fluvial system is a broad one embracing interaction between a wide range of geomorphic variables in time and space (Table 4.1). At any one time the system has three elements in plan form (Fig.4.2): (1) a *source area,* which supplies sediment, corresponding here to the Pyrenean mountain chain (2) a *transfer system,* by which sediment is carried to (3) a *sedimentary basin,* in this case he Ebro basin, in which the sediment is deposited (Schumm op.cit.).

The geological data set is most complete in the sedimentary asin where the rock sequence records the fluvial system at a point in time and through time.

For the geologist, the fluvial system is enigmatic upstream of the sedimentary basin. The key variables measured by geomorphologists (catchment area, relief, climate, vegetation, drainage density, drainage morphology) and changes in these variables through time cannot be studied directly. However, the nature of the transfer system and the drainage basin can be constrained by (i) inference from sedimentary sequences preserved within the basin e.g. given a suitable geological framework, clast types can closely constrain catchment areas, while coarsening C.U. or F.U. megasequences can be used to infer rejuvenation or denudation in source areas (Heward 1978; {4.3.2(c)}) and (ii) by the chance preservation of parts of the transfer system or drainage basin since the divisions between elements of the fluvial system are not inviolate. For example, sections of the transfer zone can be preserved as palaeovalley fills and thus studied directly, with the proviso that valley fill deposits represent sediment which was not transported: the fill stage may differ significantly from the transport stage {4.3.4}.

The study area is located at the margin of the Ebro basin, between the basin and the source area, so that it shows characteristics both of a transfer system and of the proximal part of a large alluvial fan.

4.2 DESCRIPTION OF FACIES

The observed facies are be divided into two groups: (i) primary depositional facies, the deposits of sediment gravity flows and traction carpets, and (ii) pedogenic facies.

Individual facies are characterised by lithofacies, sedimentary structures and sediment body architecture. These features are summarised on Table 4.2 where each facies is assigned a facies code. The table is modified from Miall (1978) who listed facies observed in braided streams {4.3.1(b)}. It has been expanded to cover other depositional environments, particularly facies seen on alluvial fans {4.3.2} and terminal fans {4.3.3}. The table is composite in that it includes both (i) sedimentary facies, and (ii) event deposits (sediment gravity flows) which represent small scale facies sequences.

4.2.1 Sediment gravity flows

4.2.1(a) Classification

Middleton and Hampton (1973; 1976) define: *sediment gravity flow* as sediment transport in which movement parallel to the bed is provided directly by the pull of gravity; and *fluid gravity flow* where the fluid is moved by gravity and it drives the sediment parallel to the bed - essentially as bedload, in a traction carpet.

Lowe (1979) offers a slightly different definition of sediment gravity flows:

> processes driven by gravity acting on the excess weight of entrained solids

Lowe considers that the key feature is that the entrained sediment causes the flow to have excess weight over its surroundings. This concept is easy to grasp in subaqueous settings and has led to a rule of thumb definition of sediment gravity flows: "if you took away the sediment the flow wouldn't move; while with fluid gravity flow if you take away the sediment flow would continue". However, the rule of thumb does not translate well to the subaerial environment where everything has excess weight over air: consider a turbulent subaerial sediment gravity flow if you took away the sediment the water would still flow downhill. To clarify this situation the author would define a *sediment gravity flow as a sediment-fluid dispersion moved by gravity where the grains are supported above the bed by properties of the flow and grain interactions.*

Lowe (op.cit.) divided sediment gravity flows by rheology and grain support mechanisms into two groups and five sub-groups Fig.4.3). His primary division recognises *fluidal flows* which behave as viscous fluids, and *debris flows* which behave plastically.

His secondary divisions closely follow the "end member" support mechanism categories suggested by Middleton and Hampton (op.cit).

Fluidal flows are dominated by: fluid turbulence (turbidity currents); escaping pore fluids (fluidised flow); or partial support from escaping pore fluids (liquefied flow). Of these t ree, fluid turbulence is the most important. The latter two are t ought unlikely to transport gravels on their own because, e cept in pyroclastic flows where air is drawn into the flow by c nvection, the fluid which supports the flow is expelled upwards a d outwards so that the process is not autocyclic. However, es aping pore fluids may well be important at the end of a flow history.

Debris flows are divided into debris flows and grain flows, in which cohesion and dispersive pressure repectively are the dominant grain support mechanisms.

Nemec and Steel (1984) follow Lowe's classification of sediment gravity flows but modify the debris flow category slightly by emphasising a rheological distinction between *cohesive* debris flows (or mudflows) and *cohesionless* debris flows (likely to be dominated by dispersive pressure). In doing this they argue that rheology controls clast support and the mode of en masse deposition, and that it is difficult to imagine a friction-free (and, therefore, dispersive pressure free) flow.

In subaqueous settings flows of sand-sized material moving purely by grain flow are likely to occur only on slopes ranging from 18-28°; furthermore, in the absence of modifying infuences (e.g. a more rapidly moving overlying current, or a high-density interstitial matrix) they will be less than 5cm thick (Lowe 1979). So grain flow per se, like fluidised flow and liquefied flow, is unlikely to be the dominant support mechanism for discrete beds.

For the field geologist, different rheologies and different rain support mechanisms can be expected to produce different bed ontacts and different internal structures. Primarily: debris lows are characterised by the absence of significant erosion at bed bases, reflecting non-turbulent plastic flow in which shear stress reduces to zero at bed bases; and turbulent flows are characterised by erosive bases reflecting shear at the bed base. The effects of the support mechanisms and rheologies which form the basis for Lowe's sub-division of these flow types can also be t bulated: they are discussed below in separate sections on debris flows and fluidal sediment gravity flows (Table 4.3; {4 2.1(b)}; {4.2.1(c)};).

D spite being able to tabulate the expected effects, Lowe's cl sification is difficult to apply. There are three main areas of difficulty (Lowe 1979; Nemec and Ste l 1984):

(i) it is unlikely that an indiv'd al flo will be supported by a single grain support mechan' m. Different mechanisms may operate on different grain populat'ons, or a single clast may be supported by several mechanisms operating serially over the life of a flow. Consequently an exhaustive clas ification based on support mechanisms would be extremely complex and difficult to apply. With a wide range of grain sizes this problem becomes important in attempts to classify a flow as sediment gravity or fluid gravity flow: was the bulk of the material supported by the flow, or was it driven along as bedload?

(ii) Different flow types (sediment/fluid gravity), different support mechanisms, and different rheologies may produce the same structures, so that even the most exhaustive of classifications is difficult to apply.

(iii) The classification is based on end member support mechanisms, so that bouyancy cannot be included, since, although it may be the single most important support mechanism it cannot act as an end member.

Of the theoretical range of sediment gravity flows only debris flows have been recognis d with confidence in the Collegats Gr up. Beds which may be repr sentative of fluidal sediment gravity flows occur but they cann t be d' tinguished from some typ s of fluidal flow. For flows of this type the term sheetflood is suggested. The following sect'ons discuss the general nature, and subdivisions used in this udy, of debris flows and then sheetfloods.

4.2.1(b) Debris flows

Debris flows lack sign'ficant erosion at bed bases. In plan th y are finger-like; in secti n they have sheet geometries, with mud-rich flows being convex upwards, while mud-poor flows have marginal ridges.

However, the diagnostic features are not inviolate. In pa ticular, deposits with the texture of debris flows may overlie er sive bases. Such eros've bases can arise in two ways. Firstly, by the draping of a pre-existing erosion surface, for example a fluvial channel, and, secondly, by the ploughing effect of clasts too large to be fully supported by the flow, that are rolled or caused to slide by the moving flow. Such clasts constitute a "kind of" a traction carpet, and can commonly be recognised as outsized clasts at the base of flows (Lowe 1972).

A number of other features which can be used to confirm a debris flow origin, and infer the flow rheology, may, or may not be present (Table 4.3; Nemec and Steel 1984; Lowe 1979):

(i) Poor sorting reflecting the transport mechanisms of debris flow which, largely, do not sort clast sizes, and the tendancy of such flows towards rapid freezing, preventing sorting at deposition.

(ii) inverse grading at the base of flows, probably the result of the formation of a zone of high shear at the base of a flow with high frictional strength.

(iii) "Preferred clast orientation" often sub-horizontal and parallel to flow, may originate from strongly sheared laminar flow, most likely due to clast interactions and "dispersive pressure" - Nemec and Steel (1984). Such strongly sheared laminar flows are favoured by relatively low viscosity, low shear strength and low cohesion.

(iv) Upflow imbrication of clasts with (a) or (b) axes imbricated. Upflow imbrication may be the result of "grain interaction" but the feature is essentially unexplained (Hiscott and Middleton 1979).

(v) Crude layering interpreted as due to surging flows, perhaps the result of variable matrix or matrix contents along a flow.

The debris flows seen in the Collegats Group are divided into fo r types, solely on textural grounds: clast supported gravel Gd ; matrix supported gravel Gdm; poorly sorted gravel Gdp and sa dy debris flow Sd (Table 4.2(a); Fig.4.4(a)).

4.2.1(c) <u>Subaerial fluidal s d'm nt gravity flows - distinction fr m stream flow deposits</u>

Nemec and Steel (1984) consider that fluidal sediment gravity flows exist in the subaerial environment, and cite the work of Lawson (1982) and Pierson (1981) to support this claim. They go on to make an analogy between these flows and the hyperconcentrated flows of Beverage and Culbertson (1964) which develop at sediment water mixtures of 40-80% sediment by weight, intermediate between stream flows and debris flows. Nemec and Steel do not seek to define flow type by weight % of entrained sediment but use the analogy to introduce the concept of a continuum of flow types - in terms of the sediment gravity flow/fluid gravity flow classification the transition can be thought of in terms of the swop-over point between sediment being dominantly translated as bedload and sediment dominantly held in suspension. Nemec and Steel (op.cit) consider that these intermediate flow types have been described extensively in the literature under a number of different terms "intermediate deposits" (Bull 1963) "streamflood deposits" (Bluck 1967; Steel 1974) and "sheetflood deposits" (Wasson 1977,1979; Heward 1978a; Nemec and Muszyski 1982).

However, because of the continuum in flow types it is unclear what features can be used to recognise the deposit of a fluidal, fully-turbulent, subaerial sediment gravity flow: Nemec and Steel consider that

> ...intermediate flows tend to form sheet-like beds, but these may probabaly range from only slightly channelised and weakly graded to strongly channelised normally graded and even thoroughly stratified.

Although characteristic none of these features, individually or collectively, can distinguish fluidal sediment gravity flows from fluid sediment gravity flows.

Three facies in the Collegats Group can be considered to relate to this problem. They occur in a fluvial setting and are most common in sequences considered to represent a terminal fan environment {4.3.3}. Each facies is characterised by an erosive base, fining upwards motif and a sheet geometry; the facies are subdivided on textural grounds into clast supported gravel Gs, sandstone with outsized clasts at the bed base Sde, and sandstone Sf (Fig.4.4(b)). Pebble clasts are imbricated a(t)b(i); the majority of the overlying sandstone is massive, but it may be replaced upwards by plane lamination, climbing ripples or antidune formsets.

The erosive base to each facies is indicative of turbulent flow. The fining-upwards motif and occasional upwards replacement of plane lamination by climbing ripples indicate waning flow. The plane lamination, the ripples, and the imbricated pebble clasts indicate that the sediment was deposited under the influence of a traction carpet. However, Lowe (1982) argues that in subaqueous turbidites

> Other than the common occurrence of the Ta division [massive sand] below Tb [plane parallel laminae] there is no experimental or theoretical evidence that Ta forms by high

velocity traction sedimentation, and experimental results suggest that it is deposited direct from suspension from high density [sediment gravity] flows.

Therefore, massive sand at the base of these beds may indicate that at least the sand fraction was transported largely by fluidal sediment gravity flow. If this is accepted then application of Lowe's classification (Lowe 1979, 1982) of sediment gravity flows suggests that these three facies are the deposits of turbidity currents. However, given (i) the entrenched subaqueous connotations of the term turbidite and (ii) the difficulty in distinguishing between flows dominated by turbulence and flows dominated by traction carpet transport - for example, were sand grains deposited under the influence of a traction carpet largely transported in suspension ? - the author draws back from using the term turbidity current and hence the term turbidite in the subaerial environment. The author prefers to use the term sheetflood in a non-genetic sense to describe the r nge of facies described above: gravel sheetfloods, Gs; g avel-lag sheetfloods, Sde; and sandy sheetfloods Sf.

4.2.1(d) Statistical analysis of sediment gravity flows

Sediment gravity flow units from a given depositional system commonly show a significant positive correlation between maximum particle size (MPS) and average bed thickness (Bth); by contrast beds deposited "grain by grain" by traction currents show lower correlations (Bluck 1967; Steel 1974; Nemec and Steel 1984).

Nemec and Steel (1984) have developed a model for debris flows which explains these correlations. It is particularly interesting because in addition to explaining this relationship, data sets which are in close accordance with the model can be classified as the product of either *cohesive* or *non-cohesive* debris flows. This is a significant advance on the field-based textural division of debris flows presented above. So the model was applied to the conglomerate sequences in the area.

The model is founded on two equations: the first describes the rheological properties of debris flows, and the second describes the competence of debris flows. By combining and rearranging these equations Nemec and Steel (1984) produced "a conceptual model" which relates maximum clast size to bed thickness.

$$D = F_c + i\ F_i(Y,...) \qquad \text{Cohesive debris flow (a)}$$

and $$D = i\ F_i(Y,...) \qquad \text{Cohesionless debris flow (b)}$$

where: D = maximum clast size

F_c = cohesive strength factor

$F_i(Y)$ = other physically possible support factors considered only in terms of their dependence on the flows thickness (Y).

For simplicity the equations are taken as first order linear equations. Figure 4.5 summarises the model. The equations can be considered in terms of two components: a bed thickness independent component and a bed thickness dependent component. The thickness independent component is represented by the cohesive strength factor, where cohesion is defined as:

> ....the resistance of a sediment against shear along a surface which is under no pressure - Nemec and Steel (1984)

and is thus a material constant. All other supportive factors are related to the normal stress operating on the flow, essentially the flow thickness.

Although Nemec and Steel (1984) state that equation (b) expresses the competence of cohesionless debris flows it, in fact, expresses the competence of all non-cohesive sediment gravity flows. Indeed, in interpreting plots of maximum clast size against bed thickness they use the equation in that manner.

Application of the model

The model is applied by plotting maximum pebble size (MPS) against bed thickness (Bth) for a sequence of beds believed to have been deposited by the same system. MPS is calculated by measuring the maximum diameter of the twelve largest clasts within a bed, ignoring the largest two clasts, and calculating an average for the remaining ten clasts (Nemec *et al.* 1984). The plot has three significant features: (i) the correlation coefficient and its statistical significance; (ii) the intercept of the regression line with the Y-axis, and (iii) the slope of the regression line.

(i) Ideally the value of the correlation coefficient is a measure of consistency in the physical behaviour of flows: it will be high if all flows have the same rheology and support mechanisms, but low if these factors vary between flows. Unfortunately, three ther factors also bear upon the value of the correlation coefficient (Nemec and Steel 1984):

(a) overestimation of bed thickness may result from bed amalgamations not detected in the field;

(b) underestimation of bed thicknesses may result from measuring layering produced by flow surges rather than true bed boundaries;

(c) measured beds may not be sediment gravity flows.

(i) Where the value of the correlation coefficient is high and s atistically significant the data can be considered to conform to the model. So if the regression line makes a positive intercept on the Y-axis the data set can be inferred to have been produced by cohesive debris flows.

The comparison of different intercept values for different data sets can give an idea of "relative cohesion". In this respect Nemec and Steel (1984) recommend that true maximum clast diameter should be used rather than MPS, to give a better estimate of flow competence. However, for almost all the sequences detailed in this study a higher correlation is achieved between Bth and MPS than between Bth and maximum clast diameter. This may be because outsized clasts which could represent the deposits of "traction carpets" are not included in the calculation of MPS.

(iii) The slope of a statistically significant linear correlation relates to the sum effects of support forces other than cohesion. In this way inferences can be made from different data sets within the same system about variations in support mechanisms (for an example see Nemec and Steel, 1984).

Unfortuntely the results produced in this study were not consistent. Although a number of sediment gravity flow dominated sequences show high, positive correlations, in accordance with the model, others show low, statistically insignificant correlations. Furthermore, while the majority of conglomerate sequences deposited by traction currents had low positive correlations some showed high positive correlations at odds with the model. The inconsistencies make it impossible to attach any significance to those data sets which accord with the model. No specific problems can be outlined. Assuming that the model is valid, any of the problems noted above could have operated to bias the data sets away from accordance with the model. In summary, the method is not recommended.

4.2.2 The deposits of traction carpets

The deposits of traction carpets can be divided into two groups: gravel deposits and sand deposits. The sandstone facies and facies codes used in this study closely follow Miall (1978; Table 4.2(c)). Sedimentary structures formed by unidirectional luid flow over sand are well understood (Harms et al. 1982). So no explanation is given of the sandstone facies. By contrast, structures formed in gravel are enigmatic. So the six distinct f cies recognised in the study area are discussed individually below {4.2.2(a)-(f)}.

4 2.2(a) Gm massive or horizontally bedded gravel

D scription

The facies is characterised by clast-supported conglomerate beds, 0.5-2m thick which have sheet geometries (Table 4.2(b); Fig.4.6). Bed bases are erosive and may have a relief of up to 0.5m. Internally the beds are either massive or they show indistinct horizontal bed 'ng. The internal bedding is formed by slight textural chang (gr in ize, degree of sorting) or by th n intercalated sand e beds (Sp). The top part of the beds may fine upwards. M d at to poor imbrication may be developed a(t)b(i).

Interpretation

The markedly erosive b se a d s et geometry are interpreted as reflecting stream flow within a wide or unstable channel. The imbrication and locally poor sort'ng, suggests that the facies may be the result of a combination of processes: a traction carpet associated with a high density turbulent flow. The facies is similar to the gravel sheetflood facies (Gs), but differs in that the base is more irregular, beds are thicker, and F.U. is restricted to the top of the bed.

Similar facies have been recognised in other studies. For example, the fills of the main braid channels of the modern Kicking Horse river in the Canadian Rocky Mountains are closely comparable (Hein and Walker 1977). Bluck(1979) considers massive conglomerates of this type to be formed at high stage when the river is devoid of bar features and bedload moves as a wide featureless sheet. The Gm facies is interpreted similarly as the highest stage deposit of a braided river, so that imbrication from this facies is the best indicator of palaeoflow.

The indistinct horizontal bedding is interpreted as the result of stage fluctuations while intercalated horizontally laminated sandstone beds (Sh) are believed to represent stage fall (Rust 1972, 1975; Boothroyd and Ashley 1975; Hein and Walker 1977; Bluck 1979; Kraus 1984).

4.2.2(b) Gsc Massive gravel scour-fills

Description

The facies is characterised by superposed, clast-supported, concave-up, scour-based gravels. The scour fills are massive and moderately well imbricated a t b(i) (Fig.4.7). The facies differs from the massive gravel facies (Gm) in that the scour fills have a ribbon geometry: up to 20m long in a flow parallel direction 3-5m across and 0.2-1m thick. Individual ribbons are sub parallel to each other and to the high stage palaeoflow determined from the Gm facies. Where the two facies occur together the Gsc facies commonly overlies the Gm facies in a small-scale fining upwards sequence (4.3.1 b [illegible]).

Interpretation

The elongate scour base and ribbon shape of this facies indicates that it is the product by the filling of a minor channel. The channels may either have separated discrete bar features, or have been scour irregularities on the floor of a major channel. On the grounds of the associated fining upwards motif the commonly observed facies transition from Gm to Gsc upwards is interpreted as the product of falling stage (4.3.1 b [illegible]) i.e. the Gsc facies is a lower stage facies than the Gm facies (Fig.4.8). The moderately well developed imbrication indicates that traction currents were dominant in transporting sediment and that the channels were filled longitudinally. Superposition of the channels indicates that they were unstable regularly switching positions probably in response to discharge fluctuations.

4.2.2(c) Gt Trough cross-stratified gravel

Description

The facies is characterised by trough-shaped, erosionally-based gravel beds. It differs from the gravel scour-fill facies (Gsc) in being trough cross-stratified in sets 0.1-0.2m thick. Foresets may fine upwards individually and systematically through a bed (Fig.4.8). Individual troughs are commonly superposed; unfortunately they were nowhere traceable in flow parallel sections. Palaeocurrents from this facies have a low dispersion about the high-stage palaeoflow measured from the massive gravel facies (Gm).

Interpretation

The facies is interpreted similarly to the gravel scour-fill facies (Gsc) as the result of the filling of minor channels. The fining upwards of individual scours indicates a gradual waning of the flow, either during abandonment or during falling stage. The fining upwards cross-beds indicate sorting of the material supplied to the scours, or sorting during fill - possibly the result of the sediment being emplaced by sediment gravity flows.

On the grounds of better sorting the facies is believed to represent a slightly lower stage than the Gsc facies (Fig.4.8).

4 2.2(d) Gp Planar cross-stratified gravel

Description

The facies is characterised by gravel beds with low to high angle (18-27°) planar cross-stratification (Fig.4.9). Beds are 1-1.5m thick, and erosionally based. The facies initiates from inclined erosional surfaces which overlie other traction current facies (Gm; Gsc; Gt; Gf). The cross-stratification is defined either by discrete 20-30cm thick foresets that commonly fine upwards, or by a pronounced pebble fabric. Convex-upwards reactivation surfaces may occur near the tops of beds. The reactivation surfaces are overlain by relatively small-scale cross-beds which rebuild the cross-bed angle. Toe-sets are either straight or tangential in their lowermost parts. Groups of concordant forests have been traced for distances of 23m, 6m and 11m. Cross-beds may either be truncated by the erosional base of an overlying bed or they may die out at the top of the bed and pass up-palaeoflow into a well imbricated conglomerate sheet (Fig.4.9(a)).

Interpretation

Kraus (1984) considers that high-angle, planar cross-stratification in braided river deposits can originate in two ways:

(i) as falling stage modifications to longitudinal bars;

or (ii) as the product of flow transverse migrating bedforms that were stable under flood conditions.

To these possibilities a third can be added:

(iii) as the product of "Gilbert-type" deltas filling abandoned channels or scour pools in the channel floor.

Of these three possibilities the least likely origin of the Gp facies in the Collegats Group is as Gilbert type deltas, which need not prograde in a direction that is either constant or downstream, so that they are not favoured to have produced the observed cross-bedding. The importance of the remaining two possibilities is difficult to assess as they need not be mutually exclusive.

Of these three possibilities the least likely origin of the Gp facies in the Collegats Group is as Gilbert type deltas. These deltas have a lobate shape in plan and need not prograde in a direction that is either constant or downstream, so that they are not favoured to have produced the observed cross-bedding. The importance of the two remaining possibilities is difficult to assess as they need not be mutually exclusive.

A bedform is a repeated topographic pattern which develops under given flow conditions (velocity, viscosity etc.) over a given substrate. Consequently a bedform origin of cross-bedding can be inferred in two ways: either by demonstrating a repeated topographic pattern or by demonstating growth of a single element of that pattern from a plane bed. In the deposits of bedload dominated rivers the former requires exceptional exposure given the scale of bars seen in modern rivers, and the author knows of no study where this approach has been used. Similarly the author knows of no study where growth of planar cross-beds from a gravel bed is conclusively demonstrated. However, Kraus (1984) does consider planar cross-stratification in the Fort Union Formation, in the Big Horn basin Wyoming, to be formed in this way. She supports this interpretation in three ways: firstly, by the great lateral extent of cross-bedding, citing an example of a group of c ncordant foresets 450m long in the flow parallel direction which indicates a stable form produced by steady flow; secondly, by the negative evidence of the absence of any observable lateral t ansition between the Gp facies and other traction current facies, and thirdly by sequences in which the Gp facies is overlain by the massive imbricated gravel sheets. These sheets tr nsitionally overlie the foresets, and are interpreted to be their upstream time equivalents, reflecting a bipartite bar morphology (Bluck 1971, 1976; Steel and Thompson 1983).

B cause the Gp fac'es observed in this study always initiates from erosion surfaces it is interpreted as having been produced by flow changes: as a stage fall when massive gravel facies, Gm passes via an erosion surface into Gp and as channel switching when the gravel scour f'll (Gsc) or trough cross-stratified (Gt) facies pass into Gp. H w ver, th's does not preclude an origin as a bedform as exposure 's i ff' ient to test the possibility that the pattern m b spatially repeated. Indeed the persistence of concorda t f res ts for distances of up to 23m does suggest that the fo e are produced by a form in equililibrium with a steady f , a f ature of bedforms.

The fining upwards f os b ds as in Gt facies, can be interpreted in two ways: e th r as th r sult of sorting on the bar top prior to avalanching, or as the result of sorting during sediment emplacement by tu bul nt ediment gravity flows (Gs). Both mechanisms are plausible 'n the examples studied. The tangential toeset portion of the cross-beds and their erosional base are interpreted as the result of a separation eddy which forms in the lee of the bar face and is attached to the bed producing a scour pit into which the bar migrates.

In summary, the planar cross-bedded gravel facies is considered to initiate by flow change, either falling stage or channel switching and to be maintained by steady flow. During steady flow a pronounced slip face and flat bar top developed. The geometry may be the product of a true bedform, but the exposure is insufficient to demonstrate a repeated pattern which would confirm this interpretation.

4.2.2(e) Gl - Low angle cross-stratified gravels

The facies is characterised by low-angle, cross-stratified, c ast-supported conglomerate beds, on average 1.5m thick. The cross-beds dip at a low angle downstream, around 10° and are imbricated internally. Critically imbricated clasts indicate flow down the dip of the surfaces. The beds themselves may be bounded by erosional surfaces which dip downstream at still lower angles of 0-4° (Fig.4.10(a)). Recognition of such surfaces requires excellent exposure.

Interpretation

The major bounding surfaces are interpreted to represent erosional high stage flood events that partially plane off the relief of the downstream margins of longitudinal bars. The origin of the overlying low angle cross-beds is unclear: they may have developed either from an upstream erosional surface of the same angle, or as a result of initial aggradation to the major erosion surfaces (Fig.4.10(b)). The imbrication is critical as it precludes an origin as large-scale cross-bedding.

4.2.2(f) Gf-Coarse scour plugs

The facies is characterised by matrix free gravels which overlie steeply incised minor scours (Fig.4.9). The scours are 0.5-0.8m wide, 0.3-0.5m deep and filled with a concentration of the largest clasts. The scour fills are typically capped by surfaces intepreted as the result of high stage flow. The scours are orientated at a high angle (40-60°) to the high stage palaeoflow.

Interpretation

The large clasts and the lack of matrix suggests that the scours are the product of high flow energies. But, obliquity with the high-stage palaeoflow suggests that the scours formed at relatively low stage when flows responded strongly to bar topography. The erosion surfaces which cap the scour-fills suggest that the fills immediately predate the highest flow stages. They are tentatively interpreted as having been formed at low stage when bar topography was marked and filled during rising stage when clast reorganisation at the bar head liberated the coarsest material.

4.2.3. Palaeosol facies

4.2.3(a) Introduction

Palaeosols are fossil soils. Modern soils are believed to be a product of five variables - parent material; climate; relief; organic matter and time - with distinct profiles developing as a result of a different combination of these. By analogy with modern soils these factors can be inferred for palaeosols - the present is the key to the past. The importance for the fluvial sedimentologist is that these same factors control the main variables of the fluvial system. The potential for palaeosol studies to give an insight into these variables was first shown in seminal papers by Buurman (1980) and Bown and Kraus (1981) but it has only recently been fully realised: the work of Retallack (1983, 1986) being quite exceptional in showing the breadth of indices that can be produced from palaeosol studies to indicate changes in factors which control the fluvial system. Although lack of time and resources has forced a less intensive study than hat of Retallack (op. cit.) key field observations do allow some of these factors to be outlined.

Before discussing and interpreting the observational soil facies developed for the studied sequences some of the key features and problems of palaeosols are discussed.

4.2.3 (b) Key features of palaeosols in the fluvial environment

A number of features can be used to recognise and classify fossil soils, in particular: colour and colour distribution; soil structures; authigenic minerals; clay mineralogy and micromorphological features.

Of these, colour is perhaps the most important. Many palaeosol sequences are characterised by a sub-horizontal, metre-scale colour banding which shows a frequent disregard to grain size and depositional units (indicating that it developed post deposition). An understanding of what these different colours represent in terms of soil chemistry is a fundamental building block in palaeosol analysis (Table 4.4)

In detail, the majority of the tabular units are not uniformly coloured - colour mottling is common. Such mottling reflects variations in the oxidation states of iron and manganese during pedogenesis. Buurman (1980) recognises two distinct types of colour mottling, reflecting two distinct types of *hydromorphic soils*: gley soils and pseudogley soils (Fig.4.11). In both, the uppermost horizon is commonly reddened and not mottled, reflecting well drained oxidising conditions; although in gley soils the highest groundwater level may reach the surface preserving organic matter, so that the topmost part of the profile is blacker.

Gley soils have their subsoils permanently saturated, and their topsoils periodically saturated dependent upon seasonal fluctuations. So a grey reduced horizon exists below the lowest groundwater table (CG horizon). Above this horizon alternating conditions of oxidation and reduction cause segregation of Fe and Mn compounds into discrete mottles (Cg horizion). The mottles may pass sharply or gradationally into the matrix which is grey or yellowish in colour. Critically, oxidised Fe compounds may penetrate into the CG horizon along cracks and root conduits.

Pseudogley soils are formed when seasonal soil water stagnation affects soil development. Early downward movement of clays (eluviation) creates an impervious layer, an illuvial horizon, where the clays accumulate. Above this horizon Fe and Mn compounds are segregated in a fashion comparable to that seen in gley soils. However, below the saturated horizon reduced Fe and Mn ions are transported down through cracks etc. where they accumulate. This situation "pseudogley mottling" is in direct contrast to the situation in the saturated horizon in pseudogley and gley soils, where oxidised compounds accumulate in conduits "gley mottling". On this point gley and pseudogley soils may be distinguished.

Small-scale soil formed structures are discussed at length by Brewer (1964) and USDA (1975). Root formed structures are particularly useful supportive evidence of pedogenesis (Klappa 1980) as are clay cutans (Brewer 1964; Buurman 1980) and burrows, when considered carefully (Brewer 1964; Seilacher 1978; Buurman 1980; Bown and Kraus 1981; Atkinson 1983).

The study of authigenic minerals formed during pedogenesis has a longer history than the interpretation of colour mottling. Calcretes have been described extensively, and examples of gypcretes, silcretes and ferricretes are also well documented (for reviews see Goudie and Pye (1983) and Esteban and Klappa (1983)). The author recognises five types of calcrete within the Collegats Group. Their nature and origin are discussed below (4.2.3(f)).

Clay minerals formed within the soil profile are an important part of soil studies. Like colour mottling they can be used to infer the palaeosol chemistry and so aid classification and interpretation (Weaver and Beck 1977; Watts 1980; Atkinson 1983; Goudie 1983).

Micromorphological studies can prove the action of *pedogenic* processes (Brewer 1964; Fitzpatrick 1984). But such studies add relatively little to the interpretation of soil sequences when the goal is controls on fluvial sedimentation.

Due to time clay mineral studies and exhaustive micromorphological studies were not undertaken. Instead emphasis was lain on careful field observations of palaeosol colours and colour distributions, and on the presence and form of authigenic minerals.

4.2.3(c) Problems with the study of Palaeosols

There are many difficulties in the study of palaeosols. Laying aside our incomplete knowledge (Wright 1986) there are two key problems: firstly, the difficulty in recognising distinct soil profiles in the ancient, and secondly, that in detail it is difficult to compare ancient and modern soils.

Because units of sedimentation are usually thinner than complete soil profiles (around 2m) as any individual sediment layer is buried it will pass gradually through the entire soil profile and suffer the pedogenic effects of all the different horizons within the soil. The process is known as compound pedogenesis (Bowen and Kraus 1981) and, as a result, distinct soil profiles tend to be rare. However, some studies do describe thick sequences of discrete palaeosols (e.g. Retallack 1983; 1986) and the problem can be circumvented by recording observational soil facies which can be interpreted as distinct soil types (Atkinson 1983).

The difficulty in comparing palaeosols with modern soil classifications is that the latter are commonly based on features that are either not measurable in the ancient (for example moisture content through the year) or require laborious analytical work (percentage clay, $CaCO_3$, or organic matter through the profile (USDA 1975)). However, the wide range of features upon which modern soil classifications are based does allow preliminary classification.

4.2.3(d) Soil facies recognised in the Collegats Group

Six observational soil facies have been recognised in the Collegats Group. Five are based on palaeosol colour and colour distribution, the sixth is a calcrete facies which is subdivided into five further facies by calcrete distribution, form, and abundance. The observational facies are interpreted in terms of five distinct soil types based on the United States Department of Agriculture Soil Survey classification (1975).

4.2.3(d)(i)S_1 Hydromorphic pseudogley mottling

Description

The facies is characterised by colour mottling: mottles are grey at the centre and may or may not have a haloe which is redder than the matrix (Fig.4.12(a)). The range of observed colours is recorded on Table 4.5.

Associated features which may or may not be present are : the preservation of original bedding; minor $CaCO_3$ accumulations (sparry calcite filled burrows) and an increased reddening towards the top of the profiles.

Interpretation

The mottles are interpreted as being the result of reduction along cracks within an essentially oxidised soil profile - a situation directly comparable with the lower parts of pseudogley soils.

However, reddening towards the top of some of these horizons is in contrast with a model of pseudogley soils (Buurman 1980; 4.2.3(b)) which suggests a more complex upwards transition

(Fig.4.11). Where reddening occurs it is suggested that reduced cracks represent the downward percolation of pluvial water which accumulated at the sediment surface and that reddening was imposed on the uppermost layer of the soil profile during dry periods.

The preservation of initial bedding, and the absence of calcrete accumulations suggests that these palaeosols are relatively immature. The sparry patches are interpreted as later calcite cements.

4.2.3(d)(ii) S_2 Hydromorphic gley mottling, brown background

Description

The facies is characterised by colour mottling: mottles have iron accumulations at their centres which are redder than the matrix and may or may not have a bleached haloe. The mottles are typically elongate vertically (Fig.4.12(b)). The background ranges from moderate pink to dark yellowish orange in colour. The range of colours is recorded on Table 4.5. In vertical sequences the facies is commonly associated with soil facies S^1.

Accessory features are as in soil facies S_1.

Interpretation

The colour mottles are interpreted as the result of oxidation along roots etc. within a relatively reduced soil horizon. The mottles are directly comparable to those described in gley soils (Buurman 1980; Atkinson 1983). However, the matrix colour of soil facies S_2 differs from that of true gley soils, which are grey below the permanent water table (CG) and grey or yellowish between that point and the highest water table (Cg) (Fig.4.11).

The facies is interpreted as representing the top part of pseudogley soils where the matrix is only periodically saturated, so that an even reduced colouration is not developed. The mottles develop during periodic stagnation by penetration of oxygen down cracks and root systems. The intimate association with facies 1 and the occasional upwards reddening seen in this facies support this interpretation (Fig.4.12(b)(ii)&(iii)).

Together soil facies S_1 and S_2 can be interpreted as evidence for the development of pseudogley soils. In terms of modern soil classifications they represent examples of entisols: soils which lack well developed horizons (USDA 1975). More specifically they can be placed in the fluvent suborder of entisols, since they were formed in a fluvial setting. They imply a seasonal climate with periodic inundation of the floodplain by pluvial water.

4.2.3(d)(iii)S_3 Hydromorphic gley mottling, grey background

Description

The facies is characterised by colour mottling: mottles are red to yellow-brown (pale grey haloes occur but are rare) and contrast with the matrix which is even grey in colour (Fig.4.13). The range of observed colours is recorded on Table 4.5.

Associated features differ from soil facies S_1 and S_2: organic matter may be scattered through the profile in the form of lignite; strongly oxidised iron accumulations occur within discrete burrows commonly associated with sulphur; thin-shelled bivalves and an increased reddening towards the top may be present in some profiles. In vertical sequences this soil facies tends to be developed in the finest grain sizes, and to be separated from associated conglomerate and sandstone facies by immature and mature Entisols (S_1, S_2) and Ustalfs {4.2.3(e)}.

Interpretation

The colour mottles are interpreted as the result of oxidation along cracks, roots burrows etc. within a reduced soil. The mottles and the matrix colouration are directly comparable to the gley soils of Buurman (1980) and Atkinson (1983).

The oxidised Fe-burrows with associated sulphur are interpreted as the product of the recent weathering of iron pyrite (FeS_2). Pyrite cannot accumulate in a periodically oxidising environment (Buurman 1980). So in contrast with the mottles these burrows record a permanently reduced soil horizon, compatible with the accumulation of organic matter and the bivalves which suggest permanent lakes.

Atkinson (1983) has described similar palaeosols from the Escanilla formation. By analogy with modern "acid suphate" or "cat gley" soils, which show the presence of large amounts of pyrite and occasional straw coloured mottles, he argues that they may have been formed in a coastal (lake-shore?) setting where brackish groundwater conditions prevailed. The bivalves, lacustrine limestones and evaporites within the study area {4.4.7(d); 4.4.3(e); 4.4.4(b)(ii)} indicate the presence of freshwater and saline lakes, allowing a similar interpretation to be offered here.

The facies lacks any distinct horizons and must, therefore, be interpreted as a second type of entisol (USDA 1975). The inferred lake margin setting suggests it can be assigned to the Aquent suborder: examples where pyrite is believed to have occurred can be assigned to a further sub-group - the great group "Sulfaquent".

The persistent vertical facies sequence of channel conglomerate and sandstone - fluvent Entisol - Ustalfs - Sulfaquent, is interpreted as a catena, with the sulfaquents developing away from elevated channel margin levees where Entisols and Alfisols are favoured {4.4.7(d)}.

S_4 Hydromorphic mottling

Description

This facies is characterised by colour mottling in which no distinct pattern could be discerned. The range of colours is shown on Table 4.5.

Associated features are as in soil facies S_1 and S_2.

Interpretation

The colour mottling is interpreted as the result of fluctuations in the soil water content, i.e. hydromorphic in origin (Buurman 1980). The lack of systematic mottling could reflect: compound pedogenesis; variable pedogenic influences through time during soil formation; or lack of time for a systematic colour distribution to develop.

Like soil facies S_1 and S_2 the facies can be classified as a relatively immature Entisol (USDA 1975) of the fluvent suborder. Again a seasonal climate is implied.

4.2.3(d)(v) <u>S_5 Even brown colouration</u>

<u>Description</u>

The facies is characterised by an even red-brown colouration and the absence of colour mottling. With the exception of rare green-grey reduction spots. The range of colours is shown on Table 4.5.

Associated features include: the preservation of original bedding; rain prints; burrows and planispiral gastropods.

<u>Interpretation</u>

The even brown colouration is interpreted as the product of a well drained, oxidised soil.

4.2.3(e) <u>Authigenic minerals</u>

Two important authigenic minerals occur in the studied sequences: gypsum and calcite. Gypsum occurs as a precipitate from gypsiferous groundwaters in the Collegats Group {4.4.6(b)(ii)&(iii)} and as a soil-formed mineral in the Escanilla Formation. The origin of soil-formed gypsum (gypcretes) within the Escanilla Formation has been discussed by Atkinson (1983) and is not discussed here. Authigenic soil formed calcite, i.e. calcrete, is widespread in the Collegats Group.

The most important factor in the formation of modern calcretes is climate. They are favoured by an annual rainfall of 400-600mm and a markedly seasonal climate (Goudie 1983). Seasonal rainy and dry periods mobilize and then precipitate carbonate. In many modern soils there is a close relationship between rainfall and depth to the $CaCO_3$ horizon from the surface (Jenny 1941) implying that calcium ions are supplied from the surface and taken to depth by percolating pluvial water which evaporates and precipitates carbonate during dry periods - the "per descendum" model of Goudie (1973). This is the favoured model for many modern calcretes with calcium being supplied to the soil profile as dust or in pluvial water. Where the host materials are rich in carbonate, as they are in the sediment discussed here, the host material can be expected to be an important supplier of calcium ions greatly accelerating the rates of calcrete formation (Goudie 1983; Atkinson 1983, 1986).

The calcretes studied within the Collegats Group are strongly associated with hydromorphic colour mottlings. A wide range of mottle colours has been recorded (Table 4.5.) the vast majority of which are compatible with soil facies S_2 or S_4. These soil facies, the pseudogley and hydromorphic "soil types", are interpreted above as being relatively immature soils produced by seasonal wetting and drying. It is suggested that the calcretes represent more mature examples of the same soil types. However, these more mature soils cannot be classified as Entisols, soils which lack distinct horizons (USDA 1975) since the carbonate commonly accumulates in distinct horizons. They are better interpreted as Alfisols, soils which: have an ochric epipedon (essentially a pale surface layer - the result of a low organic matter content); have undergone clay translocation to orm an illuvial horizon (as required by pseudogley soils) and ere formed in a markedly seasonal climate. More specifically hey can be assigned to the suborder Ustalf which is aracterised be $CaCO_3$ acumulations. If this interpretation i correct it has precise implications for palaeoclimate, since m dern Ustalfs

> ... form belts between the Aridisols of warm arid regions and the Ultisols, Oxisols and Inceptisols of warm humid regions.
>
> ... trees are either deciduous or xerophytic. Many of these soils have ... a savana vegetation. - USDA (1975).

The actual palaeoclimate can only be assessed by careful consideration of a large number of soil profiles, in particular the relative abundance of different soil types (Retallack 1986) and the effects of factors such as variable subsidence rates and drainage patterns (Atkinson 1986; Retallack 1986). These problems are addressed in a number of thick palaeosol sequences below.

4.2.3(f) Calcrete Facies

Five calcrete facies have been recognised in the Collegats Group C_{1-5}. They are considered as steps on an evolutionary pathway, so that as a calcrete profile develops through continued carbonate precipitation one facies evolves into the next (Allen 1974; Esteban and Klappa 1983). In addition, in well developed calcrete profiles the amount of $CaCO_3$ commonly increases upwards so that a similar sequence of facies may be seen within a single profile e.g. Figure 4.15(b). The divisions are based on field observations within the Collegats Group, but they compare closely with divisions of calcretes produced by other workers (Esteban and Klappa 1983; Goudie 1983).

Although neither pedogenic nor secondary in origin, lacustrine limestones are described here because of their close association with well developed calcretes {4.2.3(g)}.

4.2.3(f)(i) <u>C_1 - Calcified soils</u>

The facies is characterised by powdery calcite disseminated through the plaeosol, together with concentrations at roots and other conduits (Fig.4.15(a)). No ancient examples of this facies were clearly recognised. Some Collegats Group palaeosols are characterised by burrows and root cavities filled with calcite spar. At present it is unclear whether the spar has a soil or a diagenetic origin. Micromorphological studies are required to discriminate between these two hypotheses.

4.2.3(f)(ii) <u>C_2 - Nodular Calcretes</u>

Nodular calcretes composed of rounded "glaebules" are by far the most common type of calcrete accumulation in the Collegats Group. Glaebules are hard, disorthic (easily removed from the matrix (Wright 1982)) and tend to be either 1-5mm or >1cm in diameter. Usually they are elongate, 2-3 times longer than wide, with the long axis vertical. The elongate nodules may be scattered, or ordered to define a vertical fabric, perhaps coalescing upwards to form columnar calcretes (Fig.4.15(b)).

The vertical elongate nature of the glabules, and their arrangement in ordered vertical fabrics, is believed to reflect the initial control of descending $CaCO_3$ rich pluvial waters by vertical roots and burrows which are the most likely conduits for pluvial water. In addition, there is some evidence that roots, or at least microrganisms associated with roots, may be the initial trigger to calcite precipitation (Klappa 1978, 1980; Esteban and Klappa 1983). For these reasons the scattered glabules observed in some profiles may reflect a more dense ramifying network of roots than the vertically ordered glabules which probably reflect tap root systems.

A wide range of nodule colours has been recorded. Variations in colour between the nodule core and rim are common but no consistent trends have been observed Table 4.5. Where vertical nodule fabrics are developed, passageways between nodules are characterised by the mottling of soil facies S_1 and S_2, interpreted as being produced by pseudogley soils.

4.2.3(f)(iii) C_3 - Platy Calcrete

Platy calcretes are comprised of horizontal platy wavy carbonate in beds 2-5cm thick. The beds are separated by less well cemented red silts, and linked by irregular carbonate pillars 2-3cm in diameter (Fig.4.15(b)).

Platy calcrete is believed to develop when the continued recipitation of carbonate causes the direction of easiest water movement to change from vertical to horizontal. Horizontal root ystems are then favoured and the corresponding rhizoliths (root associated mineral accumulations -Klappa 1980) form the bulk of the platy horizon (Esteban and Klappa 1983).

4.2.3(f)(iv) C_4 - Hard-Pan

Hard-pans are well indurated continuous layers of calcrete which lack visible porosity. The term is synonymous with "petrocalcic horizon" (USDA 1975). Non-tectonic brecciation, attributable to root action and karstic dissolution is commonly observed (Fig.4.15(c)).

4.2.3(f)(v) C_5 - Boulder calcrete

Boulder calcrete varies from discrete to coalesced boulders and down to small fragments of hard carbonate. The individual boulders show evidence of a pedogenic origin and are interpreted as the product of the breakup of petrocalcic layers by solution and root action Goudie (1983).

Of the five calcrete facies, nodular calcretes (C_2) are by far the most abundant. Layer calcretes i.e. platy calcrete (C_3) hard pans (C_4) and boulder calcretes (C_5) are relatively rare but they do ocur. Layer calcretes are commonly grouped together, forming part of a trend towards increased or decreased amounts of calcrete. Such trends are difficult to interpret: they could represent either slow sedimentation rates, or drier climates. No studies relating to the lateral extent of calcrete horizons were carried out.

4.2.3(g) Lacustrine Limestones

Lacustrine limestones within the Collegats Group were recognised in the field as thin sheets of cream-coloured, micritic limestone which lack evidence for a pedogenic origin. They occur only in the fine grained deposits of the main fluvial system. Beds range from 10-60 cm in thickness and invariably transitionaly overlie a zone of more or less coalesced glaebules (Fig.4.15(d)). Pedogenic features may be superposed upon the lacustrine limestones, producing facies comparable to hardpans (C_4) and boulder calcretes (C_5), or resulting in a gradual upward transition into nodular calcretes. A thin section and slabs of this facies show a well developed blocky texture of cream-white fragments in a grey matrix.

The sheet geometries are interpreted as representing an interchannel setting: the geometry contrasts with some of the lacustrine limestones in the Escanilla Formation which were deposited in abandoned channels (Nickel 1982). The nodular calcretes which over- and underly the micritic limestones are interprted as "palustrine" or lake margin deposits. The blocky texture observed in thin section and slabs is directly comparable to the crumbly texture seen in lacustrine limestones by Freytet (1973). Freytet (op.cit) interprets this texture as the product of periodic dessication of the lake and cracking of the lake floor - his interpretation is followed here. The lacustrine limestones are much thinner than those described elsewhere (1-2.5m Nickel (1982); 10's m Freytet 1973) and than those described from approximately time equivalent systems towards the middle of the Ebro Basin (10's m Cabrera and Saez 1987). The reduced thickness can be interpreted in two ways: firstly, as the result of a relatively unstable fluvial system, so that long lived lakes were not favoured in any one position and, secondly, as a consequence of the studied outcrops being at the limits of lacustrine limestone development.

4.2.3(h) Palaeosols - conclusions

In summary, five distinct soil types are recognised within the Collegats Group. Three of these soil types are considered to form part of evolutionary sequence: with time immature fluvent entisols (S_4) develop firstly, into mature fluvent entisols (S_1; S_2) and then into Ustalfs as calcrete accumulates.
Each of these soil types indicate a seasonal climate, with the Ustalfs indicating, a climate between warm-humid and warm-arid, and a savanna-like vegetation. The range of carbonate accumulations is considered to reflect time and climate - the greater the percentage of $CaCO_3$ the longer the palaeosol was developing and/or the drier the climate. The fourth soil type, the sulfaquent entisol is considered to be a lake margin or "palustrine" soil, and to have developed as part of a catena away from elevated levee deposits where fluvent Entisol and Ustalf palaeosols developed. The fifth palaeosol type is an evenly oxidised palaeosol. It commonly occurs above coarse conglomerates and adjacent to actively developing highs, and is considered to reflect good drainage either due to high porosity, or to slopes produced by tectonic uplift; as such it could be considered as a third element in a Collegats Group catena.

The warm seasonal climate implied by the palaeosols is supported by palaeoclimatic computer models (Scotese and Summerhayes 1986). These models suggest that Miocene weather systems and climate were similar to those of the present day. Three key features are apparent: (i) a marked seasonal climate; (ii) precipitation dominated by the passage of anticyclones, mid-latitude storms, which occur mainly in the winter and may last 2-3 days, and (iii) there is a marked climatic gradient, especially in the summer months from the colder Pyrenees to the hot, dry Ebro basin.

4.3 ALLUVIAL ENVIRONMENTS WITHIN THE COLLEGATS GROUP

Five key environments are recognised within the Collegats Group: braided river; small alluvial fan; large alluvial fan; terminal fan, and palaeovalley. A two-part discussion of each of these environments is presented in this section. The first part considers examples and models in the literature, so that criteria can be set up by which the distinct environments can be recognised. The second part addresses examples within the Collegats Group to allow characterisation of environments within the fluvial system and comparison with other studies.

In the following section, 4.4, the key localities from which these environments have been recognised are described in detail.

4.3.1 The river environment

(a) The classification of river type by percentage bedload

In a seminal review study of the fluvial system Schumm (1977) concluded that sediment load was the most important control on channel patterns, governing channel width, pattern, and sinuosity as well as multiple channel patterns. Accordingly he classified river channels into three types: bedload; "mixed-load" i.e. bedload and suspended load; and suspended load channels. Galloway (1981) applied this concept to ancient fluvial channels, and divided them into the same three categories by the relative amounts of in-channel sand-pebble grade material (considered to be bedload) and silt-mud grade material (considered to be suspended load; Fig.4.16).

Galloway (op.cit) considers that bedload channel deposits are characterised by in-channel deposits dominantly of sand or coarser grade. He also recognises a number of secondary characteristics: coarse sediment bodies tend to have multilateral sheet geometries, and to volumetrically exceed overbank silt and mudstone facies; channel-fill sequences F.U. slightly in an irregular fashion and pass with marked reduction in grain size into overbank facies; bed accretion dominates channel fill; and channels in plan are straight to slightly sinuous, producing broad continuous sand isoliths.

Mixed-load channel deposits are characterised by in-channel

Mixed-load channel deposits are characterised by in-channel deposits which are a mixture of sand or coarser material, and silt and mud. Secondary characteristics include: coarse sediment bodies which tend to be multistorey and generally subordinate to surrounding overbank deposits; channel-fill sequences which F.U. systematically; evidence for both bed and bank accretion; and a sinuous channel trend in plan producing complex, typically beaded, sand isoliths.

Suspended-load channel deposits are characterised by in-channel deposits dominated by silt or mud. Secondary characteristics include: multistorey, coarse sediment bodies which are markedly subordinate to overbank deposits; channel fills which may be of constant grain size or fine upwards systematically; the dominance of bank accretion in channel fills; and channels in plan which are highly sinuous, or anastamosing producing shoestring or pod sand isoliths.

In addition to using this classification to describe individual channel deposits, Galloway (op.cit.) advocated the extension of the concept to the classification of entire palaeofluvial systems. He did so because he found that type models (e.g. braided, fine-grained meandering) were unable to describe such systems, for three reasons: firstly, because the variety of channel-fill deposits which he encountered was not encompassed by these type models; secondly, because of the variety of intergradational sequences which he recognised within one fluvial system; and thirdly, because of the exhaustive data set needed to justify the application of many standard models e.g. a meandering system can only be recognised with confidence in plan, requiring exposure not available in many areas.

The classification could be difficult to apply when discrete channel geometries are not clear and estimates of the percentage of in-channel material must be made from vertical sequences only: channel fill sequences may pass imperceptibly upwards into overbank facies, or subsidence rates may be so low that only the lower parts of channel fills are preserved - in either case the true nature of the channel fill may be difficult to distinguish.

Galloway (op.cit) recommends that "type models" are applied locally to areas of good exposure as a second stage in the classification of palaeochannels. The author is in broad accordance with this approach, and considers that the Collegats Group palaeofluvial system is dominated by bedload rivers, many but not all of which can be shown to have been braided.

4.3.1(b) Braided rivers

(i) Diagnostic facies and facies associations of braided river sediments in the Collegats Group

> Braided channels are single channel bed-load rivers that at low water have islands of sediment [bars] or relatively permanent vegetated islands exposed in the channels (Doeglas 1962) - Schumm (1977)

Therefore, to apply the term "braided", emergent braid bars must be documented. Three facies within bedload deposits of the Collegats Group indicate significant bar topography and can be used to infer a braided pattern: oblique gravel scours (Gf); low angle planar cross-stratified gravels (Gl) and planar cross-stratified gravel (Gp). Since these facies could also develop on lateral bars (bars attached to the channel bank) the further constraint of the absence of slough channels should also be met. Slough channels occur on the inner margins of lateral bars and are characterised by relatively low stage facies: sandy megaripples (St); chute bars; ripple (Sr); small Gilbert type deltas, and spits (Bluck 1979; Fig.4.8).

Bedload conglomerates of this type are common in the Collegats Group, and they are considered to be the product of braided streams. However, the dominant conglomerate facies of such sequences are massive gravels (Gm) and massive gravel filled scours (Gsc), with other traction current facies (Gf, Gl, Gp) and trough cross-bedded facies (Gt) being relatively rare. Only one example of a debris flow was seen in such sequences. In addition to gravel facies a number of sandstone and siltstone facies are recorded: planar and trough cross-bedded sandstone (Sp, St); rpples (Sr); pebbly sand scour fills (Ss); silt and mud (f). However, these finer grained facies are volumetrically small. (The significance of the relative proportions of the different facies is discussed below {4.3.1(b)(ii)}).

The author does not hold with all the inferences which can be drawn from the work of Miall (1978) on model braided river facies associations. However, for reference, the complete Collegats Group bedload river facies association is comparable with Miall's Scott and Donjek types (Fig.4.17(a)).

4.3.1(b)(ii) Small-scale facies sequences in braided stream deposits from the Collegats Group

Two types of 1-3m thick, small-scale facies sequence are recognised within braided stream deposits of the Collegats Group: F.U. sequences, which form the vast majority of the deposits, and rare C.U. sequences. Larger scale sequences are also recognised. These larger scale sequences and the concept of sequence scale are discussed elsewhere {4.3.2(c)}

Small-scale fining upwards sequences

Description

The sequences are erosively based and commence with 1-3m of conglomerate, which may fine upwards gradually, and is usually composed of a single gravel traction current facies (Gm; Gsc; Gp; Gl and Gt). The gravels may pass transitonally into 10-20cm of sandstone, characterised by traction current sandstone facies (Sh; St; Ss) at the top of the F.U sequence. Variants on this model sequence are characterised by more than one gravel facies. Such sequences usually commence with massive gravel facies (Gm) and pass upwards into other gravel traction current facies (Gsc; Gp; Gl; Gt). More complex transitions do occur, e.g. Gm-Gp-Gm-Sl, but sequences of this type tend to be thicker indicating a composite origin.

Interpretation

Because of their erosive bases and appropriate scale, the small-scale fining upwards sequences are interpreted as channel fills. Two features indicate that the channels were filled during waning flow: firstly, the fining upwards motif, and secondly, the observed facies changes - this is clear generally in the upwards change from conglomerate to sandstone facies, and where seen, by the change from the highest stage gravel deposits, massive gravel (Gm) to lower stage gravel traction current deposits (Gsc; Gp; Gl; Gt).

However, the origin of these sequences, in particular the time scale over which these stage fluctuations occurred, is difficult to assess. Do they represent a single flood, and therefore relate to a single storm, or rainy season, or do they represent the longer term abandoment of a channel tract? The problem is a common one in braided stream deposits (Nemec and Steel 1984). Two related features bear upon the (largely speculative) interpretation of the time scale: catchment area and climate. Large catchment areas may extend over areas greater than individual storms. So the effect of a storm is reduced as water derived from it is fed to progressively larger trunk streams, and rivers are characterised by relatively steady discharge. By contrast storms over small catchment areas are likely to produce peaked river hydrographs and deposits characterised by waning flow.

In the Collegats Group palaeosols suggest that the climate was seasonal {4.2.3(h)}. Consequently, seasonal river discharge fluctuations are likely to have occurred. However, Scotese and Summerhayes (1986) consider the weather patterns were similar to those of the present day which are dominated by mid-latitude storms lasting one to three days and covering areas greater than the drainage basin. So the area may also have been characterised by shorter term flow fluctuations.

4.3.1(b)(ii) cont. C.U. sequences

Description

The sequences are erosively based and are either composed entirely of gravel filled scour facies (Gsc) or are based by this facies and pass upwards into massive gravels (Gm).

Interpretation

Because of their erosive bases and appropriate scale the small-sacle C.U. sequences are interpreted as channel fills. Two features indicate that these channels were filled by a flow that increased in strength: firstly, the C.U. motif is characteristic of increasing flow competence, and secondly, the facies change (Gsc-Gm) is characteristic of increasing flow power.

The origin of these sequences is difficult to assess. Three possibilities are apparent: migration of longitudinal bars; increase in stage; and re-establishment of flow in adandoned channel segments.

Bluck (1980) suggests that the downstream migration of longitudinal bars produces small-scale C.U. cycles as coarse bar head deposits pass over finer bar-tail deposits. However, in contrast with the C.U. sequences seen in the Collegats Group, the C.U. sequences which Bluck describes are transitionally based and ideally, overlie F.U. sequences interpreted as channel fills

A simple increase in river stage is thought unlikely produce small-scale C.U. sequences. The volume of entrained sediment would have to be unusually high for it to do so, as the normal response to stage increase is the entrainment of sediment and consequent erosion rather than deposition.

The gradual re-establishment of flow in an abandoned channel segment is the preferred interpretation for small-scale C.U. sequences in the Collegats Group (Costello and Walker 1972). Intuitively such cycles may be expected to commence with low stage facies such as sandstones or siltsones. However, they could commence with gravel scours if only abandoned for a short period, or if flow was only partially diverted *before being* re-established.

4.3.1(b)(ii) cont.Discussion and summary

In summary, the small-scale facies sequences are dominated by F.U. motifs thought to represent channel fills deposited during waning flow. The temporal scale of the sequences is thought to be either seasonal, or that of a mid latitude storm. C.U. sequences are rare. They are considered to be the product of re-establishment of flow in abandoned channels. The rarity of preserved F.U.-C.U "bar sequences" (c.f. Bluck) is despite the evidence for braid bars in terms of the oblique gravel scour facies (Gf), the low angle cross bedded facies (Gl) and the planar cross-bedded gravel facies (Gp). However, the two observations can be reconciled by noting the relative rarity of the braid-bar characteristic facies, and by invoking rapid stage fall with a short period of discrete braid-bar formation so that bar migration was rare (Eynon and Walker 1977).

The Collegats Group small-scale facies sequences are intermediate in form between the Scott and Donjek type vertical profile models of Miall (1978; Fig.4.17(b)).

4.3.2 Alluvial Fans

(a) Introduction

An alluvial fan is a cone shaped body of detrital sediments derived, and radiating out, from a point source where a stream issues from a mountain front (Blissenbach 1954; Bull 1972; Heward 1978b). Deposition occurs because streams increase in width and decrease in depth, and therefore become more inefficient, as they emerge from the mountain front (Bull 1964b).

The alluvial fan is the most important environment within the Collegats Group. Two scales of fan are recognised encompassing the majority of the deposits: a single large alluvial fan, the Huesca fan, with a radius of over 60km (Hirst and Nichols 1986) and numerous small fans with areas of 0.5–2km². The nature and origin of both of these fan types is summarised below.

Prior to this the general nature of alluvial fans is discussed, and model facies associations and sequences are established from the literature. Comparison between these model fan features and sequences in the Collegats Group allows: firstly, a fan interpretation to be justified, and secondly, differences between Collegats Group fans and "model" fans to be highlighted.

4.3.2(b) Alluvial fan facies associations

Modern and ancient alluvial fan sequences have been studied extensively so that their characteristic features and facies are relatively well established. Several authors have reviewed these features (e.g. Bull 1972; Heward 1978; Collinson 1986) and they are only briefly listed here.

Two types of alluvial fan are recognised: (a) fans dominated by bedload river deposits, and (b) mixed fans composed of sediment gravity and stream flow facies. Each fan type has a characteristic facies association and both have been recognised in the modern and in the ancient.

Fans dominated by bedload streams tend to be large and characterised by prolonged periods of steady flow. Because of the requirement for steady flow this type of fan is rare in semi-arid settings. Examples include: the alluvial megacones of the Himalayan foreland basin (Parkash *et al.* 1980); humid fans of the Eocene Tremp-Graus basin (Atkinson 1983); and glacial outwash fans, such as the Scott glacier fan (Boothroyd and Ashley 1975).

Fans with a mixture of sediment gravity flow and stream-flow deposits, tend to be characterised by infrequent activity and to be smaller than fans produced by periods of prolonged, steady flow. In addition, their deposits, even the stream-flow deposits, are relatively poorly channelised - a feature which Rust (1978) considers to be the product of high flow energy. The most closely studied examples of this fan type come from semi-arid areas but they also occur in other climatic settings (e.g. the humid temperate fans of central Virginia (Kochel and Johnson 1984)).

Bull (1972) considered that alluvial fans are oxidised deposits which may be fringed by playa like deposits, but which rarely contain well preserved organic material. However, this view reflects Bull's over-riding interest in semi-arid fans: Heward (1978) records coals, and Atkinson (1983) records lacustrine limestones in fan sesquences deposited in humid settings. As with other environments, the background and marginal facies of alluvial fans have characteristics governed by climate.

Another common theme of all alluvial fans is the occurrence of strong proximal-distal trends: on small fans sediment gravity flow facies occur close to the fan apex, while stream flow facies are more important distally; on fans dominated by stream flow facies a progressive change may occur in facies type. For example, a down-fan change occurs from sheet bars to longitudinal bars over 4km on the Scott Glacier fan (Boothroyd and Ashley 1975).

Other proximal-distal trends have been reviewed by Heward (1978), they include:

(i) Sediment body architecture varies downfan: stacked coarse sediment bodies occur at the fan apex, but towards the fan margin coarse sediment bodies become thinner and increasingly poorly connected as they interface with basin core facies.

(ii) Textural changes occur down-fan: grain size decreases i.e. the fan surface is graded, sorting increases and in some studies clast shape changes down-fan - actual trends depend on clast lithology.

(iii) Channels change down-fan - Heward (1978) notes a decrease in flow depth, infiltration and channel re-emergence down-fan. In addition to down-fan changes entrenchment may occur, radically changing fan configurations.

4.3.2(c) <u>Alluvial fan facies sequences</u>

Fining upwards and coarsening upwards sequences are seen within alluvial fan deposits at all scales. Their interpretation is problematical as they can reflect one or a combination of three factors: firstly, movement of the graded fan profile by change in fan size or position, for example an increase in fan size (progradation) will produce a C.U. sequence; secondly, autocyclic features such as short term fan entrenchment, small secondary fans at the end of re-emergent streams, or channel migration, thirdly, depositional processes acting on fans e.g. sediment gravity flows. These factors may be difficult to distinguish, and even where they can be distinguished it may not be possible to determine what caused them to change. For example, did a change in fan position or fan growth cause an observed C.U. sequence and was change triggered by tectonic uplift, climate, scarp retreat, or fan-head entrenchment, amongst other possible factors?

In a major review of alluvial fan sequences Heward (1978b) attempted to answer some of these problems. He started by defining three scales of sequence: basin fill sequences 100's-1000's m; megasequences 10's-100's m; and sequences m's-10'm which he sub-divided into single and multiple bed sequences. The same nomenclature is followed here except that the "sequences" (m's-10's m) of Heward (op.cit) are formally divided into intermediate scale sequences (m's-10's m) composed of several discrete beds, and small scale sequences (dm's-m's) which may be only one bed thick.

Unfortunately because of the overlap in sequence scale, the terms are genetic not observational. Heward (1978) suggested that: small-scale sequences reflect depositional processes active on fans; intermediate-scale sequences reflect short term autocyclic fan behaviour, and that megasequences give an insight into longer term fan behaviour. His diagrams (his Figures 6-8; Fig.4.18) imply that megasequences result from movement of the graded fan profile and he suggests that they can be triggered by a number of factors: tectonic uplift; climatic change; scarp retreat; decay of initial topogaphy; and major fan-head entrenchment. Heward (op.cit.) does not fully explore the origins of basin fill sequences, but suggests that they reflect depositional basin setting.

Thus, the initiating mechanisms of megasequences, and the nature of basin-fill sequences are the most enigmatic features of alluvial fan sequences. Schumm (1976) suggests that the initiating mechanisms can be ordered by scale more closely than Heward (1978) and predicts five scales of F.U. sequence as a result: first order sequences produced by tectonic uplift; second order sequences resulting from isostatic rebound or climatic change; third order sequenres representing the influence of geomorphic thresholds (Schumm 1976; 1977; see below) and fourth and fifth order sequences reflecting complex response {4.3.4(b)} seasonal and flood effects. This qualitative assessment reflects the way many workers would intuitively rank the factors, but it does not point the way to interpreting sequences in the rock record.

Heward (1978) offers an alternative method by illustating the hypothetical response of alluvial fans to different initiating mechanisms. The key features are shown by a model of a fan that develops from initial topography (Fig.4.18(a)). The fan is coarse-grained at its apex, and remains so until its final part where it fines upwards as the topography decays. By contrast a mid-fan position is characterised by an almost symmetrical megasequence with an initial C.U. sequence produced by fan progradation, being capped by a F.U. sequence as topography decays. Fans triggered by active tectonics may be characterised by continual progradation as uplift renews initial topography, or alternatively, if uplift induces channel entrenchment a new fan at toe of the old fan may result (Fig.4.18(b). Fans produced by scarp retreat are characterised by a fining-upwards megasequence and a geometry which thins towards the retreating scarp (Fig.4.18(c)).

Inferences about initiating mechanism can also be made from fan geometries: fans which overlie marked unconformities, or show stratal wedging (Riba 1976a,b; Anadon *et al.* 1986) can be inferred to have been initiated by tectonic uplift.

Small-scale sequences, including individual beds, can be modelled by the concept of geomorphic thresholds (Schumm 1977, 1981; Heward 1978). The concept considers that landscapes can accumulate, or store, material on slopes and valley floors up to a certain threshold volume above which failure, or erosion, will occur; the trigger for such failure being an exceptional precipitational event (Fig.4.19). Critically, the model predicts that small-scale basins produce similar amounts of sediment every time that they release sediment to an alluvial fan. This is in accordance with fan sequences of constant grain size, and gradual F.U. and C.U. sequences which can be envisaged as the product of gradual change in the catchment area.

4.3.2(d) Alluvial fans in the Collegats Group

Two types of fan are recognised in the Collegats Group: numerous small-scale fans fringing the External Sierras, and two large-scale fans, the Huesca fan and the Luna fan (Hirst and Nichols 1986). The characteristics, facies associations and facies sequences of the two fan types are outlined below, and then compared with the fan models discussed above. In summary, the controls acting on each fan type are listed.

(i) Small-scale alluvial fans

Eleven small-scale alluvial fan bodies were studied in detail by the author. They are located within and to the north of the External Sierras, and contrast with the thrust front fans studied by Nichols (1987) which occur at the northern margin of the Ebro basin.

Estimates of fan area range from 0.5–2km². With the exception of one 150m thick fan sequence, measured thicknesses vary from 7–80m. All the fans were sourced from intrabasinal highs. Marked changes in palaeogeography demonstrate that these highs were generated during deposition of the Collegats Group {4.5}. Structurally the highs correspond largely to anticlines and thrusts in the hangingwall of a major thrust sheet, but also to diapiric piercement structures. Stratal wedging (Riba 1976a,b; Anadon *et al.* 1986) within fans and unconformities bounding the majority of fan sequences demonstrate that fan sedimentation was initimately associated with tectonics in most but not all cases.

As a result of being sourced from intrabasinal highs the fans are characterised by sediment derived from locally outcropping lithologies. These lithologies are dominated by well consolidated limestones of Upper Cretaceous to Paleocene age, which weather to produce angular limestone clasts, $CaCO_3$ ground and pluvial waters, but very little material of sand, silt or mud grade. Locally though source areas are dominated by rocks of Triassic age - gabbro, gypsum and dark coloured micritic well bedded limestone. Weathering of these lithologies produces a range of clast sizes. The gabbro in particular produces clay, a coarse sand, and rounded gabbro clasts. Gypsum is locally broken up into discrete clasts, and also produces gypsiferous ground and pluvial water, while the limestone produces angular tablet-like fragments.

Facies associations divide the small-scale fans into two types, (a) and (b).

Two type (a) fans exist: the single 150m thick fan {4.4.3(c)(iii)} and a relatively thin fan {4.4.7(c)}. Type (a) fans are characterised by gravel scour-fill facies (Gsc) and by the presence of small scale F.U. sequences.

Type (b) fans constitute the majority of the small-scale fans. They exhibit a mixture of sediment gravity flow and stream flow facies in roughly equal proportions. In addition, they lack well developed small-scale F.U. sequences. The percentage of matrix supported debris flows is low (2-10%) in all of these fans, except the single fan with a source area dominated by Triassic gabbro where the majority of the beds are matrix supported {4.4.6(b)(iii)}. In addition, this fan is characterised by a gypsiferous fan fringe interpreted as a playa deposit.

Intermediate-scale sequences occur within both small-scale fan types, but they are relatively rare: F.U. sequences are erosively based and 2-6m thick; C.U. sequences are not erosively based and have a similar scale. The top 10m or so of the majority of the fan sequences is characterised by a sequence of beds which fine and thin upwards. Two fans are of constant grain size throughout.

Palaeocurrents, and a consideration of sediment body geometry, allow three distinct geometries to be inferred for the small scale fans: firstly, a classic cone shape {4.4.6(b)(iv)} characterised by radial palaeocurrents and sediment bodies which thin in all directions away from a fixed apical point; secondly, a bajada where sediment bodies thin away from the "mountain front" but are of constant thickness along it, also characterised by radial palaeocurrents {4.4.5(e)} and thirdly, a palaeovalley fill {4.3.4} characterised by a palaeovalley shape and axial palaeocurrents {4.4.6(b)(i)}.

Where discrete fan apexes can be seen or inferred their position is invariably located by distinct structural elements e.g. a syncline {4.4.6(b)(iv)} or a fault {4.4.5(b)} normal to a straight mountain front.

4.3.2(d)(ii) Interpretation of small-scale alluvial fans, and comparison with model facies associations and sequences

The type (a) facies association - gravel-filled scour facies, and small- and intermediate-scale F.U. is comparable to the facies association of fans dominated by steady flow.

The type (b) facies association - a mixture of sediment gravity and stream-flow deposits, and a general absence of small and intermediate scale sequences - is directly comparable to the facies association of periodically active small fans. As noted these fans have been recognised most commonly in arid and semi-arid regions, but palaeosols suggest that the prevailing climate was humid to warm arid, so that the study provides a further example of their occurrence over a wider range of climate. The rarity of small- and intermediate-scale F.U. sequences is interpreted as indicating that the type (b) small-scale fans were not markedly channelised.

The interpretation of overall fan motifs from both fan types (a&b) can be aided by comparison with the models of hypothetical alluvial fan behaviour developed by Heward (1978). In this respect it is important to note that the measured logs record positions as close as possible to the fan apex.

The most common motif seen in the small-scale fans, constant grain-size for the majority of the fan sequence and fining and thinning upwards at the top, is closely comparable to the motifs Heward (op.cit) predicts for fans developed from initial topography, or by scarp retreat. Of these, origin by scarp retreat is not favoured, as there is no evidence for the fan reworking implied by that model. An origin in response to initial topography is in accordance with the structural data, as many of the fans onlap an unconformity and are considered to immediately postdate a discrete phase of tectonic uplift which would have generated the required topography.

The second fan motif, one of constant grain-size, is seen in two fans, each of which suggest a unique interpretation: (i) the first example passes abruptly upwards into the deposits of a different depositional system {Fig.4.45(b)}. The fan sequence is considered to be incomplete and to have been terminated by a major tectonically induced palaeogeographic change; (ii) in the second example the fan is dominated by stratal wedging, implying tectonic uplift throughout {4.4.5(e)}. So the coarse grain-size is maintained throughout.

It is likely that tectonic uplift can produce a range of fan sequences and that the possibilities suggested by Heward (op.cit) and here are merely examples within that range.

4.3.2(d)(iii) Controls on small-scale alluvial fans in the Collegats Group

(I) Tectonics

The most important control on alluvial fans in the Collegats Group is tectonics. Primarily, tectonic uplift generated source areas. In addition, the majority of fans are underlain by unconformities suggesting that fan sedimentation was initiated by discrete phases of tectonic uplift.

Through the control of drainage patterns, which within the newly generated intra-basinal relief were largely consequent, tectonic features are also thought to have controlled fan size, location and shape.

Within the intra-basinal uplifts of the External Sierras two sizes of catchment area occurred: synclines and strike-perpendicular dip slopes. All of the type (b) fans were sourced from dip slope source areas while the larger type (a) fan was derived from a major syncline. Thus catchment area (km^2) is thought to control observed facies associations: the larger catchment area of a syncline dampens flashy discharges into steady flows favouring stream deposits.

Fan shape and position reflects the nature of the basin margin: point sources, produced by synclines or strike normal faults at the basin margin, favour fans with a true cone geometry; sediment aprons or bajadas are favoured by even slopes; and flow elongate fans are produced by richly fed systems which flow axially along synclines in palaeovalley settings.

(II) Geomorphic thresholds

Individual units of sedimentation, in particular sediment gravity flows and other small-scale sequences, are considered to be the product of inherent landscape instability and the breaching of geomorphic thresholds (Schumm 1977, 1981; Heward 1978). Where sediment accummulation as a result of basin evolution occurred contemporaneously with tectonic uplift, instability may have been increased by the resulting increased slopes, and resedimentation may have been triggered by tectonic events. Such a situation would reflect both intrinsic (geomorphic evolution) and extrinsic factors (tectonics; Schumm 1977).

(III) Lithology of the source area

The lithologies outcropping in the source area of individual fans controlled the type of sediment supplied to the fan and strongly influenced fan facies. Limestone source areas produced angular gravel-sized clasts deposited as relatively well sorted gravel scour facies (Gsc) and clast supported sediment gravity flows with little matrix and rare interbedded sandstones. By contrast, source areas dominated by Triassic gabbro produced a wide range of clast sizes deposited as matrix supported sediment gravity flows. The effect of gypsum source areas was seen in two ways: firstly, as resedimented gypsum clasts deposited in sediment gravity flow and subordinate stream-flow facies; and secondly, as a playa-like fan fringe reflecting gypsiferous ground and pluvial waters.

The limestone and gypsum lithologies would have also favoured a relatively low vegetation density, which in turn would have enhanced rapid runoff and peaked storm hydrogaphs likely to produce sediment gravity flow facies.

(IV) Climate

Climate has not been shown to be an important factor in controlling features of the small-scale alluvial fans. The dominant palaeosol facies the oxidised palaeosol facies (soil facies S_5; (4.2.3(d)(v)) is thought to reflect good drainage produced by relatively high fan gradients.

However, intense local rainstorms may have been favoured by a microclimate over the intrabasinal relief of the External Sierras, and could have been important in generating flashy discharges and sediment gravity flows. Although this is highly speculative.

4.3.2(d)(iv) Large alluvial fan - the Huesca fan

Hirst and Nichols (1986) recognise two early Miocene alluvial fans at the northern margin of the Ebro Basin: the Luna fan, with its apical point at the western end of the External Sierras and the Huesca fan with its apical point in the study area (Fig.4.20(a)). From a study of outcrops to the south of the Barbastro anticline Hirst (1983) and Hirst and Nichols (1986) document medial and distal fan settings in the Huesca fan. Although correlation is problematic, part of the Collegats Group documented in this study is considered to represent the most proximal part of the fan.

The type section through the Collegats Group in the study area, the section to the south of Graus {4.4.2(a)} is characterised by a 278m thick F.U. basin-scale sequence. The high percentage of sandstone bodies near the southern margin of the Barbastro anticline (62%; Hirst and Nichols op. cit.) suggests that the Huesca Fan may correspond to the lower parts of this F.U. sequence, i.e. to the Graus Conglomerate Formation {4.4.2(a)(i)}. Thus, using the work of Hirst (op.cit.), Hirst and Nichols (op.cit.), and the data presented in this study, a complete proximal-distal trend can be outlined for the Huesca fan (Fig.4.21).

The most proximal part of the fan is characterised by stacked conglomerates interpreted as the deposits of braided streams. The sediments are dominated by channel-fill sequences deposited in a broad palaeovalley or on a braid-plain. The mid-fan area occurs immediately to the south of the Barbastro anticline and is dominated by erosively-based sheet sandstone bodies, interbedded with silts and clays (Fig.4.20(b)). The sandstones are interpreted as laterally migrating or braided sandy channels characterised by fluctuating discharge. The distal part of the fan occurs towards the centre of the Ebro basin and is characterised by ribbon sandstone bodies suggesting laterally stable fluvial channels. These distal deposits interdigitate with lacustrine sediments characterised by wave ripples and oxidised muds towards the northern margin of the basin, and by grey marls and lacustrine limestones near the centre of the basin.

The location of the different fan areas persisted only during the early part of the Collegats Group: through time the fan retreated northwards, so that by the end of Collegats Group sedimentation interdistributary areas near the basin margin were dominated by lacustrine deposits previously characteristic of the Ebro Basin core. In addition, systematic down-fan changes were disrupted by the development of intrabasinal uplifts. The fan responded to these uplifts in two ways: by rapid down-fan facies changes towards highs, whereby braided stream deposits pass over a distance of 10km into terminal fan deposits {4.3.3} and secondly, by the switching of major river systems to areas between uplifts.

Compositionally the conglomerates at the fan-apex reflect a wide catchment area: 15-30% of the clasts can typically be assigned to lithologies outcropping in the Axial and Nogueras Zones, although values may be higher than this, locally up to 90%. Hirst and Nichols (op.cit.) also record the importance of Axial Zone source areas in sandstones deposited south of the Barbastro Anticline, with the Huesca and Luna Fans showing distinct mineral and clast suites.

4.2.3(d)(v) <u>Comparison with model fan facies associations</u>

The facies association of the Huesca fan is directly comparable to that of the bedload stream, type (a) fans described above. The inferred warm-humid to warm-arid climate {4.2.3(h)} and the large scale of the fan makes it closely comparable to a number of large fans seen in the modern Himalayan foreland basin (Geddes 1960; Parkash *et al.* 1980). However, the author is unaware of detailed proximal-distal or facies association studies of these fans with which to compare the Huesca and Luna fans. One difference between the Huesca and Luna fans and the modern Himalayan fans is that the Pyrenean fans terminated in a lacustrine zone in the basin centre, while the Himalayan fans link with a fluvial system which drains axially along the foreland basin ({4.3.3}; Fig.4.22). The geometry suggests that proximal-distal trends on the Himalayan fans may not be so marked.

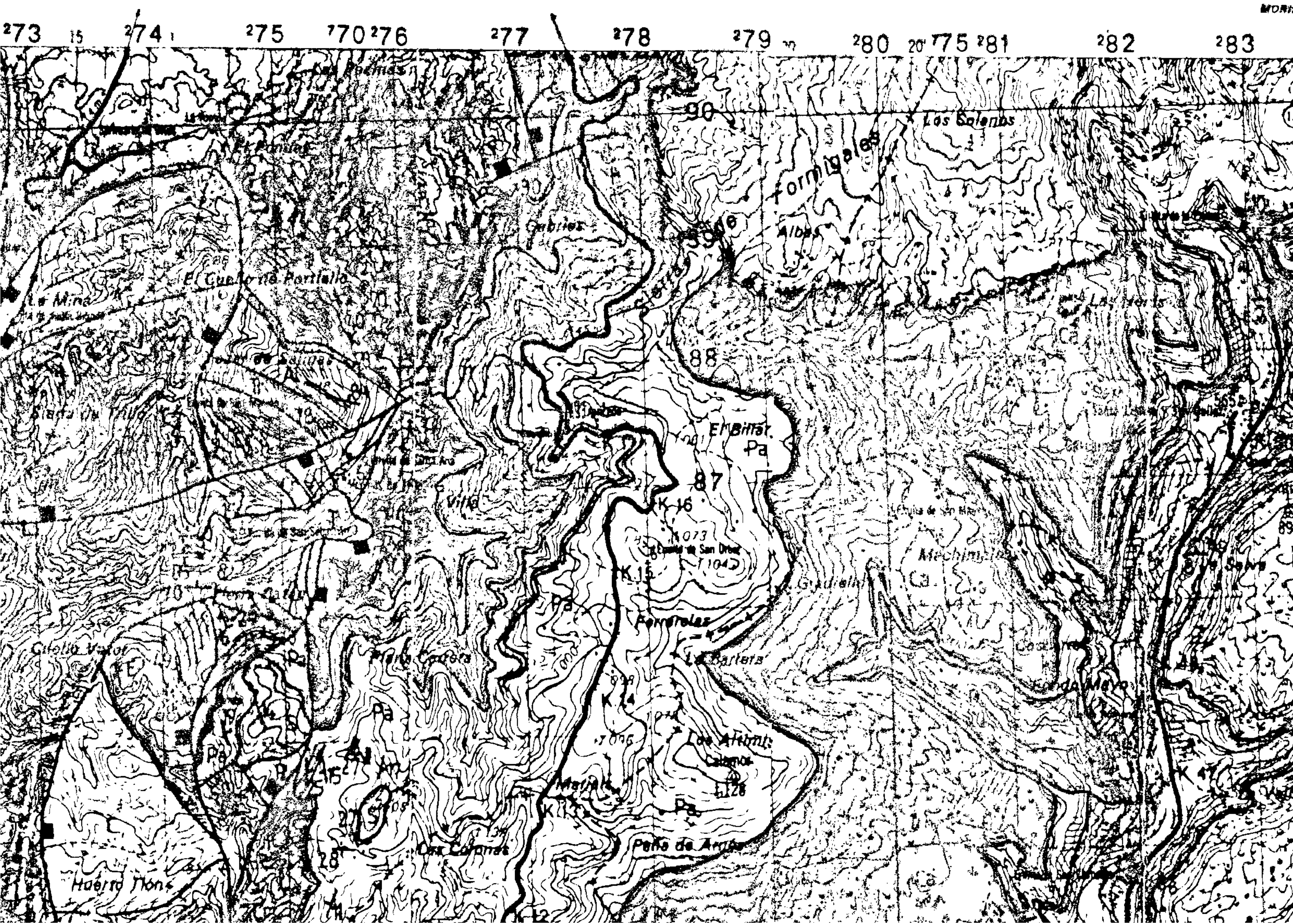
The 278m thick F.U. sequence seen in the section south of Graus is interpreted as a basin fill sequence, and is considered to reflect the gradual decay of the topographic relief of the main source area - beyond the basin margin.

A note on terminology is appropriate at this point. Hirst and Nichols (1986) use the term "fluvial distributary system" to decribe the Huesca and Luna dispersal systems, while Parkash *et al.* (1980) use the term megacone to describe similar features. In using these terms, both sets of authors emphasize the scale of the dispersal system. However, the term fluvial distributary system does not emphasise the essential fan geometry, while the term megacone is equally applicable to submarine fans. The author prefers to call the dispersal systems alluvial fans and to emphasise processes occurring on the fan to characterise the alluvial fan type - hence the term river-dominated alluvial fan following Collinson (1986).

4.3.2(d)(vi) Controls on the Huesca Fan

The fundamental controls on the development of the Huesca Fan are threefold: (i) the development of the Ebro basin and the Pyrenees which supplied it with sediment; (ii) the development of a more or less discrete input point; (iii) the internal drainage of the basin.

Hirst and Nichols (1986) argue that the position of the heads of both the Luna Fan and the Huesca Fan were tectonically controlled: the Luna Fan head lying to the west of the western tip point of the External Sierras, and the Huesca Fan head lying between the Barbastro Anticline and the Mediano (Fig.4.20(c)). Their evidence, derived from back projection of palaeocurrents to a statistically likely fan-head position, strongly suppports tectonic control for the position of the head of the Luna Fan. However, palaeocurrents for the Huesca fan back-project towards a 25km wide apical region, in part to the east of the Mediano structure, in conflict with their model (Fig.4.20(a)) but in accordance with the presence of three or more transfer systems contributing to the fan apex as suggested here (Graus, Palo and Colungo (4.5)). It is in particularly close accord with the Graus


273
274
275
770 276
277
278
279
280
775 281
282
283
Las Palmas
Los Solanos
Formigales
Albas
El Cuello na Portiello
La Mina
El Billar
Pa
87
88
90
K 16
K 15
K 14
Ferreroles
La Barbera
Los Altimir
Peña de Arrie
Las Colonas
Huerto Tions
Mechimina
Villa

within the Himalayan foreland basin (Fig.4.23). The fan has an area of $64km^2$ and a low conical form. As a result of artificial confinement the modern fan surface is dominated by only two active distributary channels. However, palaeochannels indicate that the fan was previously dominated by repeated bifurcation (Mukerji 1976). The palaeochannels also indicate that the fan is a long-lived feature, and that the position of the low sinuosity river at the fan head is relatively stable. Discharge is variable, and follows the pattern of the prevailing semi-arid monsoon climate. Typically there is flow during the three months of the monsoon, but for the rest of the year the channels are dry. Grain size analysis shows that the fan sandstones are well or very well sorted. Parkash *et al.* (op.cit) attribute this to their origin as second cycle detritus through reworking of fluvial deposits in the Siwalik hills.

Trenching revealed three distinct facies associations corresponding to channel, levee and floodplain environments.

The channel facies association commences with an erosional surface that lacks marked relief. The basal surface is overlain by a small-scale sequence which fines upwards from sandstone to siltstone and may pass into clay at its top. Sedimentary structures within the sequence indicate decreasing discharge; they commence with planar (Sp), trough (St), or ripple cross-bedding (Sr) and pass upwards into rippled (Sr), climbing rippled, or flaser bedded sandstone and silt. Parkash *et al.* (1983) note four proximal-distal variations in these facies: firstly, channel width decreases downfan from 80m at the fan apex to 25m near the fan toe; secondly, individual small-scale sequences decrease in thickness from 2m at the fan head to 0.8m at the fan toe; thirdly, multistorey sequences, 2-3 flows thick, are common at the fan head but increasingly rare towards the fan toe, and fourthly, trough and planar cross-bedded facies become increasingly rare towards the fan toe where channel deposits are dominated by plane-laminated sandstones (Sp) reflecting the shallowest flow depths.

The levee facies association is characterised by climbing ripple and trough cross-laminated sandstone and alternations of sand, silt and mud. The floodplain deposits are composed typically of massive mud or silty mud and minor amounts of cross-laminated silt.

The observations of Parkash *et al.* (op.cit) allow a model of terminal fan architecture to be developed. Critically, they consider that the majority of the sand supplied to the fan was deposited within the channels. Therefore, coarse sediment body geometry will directly reflect channel geometries, and a decrease in scale downfan, from 80m wide by 2m deep at the fan apex to 25m by 0.8m at the fan toe, can be expected. As a result of the fan geometry the number of multistorey sediment bodies, and the connectedness of sediment bodies are also likely to decrease downfan: at a rate controlled by sand supply and avulsion frequency. In this respect Parkash *et al.* (op.cit.) note that trenches in the top 2.5m of the fan indicate that the major portion of the fan is underlain by a widespread, and almost continuous layer of channel sediments.

4.3.3(b) Terminal fans in the Collegats Group

As noted above the Collegats Group forms part of a large-scale terminal fan, the Huesca fan having a radius of 70km. The terminal fans discussed here are smaller in scale, having a radius of some 10km. They developed in the proximal part of the Huesca fan, adjacent to intrabasinal highs, and are interpreted as the response of the fluvial system to blind structural lows.

Sequences interpreted as terminal fans are characterised by strong-proximal distal trends over a distance of approximately 10km. Three main features are apparent: a downstream decrease in flow depth; a downstream change from channelised to unconfined flows, and a downstream decrease in grain size and the proportion of coarse sediment (Fig.4.24).

The proximal part of a model Collegats Group terminal fan is composed of a stacked sequence of erosively-based, 1.5-3m thick, small-scale F.U. sequences dominated by the gravel scour facies (Gsc). The sequences are interpreted as channel fills deposited by a bedload river.

Over a distance of 2-4km these small-scale F.U. sequences change progressively in character, and are increasingly interbedded with monotonous silts. The erosive sequence base and sequence scale are maintained, but the gravel facies may also include gravel sheetfloods (Gs). In addition, the top half of the sequences become sandy and are characterised firstly, by sandy scour-fills (Ss) and secondly, in more distal positions by 3-4 discrete, superposed, erosively-based sheetflood deposits (Sde and Sf; Fig.4.25; Fig.4.4). The changes within the F.U. sequences are interpreted in terms of decreased flow competence, and increasingly unsteady flows, while the increased interbedding with siltstones interpreted as overbank deposits is thought to represent a distributive channel pattern.

The next change down-fan involves the loss of conglomeratic material so that the small-scale sequences are composed entirely of sand. An erosive, locally irregular sequence base is at first maintained but then lost down-fan. The sequences themselves either F.U. or C.U. and are composed of a number, commonly around five, of sandy sheetfloods (Sf). The loss of an erosive base to sequences is interpreted as the result of the loss of a channelised form marking the change to a lobe geometry. The F.U motifs are interpreted as the result of abandonment, while the C.U. motifs are interpreted in lobe-settings as the result of progradation, and in channels as the result of progressive re-establishment of flow in an old channel course {4.3.1(b)(ii)}. Beyond the lobe environment individual sheetflood deposits (Sf) punctuate a silt background. They are interpreted as recording very shallow unconfined flows at the extreme margins of the fan.

Due to flow parallel exposures, sediment bodies and fluvial architecture could not be studied systematically. In particular, variations in sediment body width are poorly known. However, channel margins are rare and most of the studied sediment bodies had sheet geometries 10's m or more wide, indicating channels and lobes 10's m wide.

4.3.3(c) Discussion

There are a number of fundamental differences between the terminal fans from the Collegats Group and the terminal fans of the literature.

Firstly, the Collegats Group terminal fans occur adjacent to highs on palaeogeographic reconstructions. They are interpreted as the response of the fluvial system to blind structural lows and as such are fundamentally the result of intrabasinal tectonics. This contrasts with the Himalayan fans where climatically induced downstream decrease in discharge is thought to be the fundamental cause of terminal fan formation.

Secondly, the Collegats Group terminal fan facies association differs in three key ways from that observed by Parkash *et al.* (1983) on the Markanda fan: (a) Parkash *et al.* (1986) recognise no F.U./C.U. intermediate scale sequences; (b) the small scale sequences Parkash *et al.*(op.cit.) recognise are thicker (2-0.8m) than those seen on the Collegats Group fans (0.1-0.3m); (c) no levee type facies are recognised in the Collegats Group fan sequences, and (d) megaripple and ripple scale cross-bedding is rare in the Collegats Group fan sequences.

The absence of recorded intermediate scale F.U. and C.U. sequences may be the product of the sampling procedure and/or a uniform grain-size - only a restricted number of the trenches dug by Parkash *et al.* (op.cit.) encountered more than one small-scale sequence, while a uniform grain-size would militate against F.U. and C.U. sequences. The lack of medium-scale cross-bedding and the dominance of plane (Sh) and low angle (Sl) lamination suggests that shallower and possibly more ephemeral flows were dominant on the Collegats Group fans. The absence of levee type facies is puzzling; it may suggest that flow was strongly confined to within channels, or that the levee facies has a low preservation potential.

The facies associations and proximal-distal trends also differ from those seen over 60km on the larger scale Huesca fan, which is characterised by stream flow from head to foot and by a lacustrine fan fringe (Hirst and Nichols 1986; {4.3.2(d)(iii)}). Thus the small terminal fans have a unique facies association and do not represent the compression of facies belts seen on the larger fan.

4.3.4 Palaeovalleys

4.3.4(a) Introduction - scales of palaeovalleys

Palaeovalleys are ancient valleys filled with sediment and preserved in the rock record.

In a study of Cenozoic Gulf Coast fluvial systems Galloway (1981) recognised three scales of river: large extrabasinal rivers; basin fringe rivers, and intrabasinal streams sourced within the basin. He considered it of paramount importance to locate the large extrabasinal rivers as they dominate sediment input and sediment dispersal. Applying this concept to the basin margin and catchment area the large extrabasinal rivers correspond directly to the transfer system of Schumm (1977; {4.1.3}) which he considers links the main catchment area to the basin, while the basin fringe and intrabasinal streams would correspond to smaller valleys and smaller drainage basins.

The author is able to distinguish two scales of palaeovalley within the Collegats Group, the transfer systsem, and the tributary valley, reflecting respectively the large extrabasinal river, and intrabasinal stream scales of Galloway (op.cit). Although the division is largely one of scale - transfer system palaeovalleys simply being larger - the recognition of the position of transfer systems is important as they control the position of the major sediment input points to the basin. Consequently the characteristics of the two scales of palaeovalley are discussed separately below.

Prior to this, models and general characteristics of palaeovalley sedimentation are discussed from the literature to allow comparison with the Collegats Group palaeovalleys.

4.3.4 (b) Palaeovalleys in the geological literature

Palaeovalleys have been studied extensively within basins as a response to sea level and climatic fluctuations (Fisk 1944; Weimer 1983, 1986) and to some extent as palaeodrainage systems (Stapp 1967; Coneybeare 1976; Padgett and Ehrlich 1978). However, at basin margins studies are less numerous, and the nature of the valley fill has been emphasised (Iwaniw 1984) rather than the part the valleys played as elements of palaeodrainage systems.

The palaeovalleys of the Collegats Group are of the basin margin type. So the following discussion is focused on literature relating to this type of palaeovalley; although in the discussion which follows the role they play as elements of the drainaage pattern is emphasised.

Iwaniw (1984) briefly reviews palaeovalley fills which are all of the basin margin type, and considers that there are two types governed by climate: arid/semi-arid, and humid palaeovalley fills, with each type divided by the presence or absence of tectonic activity (Table 4.6). However, he shows no distinct differences between valley types and the classification is considered erroneous. His review is more useful in pointing to the nature and origin of large scale sequences within palaeovalleys. Iwaniw (op.cit.) notes that palaeovalleys are dominated by F.U. megasequences. In addition, he recognises the same strong proximal-distal trends within palaeovalleys as seen on alluvial fans so that he uses alluvial fan sequence models (Heward 1978; {4.3.2(c)}) to interpret the F.U. megasequences as the results of changes in the catchment area.

The author considers that the similarity with alluvial fan sequences holds the key to the classification of basin margin palaeovalley fills and that they can be classified primarily in the same way as alluvial fans {4.3.2(b)} as being dominated by stream flow facies, or by mixed stream flow and sediment gravity flow facies. A complete classification of palaeovalley fills would allow a sub-division of the stream flow category as suggested by Schuum (1968) reflecting the three-fold division of river types (Schumm 1977; {4.3.1(a)}; Fig.4.26).

Palaeovalleys can be considered to aggrade in two ways (Schumm 1977): firstly, as a result of upstream control, whereby tectonic uplift or climatic change causes sediment supply in excess of stream capacity to transport load, and secondly, as a result of downstream control where base level rise in the lower part of the valley, produced by tectonic uplift across the valley or inclusion of the valley into the basin by basin aggradation, reduces river gradient below the value required to transport the supplied load.

Schumm (op.cit) considers that the two types of aggradation have distinct sedimentary responses: downstream controls produce backfilling in which the coarsest material will be deposited first, and the finer material will be deposited further downstream resulting in F.U. sequences at the scale of an individual channel and through time at the scale of a paleovalley; upstream controls produce down-filling and a C.U. sequence with the coarsest material being deposited first at the head of the valley, and then progressively downstream as the river adjusts its slope to the increased sediment load. The resulting sequence may well be at the megasequence scale, but Schumm (op.cit) is unclear about this. Schumm (op.cit.) also envisages a third type of sedimentary response, that of vertical filling of a valley segment richly fed by a number of tributaries. These ideas, in particular the idea that downstream controls as well as source area decay or climatic change can produce a F.U megasequence, are an important complement to the alluvial fan model interpretations of palaeovalley fills.

Schuum (1977) also considers the history of a valley and drainage basin, and shows how they may respond in a complex fashion to drainage basin rejuvenation. He reports a number of experiments involving the rejuvenation of model drainage systems at their downstream ends, by the marked lowering of base level to produce an inflection or "nick-point" in the river gradient. Downstream of the nickpoint the stream is steeper than the equilibrium profile so that it aggressively works headwards eroding the valley floor. As it does so it branches out to affect tributary valleys which provide progressively more sediment, eventually in excess of the trunk stream transport capacity. Consequently the main valley begins to aggrade and continues to do so as the nick point(s) work back up the system. When the tributaries become adjusted to the new base levels sediment production and sediment load decreases and a new phase of incision commences. In modern valleys this results in river terraces. In the ancient it could be seen as major erosion surfaces within valley fills. Schumm (op.cit) calls this kind of behaviour a "complex response" as one event can trigger a complex response from the fluvial system. The key feature of the model appears to be the development of a "nick-point" at the downstream end of the catchment area. Two factors indicate that nickpoints may be relatively common in basin margin palaeovalley settings: (i) tectonic uplift along basin margin faults would produce nick points, and (ii) Schumm (op.cit) notes that little is known of the detailed morphological response of the drainage basin to rejuvenation, so that uplift of the drainage basin by rotation about a hinge axis near the basin margin could have the same effect as the production of a nick point.

4.3.4(c) Transfer system palaeovalleys in the Collegats Group

(i) Facies associations and geometry

Two transfer systems have been documented in the Collegats Group: the Palo transfer system and the Sis-Graus transfer system thought to link discrete palaeovalley fills at the Sierra del sis {4.4.1} and at Graus {4.4.2(a)}. Of these the Sis-Graus system is the best preserved and best exposed and will be used as a model transfer system.

The most proximal part of the transfer system (at the Sierra del Sis) is preserved as a straight palaeovalley segment 14km long, 7km wide and 400m deep. Its position is controlled by a N.W.-S.E. trending syncline. The valley is based by an erosive 1km wide and 160m deep notch, above which the fill passively onlaps the margins of the syncline.

The valley fill is composed of a stacked sequence of 10-20m thick multistorey, multilateral conglomerate sheets. There is no evidence for large scale unconformities of the type produced by complex-response. Individual storeys are erosively based, 1.5-3m thick, massive conglomerate beds (Gm) which may F.U. gradually and pass transitionally but rapidly into 10-20cm thick, sandy tops dominated by plane laminated sandstone (Sh). Clasts are rounded and well imbricated indicating flow to the S.S.E. They are also poorly sorted, though usually greater than 15cm and sometimes over 2m in diameter. Provenance studies show that, with the exception of thin scree-like deposits at the margin of the palaeovalley, all the clasts are derived from the Axial and Nogueras Zones.

The coarse nature of the deposits and the dominance of Gm facies suggests that the valley fill was deposited by bedload rivers at high stage when the river bed was devoid of braid bars. The small-scale F.U. sequences and the stacked sheets indicate two scales of interruption to this flow: the former is considered to be a short term variation, perhaps on a seasonal scale or on the scale of an individual storm; the latter is thought to represent a longer term effect of uncertain origin. The dominance of Gm facies and the coarse grain size are compatible with a proximal position in a large fluvial system.

The preserved valley segment at Graus is some 25km S.W. of the Sis locality in the direction of the palaeoflow. The valley fill is erosively based, 6km wide, and preserves a relief of 60m. The fill lies with angular unconformity on older (Middle Eocene sequences). It is composed of a stacked sequence of erosively based 1-2.5m thick conglomerate beds. There is no evidence for internal unconformities within the valley fill. The beds are dominated by a variety of conglomerate traction current deposits. Individual beds commonly fine upwards at their tops and pass transitionally into sandstone facies, either massive or plane laminated (Sh). The clasts are well rounded and imbricated indicating flow to the S.E. By size they are dominantly of pebble or small cobble grade (6.4-12.8cm in diameter): the largest clasts are 30cm in diameter. Pebble counts show that only 15-30% of the clasts are derived from the Axial and Nogueras zones.

The predominance of coarse sediment bodies and the observed traction current facies suggest that the valley fill deposits at Graus are the deposits of bedload rivers. Specific facies (Gp; Gf Gl) indicate that the rivers were braided at times, but the erosive bases and the predominance of F.U. sequences interpreted as the product of waning flow suggests that sedimentation was dominated by channel-fill during decreasing discharge. The cycles are thought to either reflect the seasonal climate or to be produced by periods of intense storms {4.3.1(b)(i)}.

4.3.4(c)ii <u>The cause of incision and valley fill</u>

The causes of the incision seen at the base of both valley segments and subsequent aggradation are difficult to assess.

Incision could be the result of three factors, acting alone or in conjunction: (i) fall of base-level downstream due to thrust induced loading (ii) rise of base-level upstream due to tectonic uplift and (iii) climatic change producing increased runoff and higher stream discharges. Each of these factors would increase stream efficacy. However, there is no independent evidence for climatic change or marked basin subsidence during incision (well dated time-equivalent deposits with which to test these hypotheses do not occur) while three features indicate marked uplift in the catchment area: (i) folding at the Sierra del Sis prior to valley filling (ii) an angular unconformity at the base of the Graus palaeovalley fill, and (iii) a marked change in sediment provenance at the Sierra del Sis between the Collegats Group and the underlying sequence interpreted as the result of the development of the Nogueras Zone {4.4.1(c)}.

Valley fill was produced by supply of sediment in excess of stream capacity. As noted above {4.3.4(c)(ii)} four factors may influence palaeovalley aggradation: (i) tectonic uplift or (ii) climatic change upstream of the preserved valley segment, and (iii) tectonic uplift across the valley or (iv) inclusion of the valley into the basin by basin aggradation downstream of the preserved valley segement.

Independent evidence suggests that each of these factors, bar climate, may have influenced aggradation, but that tectonic uplift in the drainage basin was the dominant cause.

As noted above tectonic uplift in the source area is thought to have produced the initial palaeovalley incision. A model of a single thrust front during alpine compression, suggests that the thrust front remained in the Nogueras and Axial Zones until part way through sedimentation of the Collegats Group, when the thrust front became emergent within the basin {2.8}. The prolonged location of the thrust front in the Axial and Nogueras Zones would have maintained high rates of sediment production and induced aggradation.

Emergence of the thrust front within the basin, i.e. downstream control, would also have induced aggradation. However, the External Sierras fold deposits considered to be the downstream equivalents of the Graus palaeovalley fill, so that their emergence post-dates palaeovalley aggradation. Palaeogeographic maps show a marked expansion of the extent of the basin through time {4.5}. Again, though, this expansion post-dates the Graus palaeovalley so that it is not thought to have controlled palaeovalley aggradation. The possibility of climatic change influencing sediment supply rates cannot be fully evaluated, but in the light of the thickness of palaeovalley deposits (up to 400m) and the positive evidence for tectonic control it is not thought to have been a major factor.

Aggradation as a result of continuous tectonic uplift and a high rate of sediment supply is in accordance with two further features of the transfer system palaeovalley deposits:(i) the absence of valley-like erosion surfaces within the palaeovalley - if sediment production dropped off, as it would after a short lived phase of uplift, then following a complex response model (Schumm 1977;{4.3.4(b)}) incision woulld result; and (ii) approximately constant grain-size within the valley fill.

4.3.4(d) Tributary palaeovalleys

A number of tributary palaeovalley deposits are developed on the intrabasinal relief of the External Sierras. The preserved valley segments contrast with the transfer system palaeovalley: firstly, by their position, and secondly, by their relatively small size, with preserved segments ranging from 0.1-2km wide and from 0.1-7km long.

Tributary valleys are initiated in two ways: firstly, by tectonic uplift which produces incised consequent drainage flowing down synclines or down strike-perpendicular dip slopes, and, secondly, as a result of river capture which produces palaeovalleys that lie across structural trends. Valleys flowing down synclines are wider (1-2km) than those flowing down strike perpendicular dip-slopes (0.1km): a feature thought to reflect the relatively small catchment area of dip-slope valleys and the wavelength of folding in the External Sierras.

The causes of palaeovalley filling are difficult to assess. Of the four possible factors outlined above {4.3.4)c)(ii)} only tectonic uplift in the catchment area, and tectonic uplift downstream of the preserved valley segment, can be shown to have influenced palaeovalley filling. Climatic change and basin aggradation are known to have occurred but they cannot be correlated with valley fills.

Tectonic uplift in the catchment area is thought to have been important in the initial aggradation of all consequent streams, and where deformation continued, to have been important throughout valley fill. Initial aggradation is thought to have resulted from high rates of sediment production following initial palaeovalley incision after uplift. Indeed one small palaeovalley {4.4.4(b)(iii)} has a major erosion surface within the valley and is interpreted as the product of a complete complex-reponse cycle (Schumm 1977; {4.3.2(c)(ii)}).

Where tectonic uplift continued throughout sedimentation the valley fills are complicated by: unconformities which may extend across the entire valley but are most marked at valley margins, and by marginal as well as axial sediment input. Marginal sediment sources occur as discrete F.U. megasequences interpreted as fan bodies generated by tectonic uplift. Axial systems have similar characteristics including F.U. megasequences interpreted as the product of tectonic uplift. The facies associations and sequences of both these types of sediment input can be described by alluvial fan models.

Aggradation as a result of tectonic uplift downstream of the preserved palaeovalley segement, can be inferred for the palaeovalley initiated by river capture {4.4.5(d)}. The preserved valley segment erosively overlies, and therefore postdates, alluvial fan and terminal fan sequences which indicate the emergence of the External Sierras to the south and a north facing palaeoslope. The valley fill is composed of massive gravel beds (Gm) and gravel-filled scour (Gsc) facies, but lacks major erosional sufaces and small-scale F.U. sequences, interpreted elsewhere as channel bases and channel fills respectively. The clasts are well rounded, and they have a strong Axial-Nogueras Zone provenance signature. They are also well imbricated and indicate flow to the S.E. across the External Sierras. The fill is interpreted as the product of a south-flowing river which breached the External Sierras and captured part of the transfer system. It is overlain by a second alluvial fan and a sequence of terminal fan deposits which indicate the re-establishment of a north-facing palaeoslope and the end of paleovalley sedimentation. The absence of well developed channels in the palaeovalley fill is thought to represent rapid aggradation, caused by raising of base level across the External Sierras at the onset of the second uplift phase.

4.4 DESCRIPTION OF THE KEY LOCALITIES

The Collegats Group is well exposed. Locally exposure is complete, allowing facies sequences, architecture, and the structural context of the successions to be studied in detail. As examples of different parts of the fluvial system the following key localities stand on their own. No attempt at this stage is made to link localities, but they are discussed as far as possible in a downstream fashion, from proximal to more distal settings: outcrops of the transfer system are discussed first {4.4.1; 4.4.2(a)(ii)}; then outcrops representing the proximal part of the Huesca fan {4.4.2}; followed by discussion of key localities around and within the intrabasinal uplifts of the External Sierras {4.4.3-6}; the final localities {4.4.7} include the youngest and most distal facies associations seen within the Collegats Group.

4.4.1. The Sis Conglomerate

The Sierra del Sis, located between the Rio Isabena and Rio Ribagorzana (Fig.1.20) is one of several of mountains which dominate the northern margin of the Tremp-Graus basin. The mountain is composed of a number of conglomerate formations, each distinct by clast provenance and/or separated by marked unconformities. The youngest of these conglomerates, for which the formation name the "Sis Conglomerate" is proposed, is believed to be of Oligo-Miocene age and is discussed here. The Sis Conglomerate is interpreted as the most proximal part of a transfer system palaeovalley.

4.4.1. (a) The gross geometry of the Sis Conglomerate

The Sis Conglomerate is 5-7km wide and 14km long in a N.E.-S.W. direction, extending from the southern margin of the Nogueras Zone, across the outcrop of Cretaceous and Jurassic rocks to the northern margin of the Paleocene-Eocene Tremp-Graus basin. Its base is largely unconformable on older rocks. At its N.W. margin, west of Castrocit G.R.[0695] the Sis Conglomerate onlaps tilted Upper Eocene sediments that dip to the S.E. (Fig.4.27). The onlap surface preserves slopes of 26° and a relief of at least 160m, and possibly as much 380m or more (Castrocit to Las Tosas G.R.[0797]) . The northern and eastern margin of the conglomerate was not studied in detail, but field-glass study indicates the existence of an onlap surface, similar to that at Castrocit, at Obis G.R.[0892] preserving a slope of 25° and a relief of 180m. There is no obvious angular unconformity in the southern part of the Sis Conglomerate body: except below Tozal del Sis G.R.[0389] where a V-shaped notch, 160m deep and 1km wide, is cut into the underlying Upper Eocene conglomerates and filled by horizontal beds of the Sis Conglomerate (Fig.4.28).

4.4.1(b) Sedimentology of the Sis Conglomerate

The conglomerate is composed of 10-20m thick multistorey and multilateral sheets, dominated by poorly sorted, well imbricated a(t)b(i), cobble or boulder conglomerates. Clasts are usually greater than 15cm and occasionally over 2m in diameter. Where recognised, discrete storeys are 1.5m-3m thick, and composed of massive, erosively-based gravels (Gm) which F.U. slightly and pass at storey tops transitionally but rapidly into 10-20cm of plane laminated sandstone (Sh). Palaeocurrents consistently indicate flow to the S.S.E.

With the exception of a single sample (sample S_4, composed of angular clasts of Jurassic - Cretaceous age) pebble counts indicate that the conglomerates were sourced exclusively from Permian and older rocks (Table 4.7). Different samples suggest that distinct sources were present at different times. For example some sequences are composed of clasts derived exclusively from Hercynian metasediments, while others are composed solely of Permian volcanics.

4.4.1(c) Intepretation and Discussion

The Sis conglomerate body is interpreted as a palaeovalley fill. Its northeastern part, between Castrocit and Obis, is confined within a N.E.-S.W. trending syncline that folds rocks of Upper Eocene age. The fold dies out southwestwards, so that the Sis Conglomerate lies disconformably on the Upper Eocene conglomerates. The exception to this geometry, the V-shaped notch below Tozal del Sis, is interpreted as being the product of fluvial incision. As such it represents the first record of the fluvial system which occupied the Oligo-Miocene palaeovalley.

The nature of the fill, the causes of incision and subsequent filling of the valley (aggradation) have been discussed above {4.3.4(c)(ii)}.

The Nogueras and Axial Zone clast provenance is in marked contrast with the Paleocene to Upper Eocene conglomerates which underlie the Sis Conglomerate. The older conglomerates were sourced from a granodiorite body (probably the Maladeta Granodiorite) and from Jurassic and Cretaceous limestones. Major changes in the palaeovalley catchment area are implied between Upper Eocene and Oligo-Miocene times: reorganisation which largely cut off granodiorite supply to the foreland basin for the first time since the Paleocene, and exposed Permian and older rocks as major sediment sources for the first time. This reorganisation and the uplift implied from the sedimentology of the Sis Conglomerate is interpreted as the result of the further development of the Nogueras Zone {2.8.}. The distinct clast assemblages within the Sis Conglomerate suggest that a number of specific sources supplied sediment to the palaeovalley. Given the WNW-ESE trending structures of the Nogueras Zone distinct sources of this type may be interpreted as indicating a reticulate drainage pattern in the catchment area. By contrast, the importance of granitic material in the Paleocene to Upper Eocence conglomerate bodies suggests that the drainage basin at that time extended further to the north, to the Maladeta, and that the pattern is more likely to have dendritic.

4.4.2. The Graus Conglomerate Formation

A major escarpment which dominates the town of Graus [8075] can be traced from La Masada, near Naval, in the west [7074] through Graus to Cabeza de la Serra, near Viacamp, in the east [0366]. The escarpment is characterised by stacked conglomerate sheets, which have south-directed palaeocurrent indicators and clast compositions indicative of a northern source area. An informal formation name the "Graus Conglomerate Formation" (G.C.F.) is suggested for them. The G.C.F. unconformably overlies Middle to Late Eocene sediments. Across the unconformity marked changes occur, in facies, in palaeocurrent directions and in clast provenance, indicating major palaeogeographic changes between the Eocene and the Oligo-Miocene.

Three localities within the G.C.F. are discussed below together with two localities which represent its immediate downstream equivalent. The Formation is considered to occupy a transitional position between the end of a transfer system palaeovalley and the most proximal part of the Huesca fan.

4.3.2.(a) Graus

South of Graus the G.C.F. forms the basal part of a 278m thick F.U. megasequence (Fig.4.29). The sequence commences with stacked cobble conglomerates. Above these the percentage of sandstone and siltstone increases, conglomerate clasts decrease from cobble to pebble grade, and a change in facies and facies sequence occurs. However, because of the incomplete exposure it is unclear whether the changes are gradual or whether there is a marked change at around 140m. The changes are most clearly illustrated by discussing the section in two parts: (i) the base of the sequence 0-140m and (ii) a well exposed section near the top of the sequence (219-240).

(i) The base of the sequence

The base of the sequence is composed of stacked conglomerate sheets, 1-2.5m thick and 10's - 100's m wide, which onlap an irregular unconformity surface. The onlap preserves a 21°, 60m high, east-facing slope to the N.E. of Graus [790747], while field-glass study suggests a similar, west-facing, slope 6km to the east [850727]. Individual conglomerate sheets are composed of small cobble to pebble grade clasts, and occasional outsize clasts up to 30cm in diameter. Typically 15-30% of the clasts can be assigned to Axial-Nogueras Zone source areas (Table 4.7). The sheets are erosively based and usually composed of a single traction current facies, of which gravel scour fills (Gsc) are the most common; but other traction current facies (Gp; Gm; Gl; Gt; Gf) and complex facies sequences (usually based by the massive gravel facies Gm) also occur. Imbrication, Gsc and Gp facies indicate flow to the S.W. The sheets are usually of constant grain size but they may F.U gradually and/or pass rapidly, but transitionally, into sandstone facies (Sl; St; Ss) in their top 10-20cm. The rare silts and fine sandstones are red-brown in colour (soil facies S_5).

Interpretation

The preserved relief indicates that the base of the sequence was deposited in a 6km wide palaeovalley. The palaeovalley is interpreted as the downstream equivalent of the transfer system seen at the Sierra del Sis, on the grounds of: an Axial Zone source; palaeocurrents indicating a source to the N.E.; and the lack of intervening structures. The increase in Jurassic-Eocene clasts between the Sis Conglomerate Formation and the G.C.F. is interpreted as the result of tributary rivers joining with the main transfer system (Table 4.7).

The coarse grained nature of the deposits and the traction current facies suggest that the conglomerates are the deposits of bedload rivers. The erosive base, F.U. nature and falling stage facies transitions seen in the small-scale sequences, suggest that the majority of the deposits are channel fills produced during decreasing discharge {4.3.1(b)(ii)}. In addition, the facies assemblage suggests that the rivers were periodically braided: facies characteristic of bar relief occur - planar cross-bedded gravel (Gp); oblique gravel scours (Gf) and low-angle cross-stratified gravel (Gl) - while facies characteristic of slough channels are rare {4.3.1(b)(i)}. The soil facies S_5 is interpreted to reflect good drainage through the conglomerates.

(ii) The top of the sequence 219-240m

The sequence is composed of seven loosely connected pebble-sandstone sediment bodies interbedded with pedogenically altered silts and clays (Fig.4.30). The coarse sediment bodies are 1.5-3m thick and have sheet geometries. Their bases are erosive and broadly scalloped (concave upwards). Clear storeys were not recognised in the lower five sediment bodies so they can be described as simple. The uppermost sandstone body has a multistorey ribbon geometry; the overlying conglomerate is composed of gravel scour fills and fines upwards slightly.

The six lower sediment bodies F.U. They commence with 0-1m of gravel filled scours (Gsc) transitionally overlain in turn by 1.5-2m of sand-filled scours (Ss; Se), 0-1m of rippled or plane laminated sand, and variable amounts of siltstone and mud (F). Discrete soil profiles and erosively based sandstone units within the siltstone and mud are taken to indicate the transit'on between in-channel and overbank sedimentation; although the geometry cannot be demonstrated unequivocally - i.e. palae sols or sandstones cannot be shown to extend past discrete channel margins. Three soil facies are recognised hydromorphic pseud gley mottling S_1; hydromorphic gley mottling brown background S_2; and hydromorphic mottling S_4. S_1 and S_2 are associated with scattered disorthic calcrete glabules C_2. Three distinct profiles occur, each recognised by increased reddening towards the top.

A number of bedding planes preserve a broad-leaf flora.

Interpretation

On the grounds of their erosive bases, and traction current facies assemblage the coarse sediment bodies are interpreted as fluvial channel fills. The high percentage of in-channel sandstone and conglomerate allows the channels to be classified as bedload dominated. However, facies indicative of significant bar relief (Gp; Gl; Gf; {4.3.1(b)(i)}) are absent, the channel pattern is not thought to have been braided. As there is no evidence for bank accretion the small-scale F.U. motif of the five lower sediment bodies is interpreted as the product of decreasing discharge and bed aggradation. The interpretation is n accordance with a bedload stream model (Galloway 1981; {4.3.1(b)(i)}) but the systematic F.U. nature of the sequences is somewhat at odds with the erratic F.U. sequences suggested by Galloway (1981).

The soils and the broad leaf plants indicate a seasonal climate. Soil facies S_1, S_2 - C_4 are interpreted as Ustalfs indicating a warm climate intermediate between arid and humid. They probably supported vegetation with a ramifying root system {4.2.3}.

(iii) Origin of changes through the megasequence

Upward through the 278m thick basin fill sequence there is an increase in the percentage of sandstone and siltstone and a decrease the diameter of conglomerate clasts from pebble/cobble to pebble grade. Throughout the coarse sediments are thought to be largely in-channel deposits of bedload streams. Changes in facies suggest that these streams were braided in the lower parts of the sequence but not braided in the upper parts. Palaeosols are increasingly well-developed upwards.

The scale of the basin-fill sequence indicates that it was produced by tectonic uplift (Heward 1978; {4.3.2(c)}). The interpretation is in accordance firstly, with the Sis Conglomerate Formation {4.3.1(a)} and secondly, with the unconformity at the base of the Graus Conglomerate Formation which indicates basinwide uplift.

F.U. may result either from increased subsidence (so that coarse sediment bodies become progressively disconnected) or from a decrease in coarse sediment supply. Sediment supply is controlled by the source area (tectonic uplift and/or climatic change) and by sediment dispersal patterns.

The influence of subsidence rates cannot be tested: no absolute chronology exists for this sequence.

Source area controls can be assessed, in part, by (i) a basinwide palaeogeographic synthesis {4.5} and (ii) a consideration of palaeosols. The former suggests that the thrust front changed from being active in the Axial/Nogueras Zones, to being emergent at the northern margin of the Ebro basin during the Oligo-Miocene. The resulting cessation of uplift in the catchment area may have been the primary cause of F.U. in the megasequence. Although pedogenic features increase in importance upwards through the sequence it is difficult to use them to assess the role of climatic change - the change from well-drained soils (S_5) to Ustalfs may reflect a decrease in river gradient or an increase in precipitation. In the latter case if effective precipitation in the catchment area was initially low a small increase may have encouraged vegetation and therefore decreased sediment yield (Schuum 1977). The reasoning is speculative but the possibility cannot be discounted.

There is no evidence for changes in sediment dispersal patterns.

In summary, the log records a fluvial basin-fill sequence, dominated by bedload rivers, braided at the base, passing upwards into the deposits of unbraided rivers. Coarse grained sediment bodies usually have sheet geometries, are 1.5-3m thick and F.U. F.U. is believed to be the result of either gradual abandonment of channel tracts or a single rainy season. The megasequence was initiated by tectonic uplift, dominantly in the north (Axial and Nogueras Zones). Fining upwards was probably the result of switching of tectonic uplift to within the basin and consequent denudation in the source area, although increased subsidence rates and climatic change may have played a role.

4.4.2.(b) The Laguarres and Castellarnes sections

4.4.2(b)(i) Introduction

The Laguarres section is located on the Benabarre road south of Laguarres [8971]. It includes only the Graus Conglomerate Formation (G.C.F.) and is dominated by bedload stream deposits considered to have been deposited in the proximal part of the Huesca fan.

The Castellarnes section [8670] is located 2.5km south of the escarpment formed by the G.C.F. Cross-sections and palaeocurrents show that the top 50m of the section is an approximate downstream equivalent of the Laguarres section. The Castellarnes section is composed of bedload stream deposits and subaerial sediment gravity flow/sheetfood facies, and considered to represent the proximal part of a terminal fan.

4.4.2.(b)(ii) The Laguarres section

Description

The section is composed of stacked, erosively-based 0.5-3m thick conglomerate sheets (Fig.4.31(a)). Internally the sheets may be simple or multistorey. The majority of the storeys are characterised by a slight F.U. tendency. Traction current gravel facies (Gm; Gp; Gl; Gsc) comprise the major part of individual storeys, but these facies may pass transitionally but rapidly into sandstone facies (Ss; S) in their top 10-20cm. More complex facies transitions do occur e.g. Gm-Gsc-S-F, and Gm-Gp-Gm-S.

In addition to the dominant F.U. sequences, two C.U. sequences 1.2 and 2.0m thick occur. The C.U. sequences commence with an erosive base overlain by gravel scour fill (Gsc) and pass transitionally upwards into massive gravels (Gm).

Throughout the section palaeocurrents indicate flow to the S.S.E. Clasts are poorly sorted and have a strong Axial and Nogueras Zone provenance signature (Table 4.7).

Interpretation

The F.U. facies sequences are comparable to those seen in the basal part of the Graus log {4.3.2(i)a} and are interpreted similarly, as channel fill sequences deposited by braided bedload rivers during falling stage.

By contrast the facies in the C.U. sequences indicate an increase in flow power. They are interpreted as having been produced by the reoccupation of an abandoned channel segment {4.3.1(b)(ii)}.

The temporal scale of both the F.U. and the C.U. cycles is problematic. Following the arguments advanced above {4.3.1(b)} the small-scale nature of the cycles suggests that they are either seasonal or the product of single storms.

The architecture and sediment body geometry suggest that the sediments were deposited by laterally unconfined, braided rivers on a wide braid plain.

4.3.2.(b)(iii) The Castellarnes section

Description

The measured sequence at Castellarnes is 119m thick. It is composed of interbedded sandstone and conglomerate units in which sheetflood facies (Sf) are abundant (Fig.4.31(b)). Small- (1-3m) and intermediate-scale (6-20m) F.U. sequences are recognised. A single intermediate-scale (14m) C.U. sequence and a number of units of constant grain size up to 10m thick also occur, but there is no overall F.U. or C.U.

The small-scale F.U. sequences either correspond to a single event or they are composite. Single bed F.U sequences are closely comparable to those seen at Laguarres 2.5km to the north: they are erosively based and dominated by gravel traction current facies which pass near the top of sequences transitionally into sandstone facies. Composite small-scale F.U. sequences are also erosively based, but the top half of the sequence is composed of sand and the sequences are characterised by a number of facies transitions. Sequences commence with traction current conglomerate facies (Gm, or Gsc) or the gravel sheetflood facies (Gs) and pass up into a stacked sequence of sandy sheetfloods (Sf) 10-30 cm thick.

The intermediate scale F.U. sequences show two different facies sequences: a rigid Gsc-Sf-F sequence and a general Gm-Gsc-Sf-S sequence. The single C.U. sequence commences in silt and coarsens upwards gradually via a series of composite and then single bed small-scale F.U. sequences to pebble/cobble grade.

The exposures were not conducive to architectural or sand body geometry studies. However, one cliff face exposure suggests that the sandy sheetfloods occur both as fills to abandoned channels and overbank deposits (Fig.4.25).

Statistical analysis of clast size data revealed a low positive correlation between MPS and Bth (0.586, significant at the 1% level; Fig 4.32).

Interpretation

On the grounds of an erosive sequence base and F.U. motif the conglomerate dominated small-scale F.U. sequences are interpreted as channel fills deposited by bedload rivers during decreasing discharge. The absence of facies indicative of significant bar relief (Gp; Gl; Gf) suggests that the river was not markedly braided.

The erosive base of the second type of small-scale F.U. sequence suggests that it too represents a channel-fill. However, the predominance of sheetflood facies suggests that the channel was characterised by unsteady peaked discharge - each individual erosively based bed being interpreted as the product of a discrete short-lived (hours or days) flood event (c.f. McKee *et al.* 1965). As this second type of small-scale F.U. sequence is finer than the conglomerate dominated sequences, but has the same palaeocurrents and pebble composition, it is considered to be a downstream equivalent of the conglomerate dominated sequences. So the conglomerate dominated channels must have suffered the same unsteady flow.

The medium scale F.U. and C.U. sequences are too thick to be related to channel abandonment, or progradational fill. They must be the product of larger features, perhaps the progradation, or retreat, of depositional lobes on the terminal fan.

The low positive correlation between MPS and Bth suggests that the data set is not in close accordance with the model for sediment gravity flows {4.2.(i).}. The value is in accordance with the type and range of observed facies and compares favourably with similar sequences (Steel 1974; {4.2.1(d)}).

Discussion

A marked downstream fining and change in facies occurs between the Laguarres and Castellarnes sections. The change is interpreted as the downstream transition from a laterally extensive braid-plain to the proximal part of a richly fed terminal fan. The changes are believed to relate to the relief caused by a F.W. ramp to the External Sierras, expressed currently at the surface by inliers of Cretaceous and Paleocene limetones (Fig.2.17(e) and Map 1 [849674] to [908639]). The Oligo-Miocene sediments clearly onlap the limestones which, therefore, formed a high during sedimentation towards which south-dipping gradients decreased and eventually reversed. A sudden drop in gradient may well have produced the observed facies changes.

4.4.2.(c) Viacamp road section

The Viacamp road section is located near Viacamp, 10km E.N.E of Benabarre on the N-230, G.R.[0166]. The measured sequence is 82m thick and unconformably overlies the Escanilla Formation (Fig.4.33).

Description

The section is composed of 1-3m thick sheet sediment bodies, which are stacked at the base of the section but become increasingly poorly connected upwards. Sheet bases are erosive but only locally irregular. Internally the sediment bodies are simple or multilateral: individual storeys are characterised by 0.6-3m thick F.U. cycles. The cycles exhibit simple facies sequences commencing with conglomerate traction current facies (Gm; Gsc; Gt; Gl) and passing more or less abruptly in their top 10-30cm, into sandstone facies facies (S; Ss; Sh). More complex transitions occur, but such sequences tend to be thicker indicating that they are multistorey. Three 30-40cm thick matrix supported debris flow units also occur.

Palaeocurrents from imbrication in all facies suggest a S.E. directed palaeoflow.

The background is composed of even, yellow-orange (5YR 5/6) silts and fine sands, but exhibits no other pedogenic features.

Statistical analysis of plots of maximum particle size (MPS) against bed thickness (Bth) reveals a positive correlation both for all data (0.642 significant at the 0.1% level) and for sediment gravity flows, (0.85 significant at the 0.5% level; Fig.4.34).

Interpretation

On the grounds of an erosive sequence base and F.U. motif the small-scale sequences are interpreted channel fills deposited during decreasing discharge. The gravel traction current facies and the coarse grained nature of the channel fills are compatible with a bedload river model. The rarity of facies indicative of marked bar relief (Gp and Gl facies are absent and only one example of the oblique scour-fill facies (Gf) occurs) suggests that the river bed was not markedly braided.

The gravel sheetflood facies (Gs) suggests that discharge was "flashy". In showing a mixture of traction current and sediment gravity flow facies the facies association is comparable to that seen at Castellarnes differing mainly in the absence of the sandy shhetflood facies (Sde; Sf). It may be that the Viacamp section should occupy a position between the Laguarres and Castellarnes sections on a terminal fan model.

The statistical analysis partially confirms the observed facies range, since the low, positive correlation between MPS and Bth for the complete data set is comparable to values derived from similar facies assemblages in other studies (Steel 1974). However, the high positive correlation for the subset of sediment gravity flow facies is probably spurious: the regression curve has a positive intercept on the Y-axis, suggesting that the beds are the deposits of cohesive debris flows, a feature at odds with the observed facies which have erosive bases and are considsered to be the deposits of turbulent flows.

In summary, the section is interpreted as having been deposited by laterally unconfined, S.E. flowing, bedload rivers with flashy discharges. A proximal position on a terminal fan is envisaged. The lack of pedogenic features is interpreted to reflect good drainage through the conglomerates.

4.4.2(d) The Embalse conglomerate

Introduction

On the eastern side of the Embalse de Joaquin Costa at the entrance to the Esera gorge a sequence of conglomerate sheets occurs [7867]. The formation name the Embalse Conglomerate is suggested for these conglomerates. The conglomerate is considered to be a downstream equivalent to the Graus Conglomerate Formation and to predate the formation of the External Sierras.

Description

The Fmbalse Conglomerate is folded by a backthrust, the Embalse backthrust at the northern margin of the External Sierras. At its base the conglomerate is composed of stacked conglomerate sheets: upwards the sheets are increasingly interbedded with silts and sand. The conglomerate sheets are 1.5-2m thick and composed of well imbricated pebble-cobble clasts dominated by gravel traction current facies. They have not been logged in detail but 64 palaeocurrent readings from imbricated clasts, suggest flow to the south, directly towards the External Sierras (Fig.4.35). Clast compositions have a strong Axial-Nogueras Zone signature.

Interpretation

On the grounds of their coarse grain-size, erosive bases, appropriate scale, traction current facies, and imbrication the conglomerate sheets are interpreted as bedload river deposits.

Three features indicate that the conglomerate predates the backthrust at the northern margin of External Sierras: firstly, the conglomerate is folded by the backthrust; secondly, palaeocurrents indicate that the rivers flowed directly towards the External Sierras, and thirdly, the facies show no evidence for the conglomerate having aggraded as a result of the development of the backthrust (c.f. {4.4.5(d)}).

A cross-section, combined with consistently south-directed palaeocurrents and compatible provenance, suggests that the Embalse Conglomerate is a downstream equivalent of the Graus Conglomerate Formation (G.C.F). The implication is that the G.C.F. predates the External Sierras.

4.4.3. The Olvena Gorge

4.4.3(a) Introduction

The Rio Esera traverses the External Sierras in a steep sided gorge below Olvena [735656] offering three-dimensional exposures of Oligo-Miocene sediments and the folded Cretaceous to Eocene sequence on which they lie. These exposures allow the growth of the External Sierras to be documented through unconformities and the response of the Collegats Group fluvial system to uplift. In addition, they provide examples of palaeovalley, alluvial fan and braided river environments.

The structural evolution of the area is considered first, setting a framework to discuss three time-equivalent sedimentological sections (Fig.4.36). The structural evidence and the distinct logs are summarised at the end of the following section, after discussing related outcrops at the southern margin of the External Sierras (4.4.4).

4.4.3(b) Structure

(i) Description

The structure of the area is shown on Map2 and Figure 4.37. The geometry of the section is constrained by field mapping and borehole data. In addition, the structure seen at depth on Figure 2.17(e) is projected on to the section line: again the External Sierras are shown as being "allochthonous" overlying a decollement in Upper Eocene salt.

The structure of the Cretaceous-Eocene sequences is dominated by three large evaporite-cored folds. Two of these folds have no preferred sense of vergence, but one verges towards the hinterland - a backfold cut by four backthrusts T_1-T_4.

The Collegats Group records the development of these folds. It outcrops mainly within a major syncline north of the large backfold. The syncline plunges to the west exposing deeper structural levels in the east.

The basal unconformity is complex, both in a north-south and in an east-west sense. In a north-south sense the basal part of the Collegats Group is divided into two parts by parasitic anticline, anticline A (Fig.4.37). The geometry is a downplunge projection of onlaps onto anticline A seen at [656759] (Fig 4.38). In an east-west sense an important along-strike change occurs at [747652] where a normal fault downthrows the Tremp and Alveolina Limestone Formations to the east, so that the basin *margin* lithology changes along the gorge. The fault is truncated by, and so predates, backthrust T_4.

Discrete units within the Collegats Group record a number of tectonic events. The first unit, unit (a), a conglomerate lithosome, was deposited with very low angular unconformity on slightly tilted limestones, and then folded into a tight syncline (Fig.4.39). The syncline is truncated by a marked angular unconformity traceable for the length of the E.-W. section of the gorge (Fig.4.40(b)). The unconformity is most prominent below Olvena and is termed the "Olvena Unconformity" in the sections that follow. Again these relations are projected downplunge onto Figure 4.37. Above the angular unconformity a second conglomerate lithosome, unit (b) occurs. Dips within this unit gradua ly decrease upwards, particularly along the line of section Figure 4.41.

(ii) Interpretation

The deformed Cretaceous to Eocene sequence defines two distinct phases of deformation: an early phase (1) during which the majority of the shortening occurred forming the major folds; and a later phase (2) when the backfold was thrust through. The thrusts cut bedding at high angles and cannot be represented faithfully on the restored section because they affected already deformed rocks.

The geometry of the Collegats Group refines this structural history. The unconformity at the base of conglomerate (a) implies major early uplift and erosion without marked folding. The lack of major early folding is important as it implies that the Collegats Group was involved in nearly all of the deformation seen in the Cretaceous-Eocene rocks. The syncline developed in conglomerate (a) and the unconformity which overlies it represent a distinct phase of deformation. The gradual upward decrease in dip seen in unit (b) may either be the result of gradual tilting (Fig.4.42(a)) or to a single phase of folding (Fig.4.42(b)). The marked decrease in dip along the section line is largely due to the backthrust T_4 (Fig.4.42(c)). Significantly T_4 cuts unit (b) so that the early syncline, and the Olvena unconformity which overlies it, developed as a result of the main folding phase.

(iii) Discussion

The major folds in the Esera Gorge and other folds in the External Sierras either lack a preferred sense of vergence, or have an approximately equal preference for structures that verge to the foreland or hinterland (Fig.4.37; Fig.2.17(e); Fig.2.20(e)). The geometry is characteristic of folds formed above a salt decollement horizon (Davis and Engelder 1985) and, therefore, compatible with the Upper Eocene salt decollement shown at depth on the sections presented here.

But the folds are cored by Triassic evaporites, ophite and limestones which the section implies are allochthonous. So the folds may have been formed prior to emplacement over the Eocene decollement horizon (c.f. Fig.2.17(d)). However, the Triassic in the fold cores was diapirically active at least until the end of the Oligo-Miocene (3.6). This indicates that the fold cores must be linked to salt at depth, and since the Triassic is cut off as a source, the folds are likely to be cored by Eocene salt. In summary, the folds may have developed over a Triassic decollement, over the Eocene decollement, or over both.

The deep early erosion of the Cretaceous-Eocene sequence means that the deformed state section for these rocks, as shown dotted on section 2, did not actually exist - the folds were continuously eroded as they formed. Thinning of the sequence by erosion might have caused the marked vergence of the backfold, with the thinned limb deforming more easily.

For the interpretation of the sedimentary sequences that follows the structural evolution outlined here is clearly important. In addition to discrete periods of uplift and folding, compartmentalisation of the area by anticline A can be expected to have been a major influence on patterns of sedimentation.

Sedimentary logs

4.4.3(c) The Backthrust log

The section is 249m thick and located above a small backthrust T_4 to the east of Olvena ([743653]). The section can be divided into five distinct parts: (i) a basal "fan" (0-27m); (ii) a silt dominated sequence (27-78m); (iii) a thick west flowing braided stream sequence (78-228); (iv) a palaeovalley deposit (228-246m); and (v) an upper unit of markedly different provenance (Fig.4.43).

Part (i)

Description

The lowermost part of the sequence strikes parallel to and onlaps Cretaceous limestones which dip at 87° to the north, 17° steeper than the overlying Collegats Formation. The onlap is an expression of the Olvena unconformity. Part (i) is composed of stacked conglomerates and can be traced laterally along the length of the gorge. Towards the east, where deeper structural levels are exposed, the lithosome can be seen to be considerably thicker (2-3 times) and to unconformably overlie an earlier conglomerate sequence (unit (a)) which represents the oldest Collegats Group sediments in the gorge. The structural implications of this first sequence have been discussed above. Unfortunately it proved inaccessible, although field-glass study suggested it to have similar sedimentological properties to part (i) which overlies it.

Part (i) is dominated by sub-angular clasts, whose provenance could be determined on fresh surfaces. All determined clasts have lithologies which can be related to the local Cretaceous-Eocene sequence: there are no quartzites and the minor sand sized fraction is dominated by lithic limestone fragments. Clasts range up to 0.8m in diameter and sorting is generally poor.

The facies are dominated by metre-scale, clast supported, sediment gravity flows (Gs 42%; Gdp 26%) and similar scale beds composed of gravel-filled scours (Gsc 26%). Sandstone and siltstone facies and fine-grained matrix within the conglomerate facies are rare. Sequences within part (i) are restricted to a fining and thinning towards the top: there are no well-developed small-scale F.U. sequences. Locally a distinct alternation occurs between metre-scale gravel beds and thin sand beds, but facies contacts are sharp. Palaeocurrents from within the Gs and Gsc facies suggest palaeoflow parallel to the onlap surface, to the east.

Interpretation

The angular, poorly sorted nature of the clasts coupled with the close match between clast lithology and the local Cretaceous to Eocene stratigraphy suggests that the sediment was derived from a local source. The suggestion is supported by the dominance of sediment gravity flow facies and the absence of small scale F.U. sequences: features which are indicative of the proximal part of a small alluvial fan ({4.3.2(b)}; c.f Trollheim type of Miall (1978) Fig.4.17). The conglomerate which underlies part (i) is interpreted as having the characteristics of an alluvial fan deposited in a palaeovalley setting. There is no direct evidence for the direction of drainage of this early palaeovalley. However, the west directed plunge of the syncline in which it rests suggests that it drained to the west (Fig.4.44(a)).

A section which runs through the line of the log, suggests that deposition of the Collegats Formation was still divided into two distinct parts by anticline A (Fig.4.37). So the "fan" was probably deposited as a palaeovalley fill: the consistent east directed palaeocurrents support this, suggesting that the valley drained to the east and was closed to the west throughout the deposition of the sequence (Fig.4.44(b)). The gradual fining and thinning upwards at the top of the sequence is discussed in the following section.

Part (ii): 27-78m

Description

This part is dominated by light-brown (5YR 5/6) silts (F) which show reduction spots and minor hydromorphic colour mottling (soil facies S_5) some bioturbation and rare rain spot imprints.
The silts are punctuated by coarse sediment bodies composed of sandy sheetfloods (Sf) and gravel-filled scours (Gsc). Sf units

may occur individually or superposed to form m-scale F.U. sequences (Fig.4.45). The F.U. sequences have markedly erosive bases and can be seen to be channelised in one example. The Gsc facies occurs in three 2-2.5m thick C.U. sequences, two of which are superposed.

Palaeocurrents in the Gsc facies indicate flow both to the west and to the east. The single measurement of p.c.l. developed in the Sf facies gives an E-W flow orientation.

The Gsc facies is composed of well-rounded limestone clasts. But it also has a low proportion <<1% of quartzite clasts, some clearly derived from the Nogueras Zone, and a quartz rich matrix, while the Sf facies is composed of subrounded, quartz-rich sandstones.

Interpretation

Part (ii) shows dramatic changes from part (i): in clast provenance, clast texture, grain size, facies and facies sequences. The increased sorting and roundness of conglomerate clasts indicates that they are further travelled than those in the basal fan. More telling are the presence of quartzite clasts and a quartz-rich matrix in the Gsc facies, and the dominance of quartz grains in the Sf facies, both indicating that anticline A was breached and that a proportion of the sediment was sourced from as far away as the Axial Zone (Fig.4.44(c)). A qualitative assessment of this proportion can be made by comparison with conglomerates deposited by the Graus Transfer system, a time equivalent transfer system {4.4.2 (a)}. For those conglomerates 15-30% of the clasts can typically be related to lithologies outcropping in the Nogueras and Axial Zones (Table 4.7). So the contribution of far travelled clasts to the Gsc facies in part (ii) was probably minor. By contrast, the dominance of quartz grains in the Sf facies is well explained by a direct source from such a transfer system.

The F.U. sheetflood (Sf) sequences are interpreted as minor channel fills, possibly part of a terminal fan system {4.3.3} which would have fed the individual sheetflood units isolated in silts. On the grounds of provenance the channels are interpreted to be linked to the transfer system inferred to exist on the northern side of anticline A.

The C.U. gravel-filled scour (Gsc) sequences are interpreted as the result of channel abandonment and subsequent re-establishment of flow along the channel course {4.3.1(b)(ii)}. The channels are believed to relate to a local fan which reworks the Sf facies producing the observed low percentage of allochthonous clasts.

The increased rounding of the limestone clasts and the change to dominantly stream flow facies (Gsc) indicates a more distal setting for the locally sourced conglomerates than that seen in part (i). To account for this a change in the sediment dispersal patterns is suggested, with the main conglomerate input now being derived from a larger catchment area to the east: a reversal in palaeovalley gradient and opening of the western end of the palaeovalley is implied (Fig.4.47). The larger catchment area would have slightly damped flashy discharges, favouring steady flow and traction currents. In addition, its most likely input point would have been at the "elbow" of the present-day gorge, 2km from the line of the log, so that a relatively distal setting could have been established.

The F.U. at the top of the basal fan (part i) the incoming of allochthonous clasts and the postulated change in sediment dispersal patterns are interpreted to reflect the same underlying trend: gradual loss of the topography produced by the main folding phase, a result of aggradation and denudation.

Part (iii): 78-228m

Description

This part of the section is dominated by more or less stacked conglomerates which form 42 % of the section. The conglomerates comprise a number of facies: (i) traction current facies - gravel scour fills (Gsc 77%) and massive gravels (Gm 10%); and (ii) sediment gravity flow facies - poorly-sorted (Gdp 8%) and clast-supported (Gdc 2%) gravel debris-flows, and gravel sheetflood facies (Gs 5%). The conglomerate traction current deposits are either arranged in small-scale F.U. sequences 1.7-4.8m thick, or superposed with no F.U./C.U. A single sequence Gm-Gs-St has also been recognised. In addition, an intermediate-scale C.U. sequence (20m) composed of six small-scale F.U. sequences occurs at the base of the sequence part (iii).

Conglomerate clasts are typically subrounded to rounded. Observed lithologies are closely comparable to those seen in the Gsc facies in part (ii) below, although the percentage of allochthonous clasts increases to as much as 10% in some beds. An exception is the absence of allochthonous clasts in some of the sediment gravity flow beds. Large (31cm diameter) conglomerate intraclasts occur at the top of the sequence (223m).

Palaeocurrents are dominantly towards the west, clearly so from 78 to 135m, although above this reversals occur associated with a part of the sequence where sediment gravity flows are important.

Interpretation

The sequence can be divided into two parts: small-scale F.U. sequences and the deposits of sediment gravity flows.

On the grounds of erosive sequence bases, F.U. motif and gravel traction current facies , the small-scale sequences are interpreted as channel fills deposited by bed-load rivers during decreasing discharge. The absence of facies diagnostic of bar relief (Gp; Gl; Gf; {4.3.1(b)(i)}) suggests that the rivers were not braided. As in the gravel-filled scour facies (Gsc) in part (ii) above, the pebble composition of 1-10% allochthonous clasts is markedly less than in the conglomerates deposited by the laterally equivalent transfer system. So the conglomerates are interpreted to have been deposited by a local source. The additional similarities of texture and facies, indicating streamflow in a midfan setting allows the adoption of a model similar to that suggested for the Gsc facies in part (ii) below. Figure 4.37 suggests that somewhere within part (iii) anticline A is likely to have been buried by sedimentation. The postulated transfer system at this stage may have been kept away from this area by high rates of local sediment input. The model must be modified by the removal of the Sf facies as none are present in part (iii). Presumably they occured, since the quartzite pebbles are present, but they must have been rapidly reworked.

The absence of quartzite pebbles in the sediment gravity flow facies suggests that they had a source area different from the traction current deposits - one which suplied only local debris. An intermittently active local fan is suggested. The suggestion is supported by palaeocurrent reverals in the gravel-filled scour facies (Gsc) interbedded with the sediment gravity flow units - the implication being that the local fan also produced some of the Gsc units.

The 20m thick C.U. sequence at the base of part (iii) is interpreted as the result of fan progradation.

Conglomerate intraclasts at the top of part (iii) are believed to herald the initiation of uplift which caused the development of part (iv).

Part (iv) 228-246m

Description

Part (iv) fills a steep sided (30°) V-shaped, E.-W. orientated trough, eroded into the top of part (iii) (Fig.4.46). In one place the trough wall is an overturned block which exceeds 20m in length. The block is composed of facies and has a clast compositon directly comparable to part (iii). The trough fill is composed of gravel scour-fills (Gsc) and minor sandstones. It dips at 5° to the north, 17° less than the top of part (iii). Four 2-3m thick, erosively based, F.U. sequences occur, but there is no overall fining or coarsening tendancy.

Clasts within the fill are dominantly limestone (Table 4.7). Intraclasts of sandstone and conglomerate reworked from part (iii) are important directly above the trough margin where they may form 10% of the clasts. The percentage of allochthonous clasts is low, commonly less than 1%. Palaeocurrents are consistently orientated towards the west.

Interpretation

Part (iv) is interpreted as a minor valley fill. The angular discordance between beds of the valley fill and beds of part (iii) indicates tectonic uplift prior to valley fill. The uplift is believed to be the primary cause of the fluvial incision which cut the valley. The large block which lies at the base of the valley is considered to have been resedimented from part (iii) by mass wasting processes. Together with the high angle of the valley sides the block indicates that the valley sides were immature, and that the valley was cut and filled rapidly.

The erosive sequence bases, coarse grain-size, predominance of gravel scour fills (Gsc) and F.U. motif suggest that the valley fill was deposited by bedload streams devoid of braid bars.

The filling of the valley is suggested to have been the result of a "complex response" (Schumm 1977). Initial uplift raised local base level, increased stream efficacy and caused incision. But as incision worked its way into the catchment area sediment supply rates increased beyond stream competence so that the valley rapidly aggraded (4.3.4(b)).

Part (v)

Description

Part (v) forms the top part of the logged section. It is composed of stacked gravel scour-fills (Gsc) and massive clast supported gravels (Gm). No distinct sequences can be recognised. Palaeocurrents are directed towards the west. This part is distinguished from parts (iii) and (iv) by a marked increase in allochthonous pebbles which range from 15-30% of the total pebble population. The change in clast composition is reflected in a change in bedding style - part (v) forms smoothly rounded outcrops while underlying outcrops tend to be angular. The distinct style allows the unit to be traced laterally along the length of the gorge (Map 1) and to be used as a datum on which to hang logged sequences. An informal name is suggested for the conglomerate "the Olvena Member".

Interpretation

Because of the high percentage of allochthonous clasts part (v) is interpreted as the deposit of an extrabasinal river sourced in the Axial and Nogueras Zones: probably the river inferred to have been the source of allochthonous clasts in parts (ii) and (iii). The implication is that anticline A, and the effect of local sediment input have been completely swamped by this system for the first time. The interpretation is compatible with the coarse grain-size and the observed gravel traction current deposits.

4.4.3(d) Northern end of Esera Gorge

Introduction and structural setting

The N.E. end of the Esera gorge offers spectacular sections through the Collegats Group to the north of anticline A (Fig.4.38). A 215 m thick section has been measured from [765669] to [758673] (Fig.4.47).

Bed dips within the section decrease steadily upwards: the base of the section lies with low angle unconformity on the Alveolina Limestone Formation; beds in the middle part of the sequence thicken away from, and are folded by, a major fold within the Tremp Formation; while beds at the top of the sequence are apparently unaffected by the fold (Fig.4.48).

The geometry implies that deposition of the Collegats Group in this area commenced after initial uplift and erosion of Eocene sequences, and continued throughout the folding: a history closely comparable to that determined for the "backthrust log". However, in this example no discrete tectonic pulses can be determined from structural evidence alone.

The measured log can be divided into distinct parts by clast provenance, and palaeocurrent trends. Four parts have been recognised - three parts (i) to (iii) on the main log and a fourth part (iv), on a short log which is laterally equivalent to a section near the base of the main log (Fig.4.47).

Part (i) Basal Fan 0-25m

Description

The base of the sequence erosively overlies strongly weathered Alveolina Limestone. The weathering profile is composed of a transitional sequence of unweathered bedrock, closely packed angular limestone blocks each slightly rotated with respect to their original positions as bedrock, and fine-grained red sandstone with scattered angular limestone clasts. There are no pedogenic mottling, bioturbation, rhizolith or concretionary features.

The overlying sequence is composed of stacked pebble-cobble conglomerates. Discrete multistorey sheet geometries can occasionally be recognised. Sediment gravity flow deposits are predominant (Gs 27%; Gdp 12%; Gdc 10%). Traction current deposits (Gsc 36%; Gm 11%) comprise the remainder. The top 7m of the sequence fines and thins upwards, with the incoming of massive sands (S) and sand-filled scour facies (Ss). No well developed small-scale sequences occur. Palaeocurrents are dominantly to the south and southwest.

The clasts are poorly sorted and angular to subrounded. Lithologies which can be matched exclusively with locally outcropping limestones (Cretaceous-Alveolina Limestone).

Interpretation

The dominance of conglomerates deposited by sediment gravity flows, and the absence of erosively based small-scale F.U. sequences, suggests that the sequence was deposited in the proximal part of a small alluvial fan {4.3.2}. The gravel filled scours (Gsc) and massive gravel beds (Gm) are interpreted as the products of bedload rivers. However, the absence of erosively based, small-scale F.U. sequences interpreted elsewhere as channel fills suggests that the fan surface was not markedly channelised. Clast provenance and palaeocurrents indicate that the fan was supplied exclusively from local source areas to the N. and N.E. A very restricted source area of around 1km^2 is implied, bounded to the N.E. by a major N.W.-S.E. trending backthrust [7867]. The small source area would have produced flashy discharges during storms, a feature reflected in the dominance of sediment gravity flows. Palaeocurrents, sediment body geometry, and the geometry of the entire lithosome, suggests that the sequence was produced by a small sediment apron rather than a true fan.

The apron and its underlying unconfomity are interpreted as the result of a discrete uplift phase, while fining and thinning upwards at the top of the lithosome are thought to be the product of source area decay.

Part (ii)

Description

Part (ii) extends from 25 to 183m and is composed of conglomerates, sandstones and siltstones. It is distinguished by the high percentage of allochthonous clasts (5-70%) within conglomerate units. The conglomerates are dominated by gravel filled scours (Gsc) and massive gravel (Gm) traction current deposits. A single bed of planar cross-bedded gravel (Gp) and a number of gravel sheetflood beds also occur. The sandstones are typically the deposits of sandy sheetfloods (Sf or Sde). The siltstones are characterised by soil facies S_5. The conglomerates form complex sheet sediment bodies composed either of erosively-based, small-scale F.U. cycles (based by Gsc, Gm or Gs, and passing up into Ss, Sh, St or S) or superposed units of constant grain size (Fig.4.4(b)(i)). Sandstone beds also have sheet geometries, they occur in three forms: (i) superposed above an erosive base forming units of constant grain size; (ii) small-scale C.U. sequences with erosive or non-erosive bases, and (iii) isolated in silts. In addition, there is a single 3m thick sandstone unit characterised internally by west-dipping low-angle erosion surfaces, and ripple and trough cross-bedding indicating flow obliquely across those surfaces to the west (Fig.4.49(a)). Other palaeocurrent indicators suggest flow to the S or S.W. with the exception of units from 35 to 47m, where flow is to the N.W. This part of the sequence (35-47m) is also notable in having the lowest percentage of allochthonous clasts 0-5%.

Interpretation

On the grounds of erosive sequence bases, traction current facies, and F.U. motif the conglomeratic small-scale sequences are interpreted as channel fills deposited by bedload rivers during decreasing discharge. The rarity of the planar cross-bedded gravel facies and the absence of other facies indicative of bar relief (Gl; Gf; {4.3.1(b)(i)}) suggests that the rivers were not strongly braided.

The sandstone facies sequences are interpreted as a downstream progression in a terminal fan setting {4.3.3} with the sequences representing (i) in-channel deposition; (ii) progradational channel fill (erosively based sequences) and channel end lobe progradation (non-erosively based sequences); and (iii) individual flows. Clast types are the same in both the sandstone and the conglomerate facies. So it is likely that the braided river channels supplied sediment to these crevasse/terminal river systems.

The low-angle, erosive surfaces in the 3m thick sandstone unit are interpreted as the product of megaripples descending the lee face of a sandy bar or sandwave (Fig.4.49). The west-directed paleocurrents from this facies are at a high angle to other palaeocurrents suggesting that the bar may have developed in a crevasse channel. Palaeocurrents in other small-scale sandstone sequences are parallel to palaeoflow directions measured from the braided channels, indicating that both terminal lobes and crevasse channels were important (Fig. 4.44(b)). Interpretation of the facies as terminal river deposits partially explains the south-directed palaeocurrents. However, since discrete conglomerate sediment bodies can be traced throughout the gorge (e.g. the Olvena Member corresponds to the conglomerate 154-159m on this log) the braided river systems must have eventually turned westwards on encountering the relief generated by the major backthrust.

Clast lithologies are comparable with the "transfer system" of the Graus Conglomerate Formation: sediment is believed to have been sourced largely from north of the Eocene basin. This together with S. and S.E. directed palaeocurrents suggest that the backthrust shown to the N.E. of the logged section was not a major barrier to south-directed sediment supply. Even though, as discussed above, it continued to grow during this time, thus affecting the large-scale architecture of conglomerate bodies.

The anomalous palaeocurrents and clast composition in the section from 35-47m are discussed in part (iv) below.

Part (iii)

Description

The top of the section is characterised by two dramatic changes: firstly, in conglomerate clast lithologies, with the percentage of allochthonous clats dropping sharply from 70% in an underlying bed to values of 0 or <1% and, secondly, in palaeocurrents which are directed to the east (Fig.4.47).

Pebbles are typically subrounded, and moderately to well sorted. Conglomerate beds are composed of gravel scour-fills (Gsc

79%) and gravel sheetfloods (Gs 21%) and are interbedded with sandstones (Ss, Sd, Sf, S) in a background of silts (F). The silts are even light-brown in colour (5YR 5/6) with some beds showing hydromophic gley mottling, soil facies S_2. The Gsc facies are arranged into two C.U. sequences 2 and 3.5m thick. Two 1m thick C.U. sequences can also be seen within sandstone beds.

Interpretation

The gravel-filled scour facies (Gsc) is interpreted to have been deposited by bedload rivers. The absence of facies indicative of bar relief (Gp; Gl; Gf; {4.3.1(b)(i)}) and the lack of small-scale F.U. sequences suggests that the rivers were neither markedly braided nor channelised. The mixed facies association of sediment gravity flows and traction current deposits is compatible wuth a mid-proximal position on a small alluvial fan {4.3.2(b)}. Clast lithologies and palaeocurrents suggest a local source to the west of the outcrop area. Although no unconformity could be seen at the base of the sequence, tectonic uplift is believed to have initiated the fan: uplift was probably concentrated at the major backfold. The local development of soil mottling suggests that the fan surface/floodplain may have been less well drained than in part (ii) below.

The silts which dominate the sequence are of unknown origin. They may either have been supplied from a local source or from a vestige of the transfer system believed to have deposited part (ii).

Part (iv)

Part (iv) is a lateral equivalent of the basal section of part (ii) (37-61m) (Fig.4.47). The section is dominated by reddish-brown, fine sandstones containing planispiral gastropods. The fine sandstones are punctuated by conglomerate and sandstone beds which are dominatly composed of sediment gravity flow facies (Gd; Gs; Sd; Sf) but gravel scour-fills (Gsc) also occur. Gravel beds contain angular to subrounded, poorly-sorted limestones, dominantly from the Alveolina Limestone Formation, and well-rounded clasts of weathered gabbro; but they are devoid of allochthonous clasts. The coarse sandstones are composed of quartz and carbonate or fragments of weathered gabbro. Palaeocurrents within the Gsc and gravel sheetflood (Gs) units indicate flow to the west.

Interpretation

The composition of conglomerates, sandstone and siltstones indicates that a suite of sources supplied sediment to the area. The quartz/carbonate Sf sandstones are interpreted as the distal end of the terminal rivers seen in part (ii) some 300m to the N.W. In contrast, conglomerate beds composed of Alveolina Limestone clasts are believed to have been supplied from a local source. Similarly, the gabbroic clasts and sands are interpreted as being derived from a Triassic cored fold. Since a major transfer system existed to the north the most likely source for the Triassic sediment is to the S.E. - the diapir at Aguinaliu [8064] which is capped by gabbroic material. This area may have also supplied the Alveolina Limestone clasts, and the high percentage of fine sandstone. The facies assemblage and the west-directed palaeocurrents are compatible with a small alluvial fan sourced from the south-east.

4.3.3(e) The Western end of the Esera gorge

(i) Structure

The structure of the area is complex. It lies to the east of a structural transfer zone (i.e. a strike-slip accommodation fault) which is postulated to separate the south vergent recumbent fold at Urbiego ([7370]; Fig.2.17(e)) and the major backfold south of Olvena.

Figure 4.50 shows the structural evolution of the area. Along a section line 1km to the east of the major transfer fault, section X-X', the core of the major the backfold is cut by E.-W. striking reverse faults: two early south-dipping faults and a later north dipping fault. Towards the west these faults terminate against a small transfer fault, west of that fault the same shortening is accommodated by a south directed imbricate fan.

As before (4.3.3(a)) the structural history can be further refined by a consideration of the Collegats Group. Its base lies unconfomably on Paleocene and on Cretaceous limestones (Fig.4.51). It is composed of angular locally sourced limestones, not studied in detail, but believed to be deposited in a proximal alluvial fan environment. Bed dips fan upwards above the basal unconformity through an angle of 60°. Alluvial fan sedimentation is abruptly terminated by the imbricate fan (Fig.4.50; section Y-Y') which is, in turn, onlapped by the succeeding Collegats deposits. These deposits are 160m thick and preserve 100m of relief, much of which is erosional (Fig.4.51). The basal 140m of this upper part of the Collegats Group has been studied in detail and is discussed below.

Structural interpretation

A four stage structural and sedimentological history can be outlined (Fig.4.50; Fig.4.51). The nomenclature which follows refers to features illustrated on Figure 4.51.

(i) Development of the large backfold and postulated transfer zone. The anticlinal crest of the backfold was preferentially eroded, exposing Cretaceous rocks in its core, and producing a complex erosion surface (E_1 and E_2; Fig.4.50(a))

(ii) Backthrusting along thrusts T_{1-3}, shown on Figure 4.37 and on Figure 4.51(a) initiating alluvial fan sedimentation (fans

F_1 and F_2) and tilting (F_1; Fig.4.50(b)).

(iii) Forelandward directed thrusting (T) and reverse faulting (R) terminating alluvial fan sedimentation; Fig.4.50(c)).

(iv) Gradual onlap (O) of the relief generated by thrusting and erosion.

(ii) Olvena road log

Description

The logged sequence runs from [727652] to [724655] along the road to Olvena commencing above the imbtricate fan shown on section Y-Y' (Fig.4.50; 4.51). It consists of conglomerate, sandstone, silt and clay (Fig.4.52). Sediment bodies are erosively based and have sheet and ribbon geometries. The ribbon conglomerates have relatively massive fills which F.U. slightly and thin (20cm) wings which interbed with background silts. Sheet conglomerates may also be massive, more commonly they F.U. gradually, commencing with traction current facies, predominantly gravel-filled scours (Gsc) and passing transitionally upwards into sandstone facies (Ss; St; Sf). In addition, small-scale sequences of sandy sediment gravity flows also occur. These sequences may be of constant grain-size. or they may fine or coarsen upwards. One such C.U. sequence has a channelised base and each bed is separated by a layer of mud/silt with desiccation cracks. Megasequences can also be seen: a paired C.U.(50-70m)-F.U.(70-110m) sequence in the middle of the section and a C.U. sequence (117-127) at the top of the section. Both types of megasequence are composed of small-scale sequences.

The conglomerate clasts are usually subrounded and moderately well sorted. The base of the section is devoid of allochthonous pebbles. Where they first appear they constitute 16% of the pebble fraction. Above that point, the percentage falls to 6-4%, but increases upwards to 30-42% marking the base of the Olvena Member. There are no facies changes related to this compositional change. At the base of the section (6-28m) the sequence of clays and silts is punctuated by green, well-sorted, medium-grained sandstones which show a distinctive suite of sedimentary structures - particularly large-scale, trough cross-bedding and possible wave ripples. In addition, iron encrusted burrows and nodules are common (Fig.4.53).

Soil facies are well developed and vary systematically through the section: (1) gley soils at the base (S_3); (2) pseudogley soils (S_1; S_2) with or without calcrete nodules (C_2) in the middle, and (3) even light brown (5 YR 5/6) sediments with or without light greenish grey (5GY 8/1) reduction spots at the top (S_5). Significantly the green sandstones are associated with gley soils (S_3).

Interpretation

The conglomerate-sandstone sediment bodies are interpreted as the fills of relatively stable channels (ribbon bodies) or unstable channels (sheet bodies).

On the grounds of coarse-grain size, erosive sequence bases, F.U. motif, and traction current facies, the sheet conglomerate bodies are interpreted as channel fills deposited by bedload rivers during decreasing discharge. Facies indicative of bar relief are absent so that the channels are not thought to have been braided.

The coarse fill and channel shape of the ribbon bodies suggests that they too are the product of bedload streams.

The small-scale sequences of sandy sediment gravity flows are interpreted as the downstream equivalents of these channels. Those which have markedly erosive bases are considered to be channel fills; those which do not are considered to be lobe deposits. The C.U. and F.U. motifs are interpreted as the product of progradation and abandonment respectively. These facies and facies sequences are indicative of a terminal fan setting.

The conglomerate wings to the ribbon conglomerate bodies suggest that the channels which deposited them maintained their position for some time, and that their fill is more likely to be due to a flood season(s) than a single flood. This interpretation is supported by the single F.U. sequence composed of Sf units where individual units are separated by desiccation cracks. These sequences are the only small-scale sequences in the Collegats Group to which the author is able to assign a temporal scale with any degree of certainty.

The C.U./F.U. megasequences are interpreted in two ways. The first C.U. sequence develops in locally sourced material and is believed to represent the progradation of a local fan. The overlying F.U. sequence has dual allochthonous and local sources (allochthonous pebble content is 4-16%, lower on average than the transfer system seen in the Graus conglomerate (4.4.2(a))). The F.U. could be the result of the decay of either or both sources. The uppermost C.U. sequence is interpreted as the result of autocyclicity within the transfer system.

The upward change in palaeosol types is interpreted as the result of increasingly better drainage. The gley soils indicate prolonged saturation of the soil profile - the Fe-encrusted burrows may be indicative of permanent saturation. The overlying Ustalfs indicate seasonal wetting of the soil while the even coloured soils at the top of the sequences indicate well-drained soils. Better drainage is likely to be the result of tectonic uplift raising and tilting local base levels. However, the opposite of this is seen at the base of the sequence where gley palaeosols drape relief caused by the phase of foreland directed thrusting. It is possible that the initial gley palaeosols, and the change upwards to better drained soils may be controlled by base levels within the Ebro basin. No such palaeosols were recognised in the two time equivalent sections (Fig.4.36). This observation is probably best explained by better drainage conditions in these areas: sediments are coarser overall, deposited in more proximal positions on steeper slopes.

4.4.4. Southern Margin of the External Sierras

4.4.4(a) Introduction

Although undated the sediments N.E. of Estadilla [7260], at the southern margin of External Sierras, are lithostratigraphically comparable with sequences in the Collegats Group, and closely linked with the development of the External Sierras. So they are considered to be Oligo-Miocene in age.

The sediments are divided by unconformities into three units (Fig.4.54): a basal sequence which unconformably (E_1) overlies Cretaceous to Eocene rocks and is composed of locally sourced conglomerates; a varied sequence of fluvial, lacustrine and evaporitic sediments, separated from the underlying conglomerates by a cliff-like unconformity (E_2); and a number of scattered conglomerate filled tributary palaeovalleys whose bases define unconformity E_3.

A detailed analysis of the sediments and a careful consideration of the palaeogeography, suggests that the thick conglomerate sequence is a partial time equivalent of the sections described at the northern margin f the major backfold, and that the overlying sequences post-date the Collegats Group (Fig.4.36).

The sequences are important (i) in enhancing an understanding of the palaeogeography of the area, and (ii) in providing examples of braided streams, syntectonic resedimented gypsum alluvial fan sedimentation, and tributary valleys.

4.4.4(b)Description of stratigraphy

(i)E_1-E_2

E_1 is a low angle unconformity. A log extending from [733625] to [735622] is representative of the majority of the sediment between E_1 and E_2 and is discussed in detail below {4.4.4(c)}. This sequence is underlain by a lenticular lithosome composed of stacked sediment gravity flows [7462; 7461]. Clasts within the lenticular lithosome are locally sourced, palaeocurrents are to the southwest or west.

(ii) E_2-E_3

Unconformity E_2 is complex and irregular, preserving for example 20m of relief at [725624] in the form of a N.W.-S.E. trending, S.W.-facing cliff line. The overlying sediments can be considered in four formations denoted (a)-(d) which comprise the Estada Group (Fig.4.54).

(a) Formation (a) is composed of sheet sandstones in a siltstone background. The sandstones are trough cross-bedded, coarse grained and friable. Both the siltstones and the sandstones are strongly colour mottled.

The formation is interpreted as fluvial in origin with the sandstones being deposited by laterally unstable channels and the siltstones representing overbank sediments.

(b) Where best exposed, [7262], formation (b) is composed of evenly coloured grey-brown marls and siltstones. The marls and siltstones are punctuated by cm-thick, fine-medium grained sandstone beds with sharp bases and occasional matrix-supported, outsize quartzite and limestone clasts (Sde). Rare symmetric wave ripples, desiccation cracks, and scattered ichnofauna (1.5mm diameter sub-vertical tubes and 1-2cm wide grazing trails) occur. Along the C-1311 road to Graus the formation is characterised by stacked sequences of 2-10cm thick erosively based normally graded (Sde) or ungraded (Sd) sandstone sheets, interbedded with grey silts and clays (Fig.4.55(a)).

The short wavelength wave ripples, sheet sediment bodies and normally graded sediment gravity flows (Nemec and Steel 1984) are indicative of deposition in a shallow lake. The sediment gravity flows are interpreted as the deposits of a small fan delta active in the N.E. part of the lake. The sun-cracks indicate that the lake periodically dried out.

(c) Formation (c) consists of well-bedded grey gypsum interbedded with marl and plane laminated siltstone. Gypsum bedding is defined by late white anhydrite veins and primary siltstone layers and characterised by an interlocked polygonal pattern with teepee structures at polygon margins.

The facies association is interpreted as the result of periodic desiccation of a shallow saline lake.

(d) Formation (d) is composed of cross-bedded and rippled sandstones and minor siltstones. The formation is intimately related to the evolution of the Estadilla diapir and is discussed in detail above {3.7}.

4.4.4(b)(iii) <u>E_3 the Carollas Palaeovalley system</u>

Unconformity E_3 is likely to have a composite origin: it describes the base of the ten or so discrete conglomerate bodies which unconformably overlie units E_1–E_2 and E_2–F_3. The conglomerate bodies are elongate E.-W. (100-300m long and 50-100m wide) and have fills derived from local sediment sources. Four such bodies which occur in close proximity along the Barranco Carollas E.N.E. of Estada have been studied in detail [730625].

<u>Geometry</u>

The four sediment bodies are exposed along an E.-W. modern day valley and are numbered (1)-(4) on Figure 4.56(a). (1) is composed of debris flows and gravel scour-fills (Gsc) with palaeocurrents indicating flow to the west; (2) is dominated by sediment gravity flows with palaeocurrents to the west and is described in detail below; (3) is composed of sediment gravity flows and pebbly sand scour-fills (Ss) with palaeocurrents to the W.S.W.; and (4) is coarse grained and composed of stacked sediment gravity flows.

Sediment bodies (1) (2) and (4) are interpreted as representing remnants of a west-draining palaeovalley system. Sediment body (3) is also interpreted as a palaeovalley fill. It may either represent an earlier palaeovalley system or a tributary to the west-draining system (the anomalous palaeocurrents may be explained by reflection of flows at the prominent E.-W. margin) - the relationship is not clear in the field. Since sediment body

(4) is far coarser than body (3) it is unlikely that (4) is a downstream equivalent of (3).

Sediment body(2)

The facies within this sediment body are dominated by sediment gravity flows (58%; Fig.4.56(c)). There is a high positive correlation between maximum particle size (MPS) and bed thickness (Bth) both for all measured beds (0.796, significant at better than the 0.1% level) and for sediment gravity facies alone (0.811, significant at the 0.2% level). Both data sets produce regression lines with positive slopes and positive intercepts on the Y-axis. Palaeocurrents are directed towards the west in marked contrast with the underlying sequence.

The basal erosion surface (E_3) is either stepped across competent beds in the underlying sequence, or gently slopes at an angle of about 10° to the south. In addition, a stepped erosion surface with a relief of over 4m occurs within the sediment body (Fig.4.56(c)&(d)). There is no angular discordance between beds deposited above and below this internal erosion surface.

Interpretation

The observed facies assemblage is directly comparable to the deposits of the proximal part of a small alluvial fan {4.3.2(b)}. The statistical analysis showing a high positive correlation between MPS and Bth supports the interpretation that the beds are dominated by sediment gravity flows, as they are in accordance with the model presented in section 4.2.1(d). Positive intercepts on the Y-axis indicate that the sediment gravity flows were cohesive.

However, the geometry and palaeocurrents of the sediment body suggest that it formed the northern margin of a west draining paleovalley. The lack of an angular discordance between the two stages of the fill to the palaeovalley indicates that the internal stepped erosion surface is not tectonic in origin. It is interpreted as the result of a "complex response" (Schumm 1977; {4.3.4(b)}). Initial uplift caused incision producing the basal unconformity surface (E_3). As incision worked its way in to the catchment area, sediment production increased to a level greater than "river" competence. So the valley aggraded and became choked. But, the river profile remained above its equilibrium level so that incision recommenced producing the internal unconformity.

4.4.4(c) E_1-E_2 - Barranco Bardanella Log

The section is 285m long and commences 20m above unconformity E_1. The sequence is composed of more or less stacked conglomerate and (medium to coarse grained) sandstone bodies which punctuate a fine sand background. Palaeocurrents are dominantly to the east. As the exposures are parallel to flow, sediment body architecture cannot be constrained. Clast lithologies vary: allochthonous clasts form $<1\%$ to 15% of the pebble sized clasts in the basal part of the section, but are absent at the top of the sequence (Fig.4.57). Traction current facies are dominant within the sediment bodies. Distinct sediment bodies have markedly erosive bases, are 1-4m thick and are characterised by small-scale F.U. sequences. The entire log fines upwards from its base to around 100m: conglomerates decrease from 57% in the basal part of the section to 20% in the top. But, there are no marked changes in conglomerate facies associated with this change.

Interpretation

On the grounds of coarse grain-size, erosive bases, F.U. motif and traction current facies the small-scale sequences are interpreted as channel fills deposited by bedload rivers during decreasing discharge. The absence of facies indicative of bar relief (Gp; Gl; Gf) suggests that the channel was not braided {4.3.1(b)(ii)}.

The coarse sediment facies assemblage and facies sequences are comparable to the "Donjek" model of Miall (1978) which he interprets as the product of distal, gravel rivers. The sequence described here differs in that the coarser sediment bodies are interbedded with thick fine sandstone units - a feature which may reflect high subsidence rates (Mclean and Jerzykiewicz 1978; Hayward 1983).

Clast provenance studies indicate that the sediments were derived in part from a distant source in the lower part of the section, but entirely from local sources in the top part of the section. The east directed palaeocurrents indicate that the local source area was to the west, and that the rivers hugged the fold generated topography of the External Sierras.

Conclusions - Southern Margin of the External Sierras

The facies associations of the Baranco Bardanella section which indicate bedload rivers and the dominance of channel fill sedimentation, are closely comparable to the depositional style of the "transfer system" conglomerate within the Collegats Group. The section is also intimately related to the development of the External Sierras as it is under- and overlain by unconformities considered to be directly related to their growth. So although undated it is considered to be part of the Collegats Group. It differs from the Collegats Group transfer system in being dominated by locally sourced clasts, exclusively so in the top part of the section. The scale of the sequence (305m in total) suggests major local sediment input, two sources are suggested: (i) fans derived from the dip slope of the External Sierras and (ii) the major west draining stage of the locally sourced Olvena palaeovalley {4.4.3(c)(iii)} which, when the western end of the Olvena paleovalley was breached, would have resulted in a sudden increase in the percentage of locally sourced clasts in the system hugging the southern margin of the External Sierras.

The overlying Estada Group has facies associations and facies sequences unlike any sequences in the Collegats Group indicating that it postdates the Collegats Group. However, it is closely associated with the develpment of the External Sierras, as it is folded by them and unconformably overlain by the Carollas palaeovalley system. The Estada Group is most important in showing the syn-sedimentary growth of a small diapir and the resedimentation of gypsum derived from the diapir as two small alluvial fans {3.7}.

The Carollas palaeovalley system is also related to the development of the External Sierras. But in post-dating the Estada Group it is even less certain that is a partial time equivalent of the Collegats Group.

4.4.4(d) Summary of the evolution of the External Sierras in the Olvena area

The evolution of the Olvena area during deposition of the Collegats Group is summarised on three block diagrams (Fig.4.44). The first stage is characterised by a west-draining palaeovalley at Olvena defined by two folds in the hangingwall of a major thrust sheet (thrust sheet 2; {2.6.3(c)}): palaeovalley sediments are derived from local sources on the fold flanks. A time equivalent south-flowing transfer system, preserved in the Embalse log {4.4.2(d)}, is thought to have turned westwards on encountering these folds and flowed southwards into the Ebro basin via a structural low. The low corresponded to a strike-slip transfer zone separating the backfold at Olvena from the recumbent fold at Urbiego. Prior to the emergence of the External Sierras this transfer system is thought and to have flowed directly southwards.

The second block diagram shows the palaeogeography after the main phase of folding, during which time a number of structural features developed: (i) the fold at the southern margin of the Olvena palaeovalley developed into a major backfold; (ii) the palaeovalley syncline was tightened and a major unconformity, the Olvena unconformity, developed within it; (iii) a major backthrust developed to the north of the two early folds deforming the Embalse conglomerate and defining a second syncline in its hangingwall and (iv) the western end of the Olvena palaeovalley became closed. These structural features changed sediment dispersal patterns in three key ways: (a) reversal of flow in the western part of the Olvena paleovalley, so that the percentage of allochthonous clasts in the dispersal system which hugged the External Sierras was suddenly reduced; (b) by constraining the transfer system to flow around the end of the backthrust; and (c) by initiating numerous local alluvial fans.

The third block diagram shows how the system evolved as the area aggraded. The anticline at the northern margin of the Olvena palaeovalley was progressively overtopped by the transfer system, and by the valley-fill sediments. So west-directed flow in the palaeovalley was re-established and sediments there increasingly took on the compositional character of the transfer system. Eventually the combined effects of relief and local sediment input were overcome and the transfer system swamped the Olvena palaeovalley to onlap the major backfold.

The further evolution of the area is recorded solely to the south of the major backfold, where the Estada Group unconformably overlies a sequence characteristic of the Collegats Group. The Estada Group, and the Carollas palaeovalley system which unconformably overlies it, are both folded by the External Sierras. The former records fluvial, lacustrine and playa sedimentation as well as syn-sedimentary diapirism which induced alluvial fan sedimentation; the latter is an example of a tributary palaeovalley.

4.4.5. Juseau

(a) Introduction

Juseau [8365] lies at the northern margin of the External Sierras. The margin is marked by the Aguinaliu diapir, essentially the core of the large backfold of Figure 4.37. {4.4.3(b)}. Units within the Collegats Group thin towards the External Sierras suggesting syn-tectonic sedimentation (Map 2). Uplift is recorded in the Collegats Group as source, facies and palaeocurrent changes which define six distinct depositional systems: (a) an early, locally-sourced conglomerate fan, backshed to the north and overlain by (b) a sandy terminal river system flowing parallel to the backthrust front, and cut into by (c) a palaeovalley formed by river capture, which flowed south across the External Sierras, and was terminated by a second uplift phase which produced (d) a second backshed fan, in turn overlain by (e) a second terminal river system. A final depositional system is characterised by (f) a change in the drainage direction of the rivers supplying sediment to the terminal fans.

4.4.5(b) Basal fan

Introduction

A conglomerate body unconformably overlies overturned Cretaceous limestones to the east of Juseau [839635]. The western margin of the conglomerate is defined by a strike-perpendicular fault; to the east the conglomerate onlaps the Cretaceous and is lost through poor exposure. Beds within the lowest part of the conglomerate body have a wedge-shaped geometry, thickening away from the unconformity and decreasing in dip upwards (Fig 4.58). The geometry is interpreted as a cumulative wedge system (Riba 1976a,b; Anadon *et al.* 1986) produced by uplift during sedimentation. A 41m thick section has been logged through the conglomerate from the basal unconformity to the overlying sediments (Fig.4.59).

Description

The sequence is composed of sediment gravity flows (Gdp 32%; Gs 11%; Gdc 7%; Gdm 2%) and gravel filled scours (Gsc 45%). There are no small scale facies sequences: the whole log fines upwards slightly, with gravel-filed scours (Gsc) facies becoming increasingly important upwards. The clasts are sub-angular limestones derived from the Alveolina Limestone and Tremp Formations and the Cretaceous (no Triassic clasts). Sorting is generally poor, especially in debris flow units, where matrix content is variable and some clasts are matrix supported (Fig.4.4(a)(i)). Statistical analysis shows a positive correlation between maximum particle size (MPS) and bed thickness (Bth) both for all determined beds (0.597 significant at the 0.1% level) and for sediment gravity flow facies only (0.871 significant at the 0.1% level). Regression curves for both these data sets have positive intercepts on the Y-axis. Palaeocurrents fan clockwise from N.W. to N.E. directed.

The top of the sequence is defined by a marked change in clast size, provenance and facies (Fig.4.59).

Interpretation

The mixed sediment-gravity-flow and stream-flow facies association is compatible with deposition on the proximal part of a small alluvial fan {4.3.2}. The north-directed palaeocurrents and clast composition suggest that the fan was sourced from the northern limb of the Aguinaliu diapir but that Triassic rocks had not reached the surface in the catchment area at that time. Following the trend of the modern day diapir the catchment area for the fan can be estimated at under 1km². The position of the fan may have been controlled by the small strike-perpendicular basin margin fault. The radial palaeocurrents are thought to reflect the single input point and a true fan geometry is thought to have developed.

The sudden termination of fan sedimentation is considered to reflect a tectonically induced palaeogeographic change. A change in the source area seems to be the most likely cause - blocking off or diverting the fan sediment supply, possibly towards Juseau itself where a similar locally sourced fan body occurs [833637].

Sediment gravity flow units conform closely to the model of Nemec and Steel (1984; {4.2.1(c)}) confirming the field facies designations, and allowing them to be advanced since the positive intercept of the regression curve on the Y-axis implies that the flows were cohesive debris flows. This is in accordance with locally high matrix contents.

4.4.5(c) Sandy terminal river system 41-201m

Description

The sequence overlying the basal fan differs from it primarily in terms of provenance: it is dominated by allochthonous clasts - 28-56% of pebble sized clasts can be related to lithologies which outcrop in the Axial and Nogueras Zones (Table 4.7). Three beds prove the exception to this rule and are composed of angular clasts of locally derived limestone (Table 4.7).

The sequence is composed of silts punctuated by sandy, sheet sediment bodies dominated by three facies: sandy sheetfloods (Sf); pebbly sand scour-fills (Ss) and gravel-filled scours (Gsc). Sf units may occur individually or they may form multistorey bodies of constant grain size up to 3m thick. Other sediment bodies F.U. Such sequences are thin (0.6-1.6m) and invariably based by Gsc passing upwards either gradationally into Ss, or sharply into a sequence of Sf units. In addition, two medium scale F.U. sequences occur. A notable exception is the single palaeocurrent obtained from one of the three locally sourced beds which indicates flow to the N.N.E (191m on Figure 4.59). These beds are characterised by Gsc and trough cross-bedded gravel (Gt) facies, and occur randomly in the sequence.

Interpretation

The sandy sediment bodies hold the key to the interpretation of this sequence. On the grounds of coarse grain-size, erosive sequence base and F.U. motif the gravel scour-fill (Gsc) based F.U. sequences are interpreted as channel fills. Sequences which record the transition Gsc-sandy scour fills (Ss) indicate a gradual waninng flow, while sequences which are characterised by the transition Gsc to sandy sheetfloods (Sf) indicate episodic flow, decreasing in competence through time. The individual Sf units and superposed Sf units are interpreted as the downstream equivalents of these channels. These downstream deposits may also have been channelised. Their vertical sequences are closely comparable to those seen in 80m wide channels on a modern terminal fan (Parkash et al. 1983; {4.3.3}). Unfortunately exposure does not allow rigourous architectural studies to check for channels on this scale, but channel margins and marked erosive bases were not seen in these sequences. So they are tentatively interpreted as channel mouth lobes.

The clasts derived from the Axial and Nogueras Zones indicate that the channels must link back to a major transfer system, sourced in the north.

The dominant west-directed palaeocurrents are believed to reflect the relief of the External Sierras to the south, and the existence of a "through point" near Olvena to the west via which the main transfer systems drained to the Ebro Basin, so that the regional palaeoslope was to the west.

The three locally sourced beds are interpreted as recording the periodic activity of a local backshed fan. The rarity of the beds suggests that they reflect the breaching of a geomorphic threshold by an exceptional climatic event {4.3.2(c)}. The beds may be the earliest events of a second backshed "fan" ((e) below) the lowest beds of which dip steeply and strike off into this sequence (Fig.4.60).

4.4.5(d) Palaeovalley

Description

The third depositional system is characterised by stacked pebble-cobble conglomerates. The top of the conglomerate body is planar and capped by the second backshed "fan". But the base of the body descends to the east and is interpreted to be erosional into the underlying terminal fan sequence.

The conglomerates are composed of massive gravel beds (Gm) and gravel filled scours (Gsc) with scattered sand lenses (Fig.4.6). Both facies have strongly developed imbrication (a(t)b(i)) suggesting palaeoflow to the S.E. - consistently towards or obliquely across the External Sierras which trend 300°-120°. Laterally extensive major erosional surfaces, and small scale F.U. sequences, are absent.

The conglomerates are either bimodal (pebble-cobble and coarse sand) or poorly sorted. The clasts are very well rounded and have a strong Axial-Nogueras Zone signature (Table 4.7).

Interpretation and Discussion

The palaeocurrents and the erosive western margin of the conglomerate body are interpreted together as indicating flow across the External Sierras in a palaeovalley. The valley is intepreted as the product of river capture during a period of quiesence between the development of the two backshed "fans".

The palaeocurrents indicate flow on the same trend as the Mas de Nicolau palaeovalley {4.3.6}: aggressive headward erosion of this palaeovalley may have breached the External Sierras at Juseau to link with part of the main fluvial tranfer zone in an act of river capture. However, there are only small amounts of allochthonous material in the Mas de Nicolau palaeovalley; it occurs within the first part of the valley fill which is characterised by a "non-aggressive", low-energy, axial system. So the two outcrop areas cannot be linked directly. Instead river capture is considered to be the product of a largely non-depositional, actively downcutting system which operated as a first stage in the history of the Mas de Nicolau palaeovalley.

The traction current facies and coarse nature of the deposits suggests that the valley fill was deposited by bedload rivers {4.3.1(a)}. However, the absence of marked erosive bases and small-scale F.U. sequences, interpreted elsewhere as channel bases and channel fills respectively, is in conflict with this.

The absence of well developed channels is thought to reflect rapid aggradation, perhaps caused by the raising of base level across the External Sierras at the onset of the second uplift phase. For much of its history the palaeovalley was probably erosional, or non-depositional, debouching sediment into the Ebro basin at the southern end of the Mas de Nicolau palaeovalley.

4.4.5(e) Second backshed fan

Description

The fourth depositional system is a conglomerate body which forms the ridge to the west of Juseau and is capped by La Ermita ([829638]; Fig.4.66). The main body of the conglomerate caps the

Juseau palaeovalley ((d) above) but beds at the base of the conglomerate diverge downdip to interface with the first terminal fan system ((c) above; Fig.4.60). The upper part of the conglomerate body is a lateral equivalent to and overlain by, the second terminal fan sequence ((f) below).

A log through the conglomerate, below and to the south of the hermitage, shows it to be dominated by gravel-filled scours (Gsc 41%; Gt 3%) and sediment gravity flows which commonly fill pre-existing scours (Gdp 19%; Gs 11%; Gdm 5%; Sd 2%; Fig.4.61). There is a low positive correlation between maximum particle size (MPS) and bed thickness (Bth) both for all measured beds (0.468 significant at the 0.1% level) and for sediment gravity flows only (0.473 significant at the 1% level).

There are no clear F.U./C.U. sequences within the conglomerate body. Clasts are dominantly poorly sorted angular to subrounded Cretaceous Limestone (>95%): some are derived from the Alveolina Limestone and Tremp Formations, and the Muschelkalk limestone but there are no clasts of ophite. Palaeocurrents spread anticlockwise from east to N.W. directed.

Interpretation

The facies assemblage is closely comparable to that of the basal backshed fan. It is interpreted similarly, as the proximal part of a small alluvial fan. Again palaeocurrents indicate a true fan geometry. However, the conglomerate body is over 1km wide at the basin margin, suggesting a sediment apron rather than a fan geometry.

The gravel-filled scour facies (Gsc) and the trough cross-bedded gravel facies (Gt) indicate that the fan surface was dissected by shallow channels up to 1m deep.

The down-dip divergence of beds in the lower part of the apron is interpreted as a cumulative wedge system (Riba 1976 a,b; Anadon *et al.* 1986) indicating syntectonic sedimentation. The lower beds indicate that the sediment apron was a long lived feature which predated the palaeovalley, but was only periodically active probably during exceptional hydrographic events. The uplift believed to have caused the rapid choking of the palaeovalley is interpreted to have also rejuvenated the apron source area, initiating fan sedimentation. The dominance of Cretaceous clasts suggests that the source area had undergone some unroofing since backshed fan 1 times, but in view of the absence of Triassic clasts, that the Triassic at the core of the Aguinaliu diapir had still not reached the surface in the catchment area.

4.4.5(f) Second terminal river system

The fifth depositional system lies above and is a lateral equivalent to, the top of the backshed apron ((e) above). The logged section commences at ([834645]; Fig.4.62).

The section is dominated by silts punctuated by sheet sediment bodies. Sediment bodies are either simple or composite. Simple bodies are composed of single sandy sheetfloods (Sf). Composite sediment bodies are 1-2m thick, they may F.U. or be of constant grain size. F.U. sequences can commence with gravel-filled scours (Gsc) or pebbly sand-filled scours (Ss), but these constitute only the lower part of the sequence and are overlain by sharp-based (Sf) beds. Other F.U. sequences and all sediment bodies of constant grain size are composed solely of Sf beds. A pebble count revealed that 49% of the clasts were derived from lithologies outcropping in the Axial and Nogueras Zones. Palaeocurrents indicate flow to the E.N.E. - thin section work on plane laminated facies consistently confirms the current ripple direction indicators (for techniques see Bouma (1979)).

Interpretation

The facies, facies sequences and architecture of the coarse sediment bodies are closely comparable to those observed above the first backshed fan. They are interpreted similarly, as the products of a terminal fan. The main differences are threefold: firstly, the decreased number of clearly channelised facies (erosively-based, F.U. sequences with Gsc-Ss-Sf transitions); secondly, the F.U. sequences based by Gs and Ss facies are all sharply overlain by Sf facies, indicating that none of the channels were characterised by steady flow, and, thirdly, the increased proportion of background silt. The differences are interpreted to represent a more distal setting.

The E.N.E. directed palaeocurrents indicate flow obliquely away from the External Sierras. Clast compositions indicate that the terminal fans were linked upstream to a major transfer system - probably the transfer system seen to the N.W. at Graus.

4.4.5(g) Axial System

A 30m thick section located at [845657] characterises a sixth depositional system (Fig.4.63). The section is stratigraphically above the second terminal fan ((f) above) and 1.5km further from the External Sierras. In addition, it is considered to be statigraphically above the Castellarnes log {4.4.2(b)}.

Description

The observed facies divide the section into two parts: a lower part dominated by the sandy sheetflood facies (Sf), with architecture and sequence closely comparable to that seen in the second terminal fan (e) above, and an upper part dominated by gravel-filled scours (Gsc). The Gsc facies occurs in stacked sequences of constant grain size and in three small-scale sequences: two of which are based by important erosion surfaces and fine upwards, and one of which coarsens upwards. Palaeocurrents from the Gsc facies show a spread of 110° but are essentially directed to the west. Unfortunately, no palaeocurrents could be obtained from the lower part of the section. Pebble counts show a high percentage of allochthonous clasts (Table 4.3).

Interpretation

On the grounds of coarse grain-size, erosive sequence bases, and traction current facies, the F.U. and stacked sequences of gravel scour-fills (Gsc) in the upper part of the section, are interpreted as channel fills deposited by bedload streams. The allochthonous clasts indicate that the channels were linked to a major transfer system sourced in the Axial and Nogueras Zones. The west-directed palaeocurrents imply a reversal of drainage directions with respect to the E.N.E. directed terminal fan sequence below ((f) above). No obvious erosional surface marks this change (although exposure is not good enough to preclude this possibility). So the change is considered to have occurred gradually.

4.4.6. Mas de Nicolau Palaeovalley

4.4.6(a) Introduction: Structure and Stratigraphy

The Mas de Nicolau palaeovalley is isolated from the main outcrop of the Collegats Group. It lies in a complex syncline within the External Sierras [8460 to 8753]. The palaeovalley is important for two main reasons: (i) it provides an example of a palaeovalley with tectonicaly active margins; and (ii) marginal alluvial fans within the valley fill show how the lithology of the source area can control fan facies.

The valley fill is believed to be Oligo-Miocene in age on two grounds: firstly intimate relations can be demonstrated between the fill of the basin and the folds of the External Sierras, considered elsewhere as Oligo-Miocene in age {4.4.3; 4.4.5.} and, secondly, lithostatigraphy and facies patterns are closely comparable to sequences observed elsewhere in the Collegats Group.

An unconformity south of Mas de Nicolau [8657] can be used to divide the basin fill into two parts: a northern stratigraphically older N.W.-S.F. trending part and a southern stratigraphically younger N.-S. trending part.

The trend of the northern part of the basin fill parallels local fold axes. The sediments thicken towards synclinal axes [8559] and onlap the limbs of the syncline [8557] suggesting that the N.W.-S.E. folds delineated the early basin geometry.

By contrast, the southern part of the basin fill trends across the N.W.-S.E. folds. Its outcrop lies to the east of a small diapir near Minas de Santa Teresa [8655] and parallel to a large diapir and faults related to it [8856]. The margins of the second phase of the basin fill are relatively poorly exposed. However, unconformities can be inferred at both margins of the palaeovalley: between conglomerates folded around the Minas de Santa Teresa diapir, and younger fluvial sediments at [868557] and between overturned conglomerates, and younger fluvial sediments at [876557]. Above the marginal unconformities the fill dips shallowly to the S.W. except at Gabasa where a structurally complex zone causes sedimentological continuity to be lost. The unconformities and the geometry of the fill suggest that the diapirs delimited the basin during the second phase of sedimentation.

Using the results of a detailed sedimentological study the suggested basin evolution is tested, confirmed and refined below.

4.4.6(b) The Northern part of the Basin

Four sedimentological sections have been measured in the northern part of the basin: they are described and interpreted individually below. Their implications for basin evolution are then summarised.

4.4.6(b)(i) Basal Conglomerate

Description

The oldest part of the basin fill is a conglomerate composed of angular to subrounded clasts of Alveolina Limestone and Tremp Formation limestones. A 19m thick section through these conglomerates has been measured at [850595] (Fig.4.64). The conglomerates are tightly packed, clast supported and dominated by gravel filled scours (Gsc facies 49%). Other facies are massive gravels (Gm 12%), gravel sheetfloods (Gs 15%) and massive sandstone (S 24%). There is a high positive correlation between maximum particle size (MPS) and bed thickness (Bth) both for all determined data (0.760 significant at the 0.1% level) and for sediment gravity flow faces only (0.928 significant at the 0.1% level; Fig.4.65). Both regression curves have positive intercepts on the Y-axis. No small-scale fining upwards sequences are observed. The log, however, fines upwards in its top 10m. Palaeocurrents are to the S.E. parallel to the anticlines which bound the outcrop. The conglomerates onlap the limbs of the anticlines and are folded with them.

Interpretation

The dominance of coarse angular clasts, and the assemblage of sediment gravity flow and stream-flow facies indicate that the sequence was deposited in a proximal setting on a small alluvial fan {4.3.2(b)}. The palaeocurrents, the observed onlap relations and the folded nature of the conglomerate show that the fluvial system was confined to flow S.E. within a S.E. plunging syncline which developed before (probably during) and after deposition. The conglomerate sequence is believed to be the result of the folding which produced uplift and created local source areas.

The high positive correlations between MPS and Bth imply that the deposits are in close accordance with the model for sediment gravity flows {4.2.1(d)}. Since the regression curves make a positive intercept on the Y-axis, the model implies that the beds are the product of cohesive debris flows. This is problematic for two reasons: firstly, because of the wide range of observational facies, each facies would be expected to form <u>its own</u> regression line <u>if</u> it agreed with the model, and, secondly, because the majority of the sediments are traction current deposits which should not be in accordance with the model. It is believed that the high positive correlations cannot be interpreted in terms of the model i.e. they are spurious.

4.4.6(b)(ii) <u>Gabbro Fan</u>

<u>Description</u>

At the northern margin of the first part of the basin fill a thick sequence of red clays, silts, sandstones, and conglomerates occurs stratigraphically above the basal conglomerate. A section through the sequence ([855597]; Fig.4.66) shows two distinct sediment sources: firstly, a dominant "local" source, characterised by poorly sorted clasts of angular limestones (Triassic, Cretaceous, Tremp Formation and Alveolina Limestone Formation in age) and rounded gabbro which lie in a gabbroic matrix; and, secondly, an "axial source" composed of well sorted medium to coarse-grained sandstones.

Beds which have clasts indicating a "local" provenance are dominated by 5-60cm thick sediment gravity flows. Matrix supported units are common but they often have erosive bases and imbrication indicative of traction currents a(t)b(i), so that the facies codes outlined above {4.2.1} are difficult to apply (Fig.4.66 & 4.67(a)). There is a positive correlation between MPS and Bth both for all data (0.54, significant at the 0.1% level) and for sediment gravity flows only (0.565, significant at the 0.5% level; Fig.4.68). Palaeocurrents are south-directed - towards the E. and S. at the base and towards the W. and S. in the middle of the sequence.

The rare well sorted sandstone beds are characterised by the sandy sheetflood (Sf) facies.

Small-scale facies sequences are not well defined, but three intermediate scale sequence do occur: one F.U. and two C.U. sequences. The background sediments are usually composed of extensive mm-scale laminations which may be planar, or show compressional structures; extensional crack-fills occur but are rare (Fig.4.67(b)). Lozenge shaped crystals (2-4mm long) of gypsum and exterior moulds after gypsum are common. The laminations are of uncertain origin, but could relate to gypsum rich and gypsum poor layers. The background sediments are evenly coloured mostly browns, yellow-grey or olive-brown. Incipient entisols occur locally, soil facies $S_{1,2,4}$.

Intepretation

The coarse nature, and mixed facies association of sediment-gravity-flow and traction-current facies, are compatible with deposition on the proximal part of a small alluvial fan (4.3.2(b)). Radial palaeocurrents indicate a true fan geometry. The erosive base and F.U. motif of the intermediate scale F.U. sequence indicates that it is a channel-fill sequence. However, three features indicate that the fan surface was not markedly channelised: firstly, the intermediate scale F.U. sequence is a unique example; secondly, there are no small-scale F.U. sequences interpreted elsewhere as channel-fills, and thirdly, facies are dominated by sediment gravity flows which do not produce channels. The intermediate scale C.U. sequences are interpreted as the result of short-term fan, or supra-fan lobe progradation

The compressional structures in the background sediments are intepreted as the result of gypsum growth in a playa environment which fringed the small alluvial fan. The crack fills are interpreted as desiccation cracks indicating periodic drying out of the playa.

The statistical analysis is not in accordance with the facies interpretations. Although significant, the correlation coefficients of MPS-Bth plots are lower than those expected for sediment gravity flows (Steel 1974; Nemec and Steel 1984). Yet the field evidence points unequivocally to a sediment gravity flow origin for many beds, suggesting, either that the model is incorrect, or that other factors operated to bias the data set away from a good correlation. A number of possible factors are discussed above {4.2.1(c)}. In this example it seems likely that the variety of sediment gravity flows would serve to lower correlation coefficients - different processes would produce distinct regression curves. Indeed, the poorly sorted debris flow facies (Gdp) plots consistently above the gravel sheetflood facies (Gs; Fig 4.68(b)).

Both the playa facies and the dominance of matrix supported sediment gravity flows are believed to be a direct reflection of the source area. The playa facies only occurs at the northern margin of the basin in association with material derived from a Triassic source. It is likely that gypsiferous groundwater and stream water and an abundance of clay (derived from the weathering of the gabbro) were associated with the Triassic source, promoting authigenic gypsum growth in the fine grained material. Given the possibility of high rates of gypsum supply the climate need not have been particularly dry (c.f. Truc 1978). Similarly, matrix supported sediment gravity flows are confined to the northern margin of the basin - they are believed to be the result of supply of gabbroic sand.

The even brown colouration and rare, poorly developed entisols ($S_{1,2,4}$) indicate that the basin was relatively well drained and that a seasonal climate operated.

The catchment area is difficult to estimate. The present structure shows only a small area of Triassic rocks immediately adjacent to the fan ($0.5km^2$). However, the Triassic outcrop area increases to the south-east indicating that the carapace to the Triassic would have plunged to the N.W. If this carapace was pierced or eroded along the crest line of the fold the inherited slope of the Triassic would have been towards the northwest, towards the head of the gabbro fan. Therefore, the fan catchment area may have been several km^2.

4.4.6(b)(iii) The Axial system - Mas de Nicolau log

Description

The section is located towards the top of the first part of the basin fill, nearer to the basin axis than the two sections described above ([865584]; Fig.4.69). The sequence is dominated by fine sands and silts. Interbedded with these are units of medium grained sand and coarser grade material, characterised by sandy sheetflood facies (Sf 36%) and a mixture of other sediment gravity flow (Gs 7%, Sd 4%, Gdc 5%, Gdm 1%) and traction current deposits (Sh 6%, S 12%, Ss 15%, Gsc 19%, Gm 2%). The facies are arranged in multi- or single-storey sheet sediment bodies. Seven small-scale F.U. sequences can be recognised (e.g.Fig.4.70). Each commences with gravel facies: gravel scour-fill (Gsc); massive gravel (Gm); or matrix supported debris flow facies (Gdm). The gravels pass upwards into sands: debris flow (Sde); sheetflood (Sf); plane-laminated (Sh); scour (Ss) and trough cross-bedded (St) sandstone facies. No medium or large scale sequences are present.

There is a high positive correlation between maximum particle size (MPS) and bed thickness (Bth) for both all determined data (0.744 significant at better than the 0.1% level) and for sediment gravity flows only (0.736 significant at the 0.5% level). The regression curves have positive intercepts on the Y-axis.

The sediments most commonly have an even, brown colouration (5Y 7/2 - 5YR 4/4). Weak entisols, soil facies S_1 and S_4, are developed at some horizons. In addition, lozenge shaped gypsum crystals are present in parts of the background sequence. Although, in contrast to the northern fan, well laminated facies were not observed.

Pebbles are sub-rounded. Pebble counts indicate that the sequence is dominated by locally derived Cretaceous to Paleocene limestones: no Triassic clasts were recognised. The basal 60m of the sequence has a low percentage of allochthonous clasts (Table 4.7). Above 60m allocthonous pebbles are absent. Palaeocurrents indicate flow to the S.E. parallel to the local fold axes.

Interpretation

Palaeocurrents parallel to the structural trend indicate that the coarse sediment bodies were deposited by an axially flowing system. On the grounds of erosive bases, coarse grain-size and F.U. motif the small-scale sequences are interpreted as channel fills deposited by bedload rivers. The predominance of sandy sheetfloods (Sf) in the top part of these cycles suggests that flow was episodic. The isolated single storey sediment sheets are considered to be the downstream equivalent of these channels. The facies associations and sequences are compatible with deposition in a terminal fan setting. So the axial system is not thought to have been a through-going system: the southern end of the palaeovalley may either have been closed or had a very low gradient inducing rapid deposition.

The high positive correlations between MPS and Bth imply that the deposits are in accordance with the model for sediment gravity flows {4.2.1(d)}. The results are surprising in that a wider range of observational facies produces a higher correlation than the subset of sediment gravity flows. According to the model positive intercepts on the Y-axis can be interpreted as being the product of cohesive debris flows. This casts further doubt on the correlations since the majority of the facies are interpreted as the products of <u>fully turbulent</u> flows. In summary, the correlations are considered spurious.

The pebble counts indicate that the sediment was derived largely from local sources. The rare allochthonous clasts seen towards the base of the section are problematic, two explanations are apparent: either they represent the reworking of Collegats Group deposits which may have overlain the Paleocene prior to folding in the External Sierras, or they indicate that the fluvial system was at some time linked to the main Collegats Group transfer system {4.4.5}.

In a similar way to the discussion in section 4.4.6(a)(ii) above the gypsum cystals and exterior moulds are interpreted as the result of groundwater or overland flow rich in gypsum.

The even colouration of the sediments and the weak entisols are indicative of well drained conditions in a seasonal climate.

4.4.6(b)(iv) The Southern Margin

Introduction

The section is located at the southern margin of the first part of the basin fill ([8557]; Fig.4.71). It is composed of conglomerates, sandstones and siltstones. Palaeocurrents, facies and clast composition allow two distinct depositional systems to be recognised: a marginal fan and an axial system. The situation is closely comparable to that seen in log (ii) above, which is a time equivalent sequence, Map 1 [8557,8558,8559].

The Marginal Fan

A number of conglomerate bodies occur at the southern margin of the basin separated by siltstones and occasional sandstones. At their thickest they are 2-10m, thinning towards the basin axis and along the basin margin. The thickest points are superposed in two localities, at [852576] and [857575] - the log passes through the first of these (Fig.4.72(a)). The conglomerates are composed of limestone clasts derived from the Cretaceous and the Alveolina Limestone and Tremp Formations, and relatively rare pale brown Tremp sandstones. There is no evidence for a Triassic source. Clasts are tightly packed, subrounded and well sorted - matrix is rare or absent (Fig.4.72(b)). The conglomerates are dominated by gravel filled scours Gsc (51%) and a range of sediment gravity flow facies (Gdc 10%; Gdm 6%; Gdp 4%; Gs 2%). The interbedded sandstone facies comprise the remainder of the coarse sediment bodies (S 13%; Sf 5%; St 4%; Sde 2%; Sd 2%; Ss 1%). There is a very low correlation between MPS and Bth, both for all data (0.273 not significant at the 10% level) and for sediment gravity flows only (0.384 not significant at the 10% level). Palaeocurrents within the conglomerates are directed towards the north but spread from west to east directed. Palaeocurrents in the sandstones are directed towards the east. Sequences within the conglomerates are retricted to two or three poorly developed small-scale F.U. sequences, and a slight medium scale C.U. sequence that constitutes the basal part of the thickest conglomerate body.

Interpretation

The superposed conglomerate bodies are interpreted as the product of a periodically active alluvial fan. Superposition of the thickest parts of three distinct conglomerate bodies suggests that the fan apex maintained a constant position through time: a position coincident with the axis of a small syncline in the basin margin, imlpying that the fan records an element of consequent drainage. The fan catchment area was probably small, 2-3km², being constrained to the south by a prominent E.-W. anticline and to the west by the outcrop of Triassic rocks north of Salinas [8356] which cannot have been in the catchment area. Thining and loss of the conglomerate within 1km of the input point suggests that the fan was somewhat smaller, 1km². The spread of palaeocurrents suggests that a true fan geometry developed. The fan is intepreted fundamentally as the product of uplift in the source area. Its periodic activity, as recorded by three sediment pulses is believed to reflect the breaking of geomorphic thresholds {4.3.2(c)}.

The sandstones and pebbly sandstone characterised by west-directed palaeocurrents are interpreted as part of the Axial system discussed above {4.4.6(b)(iii)}.

4.4.6(b)(v) Discussion and Summary

The three logs (ii)-(iv) are interpreted as the products of three more or less coeval depositional systems: northern and southern fans and an axial system.

The fans are inferred to have had similar catchment (2-3km²) and fan (1km²) areas. In addition, both are believed to be fundamentally tectonic in origin, while their coeval nature and close proximity suggests that they developed under the same climate. Yet they differ markedly in facies. The difference is attributable solely to the lithology of the rocks exposed in the catchment area of the respective fans: the northern catchment area producing a wide range of grain sizes and clast types, and pluvial waters rich in dissolved gypsum - sediment which favoured poorly-sorted, matrix-supported deposits and a playa-like fan fringe; and the southern catchment area producing a single clast type that favoured relatively well sorted deposits, and allowed only one matrix supported flow and no playa fringe.

In summary, the first part of the basin lies in a N.W.-S.E. trending syncline which tightened during sedimentation, and controlled sediment thickness and facies patterns. Uplift at the basin margins produced fans derived from local sources. The fans interfaced with a S.E. draining axial system characterised by terminal river sediments. There is no evidence for a through-going river system to the south at this stage. However, a low percentage of allochthonous clasts seen within the axial system does suggest that the palaeovalley was open to the north at some stage (4.4.5(d)).

4.4.6(c) The Southern part of the Basin

The fill of the southern part of the palaeovalley is orientated N.-S. indicating an important phase of tectonic uplift: uplift is reflected by marginal sediment sources seen as small alluvial fan bodies, and also by a marked change in the axial system which is entirely locally sourced and indicative of a through-going river in the southern half of the palaeovalley.

Two sections have been logged in detail in the southern part of the basin. They are discussed below, together with results of mapping and sketch logging.

4.4.6(c)(i) Marginal Conglomerates

The conglomerates which are folded by diapiric intrusion at the basin margin have not been logged. However, some useful observations have been recorded. The western conglomerate, at Minas de Santa Teresa, is composed of a 20m thick F.U. sequence. The conglomerates are composed of poorly sorted, angular or subrounded limestone clasts derived predominantly from the Alveolina Limestone Formation; rare clasts of Triassic limestone occur but gabbro clasts are absent. The beds are interpreted to have been deposited by sediment gravity flows: debris flows (Gd) and gravel sheetfloods (Gs). The eastern conglomerate is relatively poorly exposed. It has similar composition and facies.

Both marginal deposits are interpreted as the products of small locally sourced alluvial fans initiated by uplift at the basin margins. The exact stage of their formation has not been determined.

4.4.6(c)(ii) Mas de Nicolau Conglomerate - Axial system

A series of conglomerate ridges occurs to the S.E. of Mas de Nicolau, the lowest of which is described here [866580]. Each of the conglomerates thins to the S.S.W. where they appear to onlap the basin margin at [863572]. It is this relationship which suggests an unconformity separating the two parts of the basin fill.

Description

The conglomerate is dominated by gravel scour-fills (Gsc 65%;

Fig.4.73 (a)). Subordinate facies are dominantly traction current deposits (Gm 8%; Ss 4%; Sh 4%; St 2%; S 13%; Gs 2%; Gdm 1%). Small scale F.U. sequences (1-2m) are important, preserving facies successions indicating decreasing discharge. Clasts are subrounded, and derived from local limestone sources (Alveolina Limestone and Tremp Formations and the Cretaceous). Some clasts of pale brown Tremp sandstone also occur. There is a marked lack of matrix within the conglomerates. Palaeocurrents are directed to the S.S.E. although there is a spread of 80°.

Interpretation

On the grounds of coarse grain-size, erosive bases, F.U. motif and traction current facies the small-scale sequences are interpreted as channel fills deposited by bedload streams during decreasing discharge. Facies indicative of bar relief are absent so the channels are not thought to have been braided (4.3.1(b)(i)).

The axial palaeocurrents suggest that the channels represent a through-going, axial, fluvial system. The observed clast suite indicates that the system was sourced locally, and that there were no Triassic rocks exposed in the catchment area.

4.4.6(c)(iii) Gabasa road section

The section is composed of a number of erosively-based sediment bodies which have sheet geometries and punctuate a fine sand background (Fig.4.73(b)) The coarse sediment bodies form 39% of the section and are dominated by gravel-filled scours (Gsc 61%; Gt 10%; Gdc 1%; Ss 18%; Sf 10%). Each coarse sediment body is typically composed of one or two small-scale F.U. sequences which have facies successions indicating decreasing discharge. The clasts are derived from local limestone sources (Tremp and Alveolina Limestone Formations and the Cretaceous) and, rarely, from intrabasinal reworking of earlier locally sourced conglomerates. All clasts are well rounded, sorting is variable. Palaeocurrents are consistently directed to the south and southwest parallel to the basin margins. Statistical analysis shows a low positive correlation between MPS and Bth (0.566 significant at the 0.5% level) the regression curve has a positive intercept on the Y-axis.

Interpretation

The coarse sediment bodies are closely comparable to those seen in the Mas de Nicolau Conglomerate ((ii) above) and are interpreted similarly: as channel fills, deposited by an axial, non-braided bed-load river which constituted a through-going fluvial system deposited in a structurally defined palaeovalley. This system shows three key differences: firstly, the sediment bodies are not stacked, suggesting higher rates of subsidence or a change in sediment supply; secondly, the sediment bodies do not thin towards the basin margins suggesting that the basin margins were inactive during deposition, and thirdly, that the main palaeoflow direction has rotated slightly towards the west to maintain parallelism with the basin margins. The low positive correlation between MPS and Bth is consistent with the facies interpretations, being too low to apply the model for sediment gravity flows {4.2.1(d)} and closely comparable to values for similar deposits recorded by Steel (1974).

4.4.6(d) Conclusions and summary

In summary, the Mas de Nicolau palaeovalley is a tectonically defined and controlled palaeovalley. The palaeovalley lies within a composite syncline formed during two major stages of folding. The first stage controls the northern part of the palaeovalley and corresponds to the dominant N.W.-S.E. structural trend within the External Sierras. The folding is thought to correspond to hangingwall deformation during the emplacement of a major thrust sheet (thrust sheet 2; {2.6}). The second stage of folding, which controls the trend of the southern part of the palaeovalley fill, is the result of diapirism.

Both fold stages produced marginal sediment sources preserved as small-scale alluvial fans dominated by mixed sediment gravity flow and stream flow facies. In detail the lithology outcropping in the fan source area controls facies: fans with limestOne source areas are characterised by well-sorted clast-supported beds, while fans with source areas in Triassic gabbro and gypsum lithologies are characterised by matrix-supported beds and playa-like fan fringes.

An axial sediment dispersal system is maintained throughout. In the northern, early part of the palaeovalley, it is characterised by terminal fan facies sequences and is not considered to have been a through-flowing system. In the southern, later part of the palaeovalley, it is characteried by channel fills deposited during decreasing discharge by bedload streams thought to have formed a through-going fluvial system.

4.4.7. Mediano

4.4.7(a) Introduction

A number of outcrops of the Collegats Group occur within the area of the Mediano diapir centred on [7284]. None of the localities have been accurately dated, but they all have a strong Collegats Group lithostratigraphic and provenance signature.

The outcrops are important in three ways: (i) they record the existence of a second major transfer system sourced in the Axial and Nogueras Zones supplying sediment to the head of the Huesca fan; (ii) they help to document the dissolution induced collapse stage of the Mediano diapir {3.5.3(b)}, and (iii) they contain facies which indicate that the Huesca fan retreated at the end of the Collegats Group; these facies include palaeosols and limestones indicative of two types of lakes and a floodplain catena.

4.4.7(b) Sierra de Trillo

The Sierra de Trillo {7387} is composed of a thick sequence, over 300m, of boulder and cobble conglomerates. The conglomerates unconformably overlie the Escanilla Formation, and are bounded to the east and west by extensional faults which downthrow into the Mediano, and define the Palo Graben. Exposure is too poor to allow palaeocurrent or facies analysis. However, estimates of clast composition suggest that 14-22% of the clasts are derived from Axial and Nogueras Zone sources (Table 4.7). Clasts are well rounded and typically boulder sized, up to 1.2m in diameter.

Interpretation

The conglomerates are very coarse - similar in grain size to the Sis Conglomerate and much coarser than the Graus Conglomerate Formation. The grain-size and the existence of a well defined palaeovalley at Graus to the east of the Mediano, implies that this conglomerate did not link with the Graus Conglomerate Formation at Graus. It is suggested that the outcrop records the presence of a second discrete transfer system, the Trillo transfer system, located within the Palo graben which is thought to have been developing by this time {3.4.2}.

4.4.7(c) Tozal Panchado

The section is located within a down-faulted block below Tozal Panchado [750775]. The sequence unconformably overlies Escanilla Formation sediments and contains the intra-Collegats unconformity discussed in section 3.4.2. (Fig.3.24).

Description

The section is composed of silts punctuated by conglomerate sheets (Fig.4.74). The conglomerate bodies lack small-scale F.U. sequences, and are composed of gravel-filled scours (Gsc). Clasts are rounded, poorly sorted and 5-39% can be related to the Axial and Nogueras Zones (Table 4.7). An exception is the topmost conglomerate body (167-187m) which is composed of large angular blocks (up to 1m) of tufa and the San Martin Formation. In addition, rare angular pebbles of the San Martin Formation can be recognised within conglomerate bodies from 93m upwards. This feature approximately coincides with a marked change in palaeocurrent directions: below 85m palaeocurrents are directed to the south, whereas above they are directed largely to the west.

Interpretation

The coarse sediment bodies are interpreted as multistorey and multilateral deposits, where continual reworking prevented the preservation of F.U. sequences. The range of conglomerate facies is comparable to that seen in the Graus Conglomerate Formation. Yet, although the sequence is considered to overlie the same unconformity, sediment bodies are poorly connected. The increased proportion of fine-grained facies, together with the changes in clast provenance and palaeocurents, are interpreted as the result of movement on the N.W.-S.E. trending fault which bounds the outcrop to the south. The first part of the sequence 47-85m is considered to be derived from the Palo transfer system, flowing south down the core of the Mediano west of the Caneto Fault System. The second part (85-168m) is related to part of the Graus transfer system/extrabasinal river swinging westwards following a phase of increased fault activity.

The topmost conglomerate body unconformably overlies the fault plane and tilted Collegats beds. The associated change in paleocurrent direction and the dominance of coarse local limestone and (fault spring-line?) tufa deposits suggests that the San Martin area had important relief at this stage and shed a small alluvial fan to the south. The cause of the sediment pulse is unclear.

4.4.7. (d) Secastilla

The section is located in a v-shaped graben to the north of Secastilla. ([751764-742743]; Fig.4.75). The graben-fill is composed of siltstones and rare clays punctuated by thick conglomerates (Fig.4.76). The conglomerates thicken towards the western margin of the graben suggesting subsidence during deposition. The displacement on the western graben bounding fault dies out upwards, suggesting that the graben fill is a time equivalent of the footwall sequence. The footwall differs markedly from the graben fill in being composed of stacked conglomerates. As with the Tozal Panchado graben described above, the situation appears to be in conflict with established models of fluvial architecture (Bridge and Leeder 1979) which suggest that rivers should avulse towards rapidly subsiding areas e.g. towards grabens. However, preferential avulsion will only occur if there is a marked gradient across the floodplain. Where aggradation prevents such a gradient developing, the normal, non-faulted models relating subsidence and connectedness of Bridge and Leeder (op.cit.) wil apply. The sequences here are interpreted in this way with the lower connectedness of the graben-fill resulting from higher subsidence.

The cause of the subsidence is discussed elsewhere {3.5.3.(b)}.

Description of the log

The section is 555m thick, and is composed of pedogenically altered siltstones and clays, punctuated by coarse sediment bodies.

In the upper part of the sequence where exposures are the most accessible, the coarse sediment bodies are characterised by 7-10m thick F.U. sequences (Fig.4.76). Sequence bases are erosional, and overlain by conglomerate traction current facies: massive gravels (Gm); gravel-filled scours (Gsc) and planar cross-bedded gravel (Gp). These, in turn, are overlain by a F.U. sequence commencing with Gsc facies and passing upwards into pebbly-sand filled scours (Ss), silts (F) punctuated by sandy sheetflood (Sf) units, and clay. The sandstones, silts and clays are invariably pedogenically altered. A single C.U. sequence also occurs (clay-silt-fine sandstone-conglomerate). Exposures are not suitable for detailed architectural or sediment body geometry studies.

Large-scale sequences are restricted to the presence or absence of conglomerate units - there is no systematic C.U. or F.U.

Palaeocurrents are to the south (parallel to the graben margins) throughout. Clasts are well-rounded and moderately well sorted. Compositionally, the percentage of allochthonous clasts varies widely, 4-59% of the clasts correspond to lithologies outcropping in the Axial and Nogueras Zones - in one bed 31% of the clasts are granitic (Table 4.7).

The complete range of observational soil facies {4.2.3(d)(f)} occurs. Pseudogley soils (S_1 and S_2) and hydromorphically mottled soils (S_4) occur throughout. Calcretes are associated with soil facies S_1 and S_2, so that they can be assigned to the soil suborder Ustalf. They are dominantly nodular (C_2) although the entire range of calcrete types is observed especially towards the top of the sequence where calcrete becomes more important. The increase in calcrete is associated with the prominent development of hydromorphically mottled soils (S_4) and occasional lacustrine limestones.

Well developed gley soils (S_3) are restricted to the finest parts of the sequence. They occur at the top of F.U. sequences and are separated from the coarse sediment bodies by other soil types (e.g. 431-457m Fig.4.76). Shelly material, lignite and oxidised Fe-burrows with associated sulphur, in addition to the gley mottling, suggests that both "gleys" and "sulfaquents" were developed.

Interpretation

On the grounds of erosive sequences bases, coarse grain-size traction current facies, and F.U. motif the conglomerate sediment bodies are interpreted as the deposits of bedload rivers. The planar cross-bedded gravel facies (Gp) is indicative of significant bar relief. In the absence of slough channel deposits this relief is interpreted to reflect a braided channel pattern {4.3.1)b)(i)}.

The sequences are much thicker than the 2-3m thick F.U. braided stream deposits described from elsewhere in the Collegats Group. They are closely comparable to the type A conglomerate-sandstone association of Hayward (1983) who interpreted the sequence thickness to be the product of high subsidence rates. This is compatible with the graben setting. The F.U. motif is interpreted as the result of decreasing discharge and shallowing of water over accreting bars (Hayward 1983). Again it is expanded with respect to cycles seen elsewhere. In addition to increased subsidence, this could reflect greater preservation or deeper channels.

Sandy sheetfloods and thin channel sequences (1-2m) not associated with the major F.U. sequences are interpreted as crevasse splays and channels formed buring periods of river flood.

The range of soil facies (S_{1-4}) all indicate a seasonal climate. The Ustalfs (S_{1-2}-C_2) can be specifically interpreted as soils which supported a savanna-like vegetation, under a climate intermediate between warm arid and warm humid. The upward increase in the maturity of calcrete profiles may reflect a drier climate. Unfortunately the influence of other factors (e.g. subsidence rates) cannot be evaluated.

The gley soils, in particular the sulfaquents, are interpreted as lake margin soils. Their persistent position in the finest parts of the sequence, and their separation from the channel deposits by other soil facies ($S_{1,2,4}$) suggests that they formed in an interchannel position separated from the channel fills by a catena, a sequence of soil types controlled by relief. The implication is that channel levees were areas of slightly higher topography, and better drainage.

The gley soils represent a distinctly different type of lake margin to the lacustrine limestones. They might represent clastic-sediment-rich and clastic-sediment-poor margins of the same lake. More likely, since the palustrine limestones record frequent subaerial exposure and the sulfaquents are the product of permanently saturated soils, they represent margins of two different lake types. The change in lake types may well be a reflection of increased aridity towards the top of the sequence.

4.5. SYNTHESIS OF LOCALITIES

The Collegats Group fluvial system was initiated by tectonic uplift in the Nogueras and Axial Zones. Its development can be synthesised into five stages which reflect episodes of tectonic uplift in the External Sierras. (Fig.4.77(a)-(e)). The correlations between the key localities on which this synthesis is based are summarised on Figure 4.36.

4.5.1 The drainage basin

The key feature of the drainage pattern is that the fluvial system of *source area, transfer system* and *sedimentary basin* is maintained throughout.

The actual drainage pattern is controlled by relief: (i) relief tectonically generated during the deposition of the Collegats Group and (ii) remnant relief the product of pre-Collegats structures and erosion.

Tectonically generated relief dominates the Nogueras and Axial Zones, where a reticulate drainage pattern (following dip slopes, and fold and thrust planes) is implied. It also dominates in the basin where the development and rejuvenation of the the External Sierras causes major changes in sediment dispersal patterns.

Remnant relief, the product of differential erosion and older structures was also important. The Montsech high and the eastern margin of the Mediano bounded the Graus palaeovalley even though they were inactive during the Oligo-Miocene. The Boltana anticline did not form the western margin of the head of the Huesca Fan, but it may have operated as an interfleuve between the catchment area of the Luna and Huesca fans.

At no stage is the Barbastro anticline thought to affect the drainage pattern (Hirst 1983).

4.5.2 Stage 1

The figure shows the first folding in the External Sierras: it immediately postdates regional uplift and the development of the triangle zone in the Axial and Nogueras Zones (Fig.4.77(a)).

The palaeogeography is dominated by three major transfer systems which are separated by wide, actively eroded interfluves and have catchment areas extending into the Axial and Nogueras Zones: the Sis-Graus {4.4.1;4.4.2(a)}; the Mediano and the Colungo transfer systems. Of these only the first is known to have been an early feature. The other transfer systems are inferred to be early by analogy - since the primary control on the initiation of the Sis-Graus transfer system is considered to have been uplift in the Nogueras Zone. They were constrained to flow through structural lows in the External Sierras e.g. the Sis-Graus transfer system flows between the recumbent fold at Urbiego and the Olvena backfold.

The uplift at the External Sierras is interpreted to have produced unit (a) at Olvena {4.4.3(b)}.

Since sediments considered to be the downstream equivalent of the Sis-Graus transfer system were folded by the External Sierras {4.4.2(d)} this palaeogeography is likely to have been predated by one without the External Sierras.

4.5.3 Stage 2

Stage 2 is characterised by a marked phase of deformation and uplift in the External Sierras which transformed the palaeogeography (Fig.4.77(b)). The most dramatic changes are at Juseau where a backshed fan developed, and at Olvena where an early palaeovalley become closed and flow along it reversed.

By this stage the basin had expanded so that the transfer systems were shorter and the head of the Huesca Fan was within and to the north of the External Sierras. Sediment supply to the mid and distal-fan areas was uninterrupted as a result of a number of structural lows in the External Sierras. At the southern margin of the Sierras a major distributary turned eastwards hugging the thrust generated topography. Elsewhere distributaries flowed into blind structural lows causing rapid downstream changes, from stacked braided rivers to sandy terminal fan sequences over distance of 10km.

Dissolution collapse of the Mediano diapir is believed to have commenced at this stage, at first localising the position of the Trillo transfer system and later causing a distributary from the Sis-Graus transfer system to turn westwards around the southern end of the Mediano and flow towards the developing graben.

The main fluvial system was slightly finer than in stage 1 reflecting switching of the active deformation front to the External Sierras and cessation of uplift in the triangle zone.

4.5.4 Stage 3

Stage 3 is one of relative quiescence, characterised by gradual aggradation of the system, further expansion of the basin and geomorphic evolution.

Through aggradation and denudation the closed palaeovalley at Olvena became breached at its western end and periodically overtopped at its northern margin by a distributary of the Huesca fan. Breaching radically altered the local palaeogeography, and increased the percentage of locally sourced clasts in the bedload rivers which dominated sedimentation at the southern margin of the Sierras.

Meanwhile active headward erosion in the Mas de Nicolau palaeovalley breached the northern margin of the External Sierras at Juseau, capturing part of the Huesca Fan distributary system. A vigourous erosive, non-depositional palaeovalley resulted, debouching coarse sediment onto the Huesca fan, at a short-lived secondary fan-apex.

4.5.5 Stage 4

Stage 4 is characterised by a second period of uplift and deformation in the Fxternal Sierras. Its effects are similar to those seen in stage 2. They are most clearly seen at Juseau, where uplift caused, firstly, the rapid aggradation and closure of the palaeovalley and, secondly, the development of a syntectonic sediment apron which passed northwards into a time equivalent terminal fan sequence. The now closed Mas de Nicolau palaeovalley may have begun to aggrade at this stage: tectonic uplift initiating a locally-sourced, coarse, axial bedload-river system.

In the Esera Gorge this uplift may relate to the erosionally based palaeovalley below Olvena {4.3.3(c)(iv)} or to the alluvial fan at the northern end of the gorge {4.3.3(d)(iii)}.

A second phase of dissolution collapse at the Mediano is believed to have commenced at this stage. The resulting fault system controlled local sediment dispersal patterns, while the associated rapid subsidence produced expanded, poorly-connected conglomerate sequences.

4.4.6 Stage 5

During the final stage the Collegats Group basin was at its largest, but compressional tectonics had ceased and slopes were denuded, so that rates of sediment supply were low. In response to low sedimentation rates, lacustrine environments, previously characteristic of the basin centre (Hirst and Nichols 1986), expanded to impinge on the basin margin. Diapiric uplift continued locally, folding the margins of the Mas de Nicolau palaeovalley to produce progressive and angular unconformities, and local fans.

A range of palaeosols developed in interdistributary settings, indicating a seasonal climate gradually changing from warm-humid to warm-arid. In response to this lacustrine and lake margin (palustrine) sediments changed from anoxic, permanently saturated Sulfaquents (USDA 1975) to oxic periodically exposed lacustrine limestones.

4.6 CONCLUSIONS - THE CONTROLS ON FLUVIAL SEDIMENTATION

A number of factors are recurrent in controlling the nature and development of the Collegats Group. Their relative importance is qualitatively assessed below.

1. Time

Conceptually, time is the fundamental control: there can be no no change without it.

At a less abstract level its most important feature is that landscapes evolve with time. They become denuded and supply progressively less sediment, producing, in this example, the large-scale F.U. megasequence of the Collegats Group, and F.U. at the top of many small alluvial fans.

A crude estimate suggests that the Collegats Group was deposited within 5-10m.a.

2. Tectonics

The fingerprint of tectonics is everywhere in the Collegats Group. Thrust loading is believed to have generated the basin, while tectonically generated relief is believed to have played four key roles. Firstly, it produced sediment sources, at the basin margin and within the basin. Secondly, it was a major control on drainage patterns, both within the catchment area and the basin - key fluvial environments e.g. small alluvial fans and terminal fans were a direct response to intrabasinal uplift, while the main transfer systems, and tributary valleys followed structural lows. Thirdly, it contolled, together with diapiric dissolution collapse, variations in susidence rates. Fourthly, the high gradients produced by tectonic uplift favoured bedload streams.

3. Remnant, erosional, and depositional relief

Remnant relief, the product of pre-Collegats Group structures and erosion, controlled (together with tectonic relief) drainage patterns in the catchment area and at the fan apex. This relief may have supplied large amounts of sediment, but it did not generate distinct fan bodies.

Channel avulsion and palaeosol development are believed to have been influenced by the depositional relief of aggraded channel belts: channels avulse away from elevated, aggraded areas, which are relatively well-drained with respect to the interdistributary areas, producing a distinct catena of soil types (in favourable settings).

4. Sediment supply

The high rates of coarse sediment supply resulting from tectonic uplift are believed to have had three effects: firstly they favoured bedload rivers; secondly, where rates of supply exceeded river competence they caused aggradation, and, thirdly, they increased the "connectedness" of coarse sediment bodies.

5. Subsidence

Variable rates of subsidence controlled fluvial architecture. High rates of subsidence in narrow grabens formed by diapir dissolution collapse, produced relatively thick sequences with low degrees of channel connectedness. Preferential avulsion towards areas of rapid subsidence is thought to have been prevented by high sedimentation rates which prevented the development of fault induced topography.

6. Source area lithology

Source area lithology is an important control on the facies of small alluvial fans (4.4.3). Fans sourced from limestone source areas are characterised by well or poorly sorted, clast supported deposits: gravel-filled scours (Gsc) and clast supported sediment gravity flows. By contrast, fans sourced from mixed limestone, gabbro and gypsum lithologies are characterised by matrix supported sediment-gravity-flows and gravel-filled scours (Gsc) and may be fringed by playa deposits.

In addition, where favoured, calcretes are likely to have developed rapidly because of the dominance of carbonate lithologies in all source areas (Atkinson 1983;1986).

7. Catchment area - km^2

Larger catchment areas are considered to damp storm hydrographs and favour steady stream flow.

8. Climate

Climate is the major control on river discharge and on palaeosols. Palaeosols record a sub-humid to sub-arid, seasonal climate.

The main fluvial system records unsteady flow at two scales: 10-30cm thick sheetflood units, recording discrete hydrographic events probably produced by a single storm, and repeated 2-3m F.U. channel fill sequences interpreted as the product of loger term (seasonal?) discharge fluctuations.

In addition, climatic change may have helped to increase (4.4.1) or decrease sediment supply (4.4.2) primarily by controlling effective vegetation cover (Schuum 1977). However, a direct link between sediment supply and climatic change could not be demonstrated at any stage.

9. Sea-level fluctuations

Sea-level fluctuations had no effect on Collegats Group sedimentation as the Ebro basin had an internal drainage system during deposition of the Collegats Group.

4.7. DISCUSSION

The study shows the importance of a linked sedimentological and structural approach. Detailed structural analysis produces a framework for sedimentological investigation, outlining the major lithostratigraphic divisions and a crude sequence of events. Sedimentology can test and refine this early picture to produce a coherent synthesis of the evolution of an area.

The approach is particularly fruitful for the well exposed key localities (4.4). By applying a range of well established techniques and by comparison with similar settings, detailed in the literature, the evolution of a relatively small area can be outlined in detail. From a number of these localities key environments can be defined and the way in which the fluvial system responds to change in the major variables can be recorded.

However, in this study, correlation between areas is not complete, so that the exact palaeogeographic evolution is unknown. In this respect magnetostratigraphic studies hold the most promise. In addition to testing the correlations implied here, accurate correlations would be particularly useful in feeding the climatic changes deduced from palaeosols into an assessment of changes in the fluvial pile.

Futher work would have enhanced the study in a number of ways. Work on the sediments at the northern limb of the Barbastro anticline would have helped to test the geometry of the Huesca Fan shown by Hirst and Nichols (1986; Fig.4.20) and perhaps tied their data more closely to the more complex sequence of events documented here. A study of the well exposed sediments south of Gabasa would have been useful for similar reasons, as well as being able to test the sequence of events implied above for the Mas de Nicolau-Juseau palaeovalley - was this area an important secondary fan apex at some stage? The area south and east of Benabarre could also be the subject of further attention: did a transfer system ever drain directly to the Ebro Basin through this area, or was the area always a blind structural low? Although these areas are likely to hold key answers they suffer increasingly from problems of poor correlation.

The application of the sediment gravity flow model of Nemec and

Steel (1984) by the exhaustive collection of clast size and bed thickness data was not a success. Spurious correlations were common - too common to hold faith in those data sets which appeared to agree well with the model. Although a wide range of factors can explain these correlations, their lack of success in consistently advancing the sedimentological interpretations leads the author to recommend that the method is not used again.

CHAPTER 5

...what sort of an earth would it have been after the following processes had passed over it. The Moon jumping out of the Pacific ocean, Wandering Continents, a shrinking Globe, Africa pushing against Europe and leaving parts of Africa on the Alps, Mountains that have travelled over volcanos, Radioactive volcanos, half of Scotland breaking loose and wandering 65 miles to the North-East. All these geological fairy tales and red herrings form part of the stock-in-trade of our professors and teachers and are sponsored by *Nature* and the Royal and Geological Societies.

C.T. Trenchmann "A New Explanation of Mountain Uplift" (1955)

CHAPTER 5

CONCLUSIONS

From each of the studied aspects of the South Pyrenean foreland basin - balanced sections, Oligo-Miocene fluvial molasse sedimentation and diapirism - a number of conclusions, and problems can be outlined:

5.1 BALANCED SECTIONS AND BALANCED MAPS

5.1.1 Conclusions

Two sequentially restored balanced sections and a series of sequentially restored balanced maps outline five major tectonic phases: (i) Upper Cretaceous extension producing approximately WNW-ESE trending, down-to-the-north faults which defined a major turbidite basin, the Vallcarga basin, in the north of the area; (ii) inversion of suitably orientated faults by transpressional, dextral strike-slip during the Maastrichtian and Paleocene; (iii) emplacement of the Cotiella thrust sheet into the foreland basin during the Lower Eocene; (iv) switching of the thrust front from within the basin to its northern margin where a triangle zone developed from the Middle Eocene to the uppermost Oligocene; and (v) cessation of uplift at the northern basin margin with the emplacement of a second major thrust sheet in the foreland basin during uppermost Oligocene and Miocene times.

5.1.2 Discussion

The balanced sections were developed in an iterative fashion: structural mapping led to the development of preliminary sections, which were then tested by comparison with the preorogenic and synorogenic sequences.

Preorogenic sequences constrain the restored section: the restored section should show the geometry of the basin in which they were deposited. Particularly useful features are: marked thickness variations, facies variations, and steep ramp angles, each of which may indicate syndepositional faulting, and a non layer-cake preorogenic sequence.

Synorogenic sequences constrain the deformed state section. Features such as syntectonic unconformities, marked changes in palaeogeography and sudden coarse clastic input can assist in defining periods of thrust movement. Conversely, the locations of uplift (culminations and large ramps) and their order of occurrence, as indicated by the thrust sequence shown on the deformed state section, should be compatible with this history. Inidividual synorogenic sequences can be used as a "regional" or reference level, for a particular stage of restoration - a particularly useful method when the basin margin is preserved so that the amount of uplift which had occurred at a given stage can be gauged.

Sections incompatible with the preorogenic and synorogenic sequences should be modified or rejected.

Balanced maps or three-dimensional balanced sections offer two advances on a series of balanced sections: (i) they demonstrate compatibility between sections; and (ii) they show a truely integrated structural and sedimentological palaeogeographic basin evolution. The maps presented here represent the first attempt to do this in the Pyrenees (indeed the author knows of no other study where this technique has been used).

Three problems were encountered in developing the balanced maps presented here: the reliability of the thrust models; poor correlation; and an incomplete data set.

The structural interpretations are based on the assumption that only one thrust front operated at any one time. Sedimentary sequences during the best constrained periods support this model. At other times high rates of sediment input at the northern margin of the basin may have been coeval with the development of thrust fronts in the basin, so that this model could be inappropriate. At present the lack of a detailed basinwide correlation, especially in the continental molasse deposits, prevents testing of this alternative. The problem has few implications for thrust geometries, but it does affect interpreted phases of uplift, and the model of switching of the thrust front between the foreland basin and internal zones.

Further work on balanced sections in the area awaits: the publication of seismic lines; a detailed basin-wide correlation, allowing outcrops separated by erosion or structural features to be linked; and the study of certain key localities especially the upper part of the Ager basin fill and Tertiary sequences preserved in synclines in the External Sierras.

5.2 DIAPIRISM

Four features have emerged from the study of selected diapirs:

5.2.1 The recognition of diapirism

Six features were found to be useful (i) evidence of halite in the core of the structure; (ii) thickening of evaporites towards the core of a given structure; (iii) intrusive contacts between evaporites and younger rocks; (iv) assemblages indicative of cap rocks; (v) structures within evaporite bodies (steeply dipping fabrics and folds) are supportive of diapiric activity but exposure is often too poor to study them systematically; and (vi) features in the carapace of the diapir.

5.2.2 The effects of diapirs on their carapace and syndiapiric sediments

Five main effects have been recognised: (i) unconformities may develop over the crest of the diapir indicating localised uplift; (ii) syn-diapiric sequences thin towards diapir crests forming cumulative wedge systems and indicating prolonged diapir growth; (iii) facies changes and changes in sediment dispersal patterns may occur towards diapir crests and indicate that the diapir was a palaeohigh; (iv) diapiric highs may become an important sediment source - by inducing resedimentation of sediments deposited on the diapir, or by the reworking of evaporites which reach the surface; (v) faults produced by dissolution collapse, can have large throws and control variations in sediment thickness and sediment dispersal.

5.2.3 Initiation of diapirism

All the studied diapirs are considered to have been initiated by alpine compression; either by thrust induced loading or by the development of a pillow or piercement structure from a salt cored fold.

5.2.4 Level of intrusion

Two of the studied diapirs are considered to have reached the sediment surface while they were diapirically active.

5.3 OLIGO-MIOCENE FLUVIAL MOLASSE SEDIMENTATION

5.3.1 Introduction

Oligo-Miocene fluvial molasse sedimentation can be characterised by four environments: bedload rivers; large and small-scale alluvial fans; terminal fans; and transfer system and tributary palaeovalleys. These distinct environments are considered to be controlled by eight factors: tectonics; remnant, erosional and depositional relief; sediment supply; subsidence; source area lithology; catchment area; climate; and time.

5.3.2 ENVIRONMENTS

(a) Bedload rivers

Many of the conglomerate succesions in the Collegats Group are characterised by erosively-based small-scale sequences (1-3m thick) which are dominated by conglomerate, but may fine upwards in their top 10-30cm to sand grade. The conglomerates are characterised by traction current facies, most commonly massive gravels (Gm) and gravel-filled scours (Gsc) with subordinate planar cross-bedded gravel (Gp), low-angle cross-bedded gravel (Gl), and oblique gravel-filled scours (Gf). Common facies transitions are Gm-Gsc-S, Gsc-S, Gm-S upwards.

On the grounds of cycle scale, erosive base, traction current facies and coarse grain-size, the cycles are interpreted as inchannel deposits of bedload rivers. The crude F.U. motif and observed facies transitions are interpreted as the product of decreasing discharge, possibly reflecting seasonal discharge variations. Where present the Gp, Gl, and Gf facies indicate that there was significant bar relief and that the channels were probably braided.

Erosively-based coarsening-upwards sequences occur rarely. They are characterised by the facies transition Gsc-Gm upwards and are interpreted as the result of re-establishment of flow in a (partially) abandoned channel.

5.3.2(b) Small alluvial fans

A number of small alluvial fans are sourced from syn-sedimentary folds and thrust generated intrabasinal highs within the foreland basin. These fans have areas of 0.5–2km^2 and are 7–80m thick, although one fan attains a thickness of 150m. The 7–80m thick fans are sourced from dip-slope catchment areas, and characterised by mixed sediment-gravity-flow and stream-flow facies; fans sourced from Cretaceous to Paleocene limestones are dominated by clast supported flows, while fans sourced from Triassic lithologies are dominated by matrix supported flows and, furthermore, have playa-like fringes. The single 150m thick fan sequence has a syncline for a catchment area. It is characterised by stream-flow facies only and by the presence of small-scale F.U. sequences (1–3m) indicative of bedload rivers.

Many of the fan bodies overlie discrete unconformities or show evidence for stratal wedging indicating that the fans were initiated by tectonic uplift.

5.3.2(c) Large alluvial fan

The proximal part of a large alluvial fan, the Huesca fan (Hirst and Nichols 1986) occurs in the study area. Combining work presented here with the results of Hirst and Nichols (op.cit.) proximal-distal trends can be outlined over the 70km of fan radius.

The fan head is composite - it was supplied by a number of discrete fluvial transfer systems, and is characterised by stacked channel-fill conglomerates deposited by braided bedload rivers. The mid-fan area is dominated by erosively-based sheet sandstone bodies (62%) interpreted as laterally migrating or braided sandy channels. The distal part of the fan occurs towards the centre of the Ebro basin and is characterised by ribbon sandstone bodies (suggesting laterally stable fluvial channels) interdigitated with lacustrine sediments. The fan has a radial geometry; channels form a distributive pattern and die out into the Ebro basin, so that the fan can be considered as a large terminal fan.

5.3.2(d) Terminal rivers

A number of terminal fan sequences occur in the Collegats Group. They developed in the proximal part of the Huesca fan adjacent to intrabasinal highs and are interpreted as the response of the fluvial system to blind structural lows. They differ from the Huesca fan in their origin, and in the nature and rate of down fan changes which occur over 10km.

The proximal part of these fans is characterised by conglomeratic bedload stream deposits. The conglomerates pass downstream into sediments with similar small-scale cycles (1-3m) which are increasingly characterised by stacked sheetfloods. By a mid-fan position erosive cycle bases are lost and the cycles are dominated by sandy sheetfloods units deposited in lobe settings. Beyond the lobes individual sandstone beds punctuate a silt background.

5.3.2(e) Palaeovalleys

Palaeovalleys in the Collegats Group occur at two scales: transfer system palaeovalleys which supplied the majority of the sediment and were sourced from the basin margin; and tributary paleovalleys which drained (or drained across) intrabasinal highs.

The transfer system palaeovalleys are the coarsest part of the entire fluvial system with boulders over 2m in diameter. The most proximal studied transfer system is characterised by 10-20m thick multistorey and multilateral sheets. Individual storeys are erosively based, 1.5-3m thick, massive gravel beds (Gm) interpreted as the deposits of bedload rivers at high stage when the river bed was devoid of braid bars. The downstream ends of these palaeovalleys merge, and have facies comparable with the head of the Huesca fan.

The fills of tributary valleys are highly variable, reflecting: their mode of initiation - tectonic uplift so that they drain synclines or strike perpendicular dip-slopes, or river capture producing palaeovalleys which lie across structural trends; the direction of sediment supply to the valley - longitudinal, with or without marginal sources; and the causes of filling.

Palaeovalley fills can be described using alluvial fan models, with the proviso that in addition to upstream controls (e.g. sediment supply in excess of stream capacity) downstream controls (e.g. uplift across a palaeovalley) can also initiate valley aggradation.

5.3.3 Factors controlling Oligo-Miocene sedimentation

The most important factor controlling Oligo-Miocene sedimentation was tectonic activity. Crustal loading generated the basin, while tectonically generated relief is considered to have had three key roles:

(i) it produced sediment sources, through a triangle zone at the northern margin of the basin, and, following a switch of the active thrust front to within the foreland basin, local sediment sources through intrabasinal highs;

(ii) it was the major control on drainage patterns, both within the catchment area where transfer systems followed structural lows, and within the basin, where small alluvial fans and terminal fans were a direct response to intrabasinal uplift, and the main transfer systems and tributary valleys followed structural lows;

(iii) thrust generated uplift and diapir dissolution collapse controlled variations in subsidence rates.

Other factors are more subtle, three are of particlar interest: source area lithology; catchment area; and climate.

The facies assemblage of small alluvial fans is largely controlled by source area lithology: limestone source areas produce clast-supported sediment gravity flows, and gravel-filled scours (Gsc); Triassic source areas composed of gabbro, gypsum and minor limestone, produce matrix supported sediment gravity flows, Gsc facies and playa-like fan fringes.

Larger catchment areas are considered to damp storm hydrographs and favour steady stream-flow.

Climate is the main control on river discharge and palaeosols. palaeosols record a subhumid to subarid climate. The main fluvial system records unsteady flow at two scales: 10-30cm thick sheetfloods units produced by a single storm and repeated 2-3m thick F.U. channel fill sequences interpreted as the product of longer-term (seasonal?) discharge fluctuations.

5.3.4 Discussion

The study of the Collegats Group was enhanced by a linked sedimentological and structural approach. A broad structural and preliminary sedimentological analysis formed the basis for the erection of hypotheses which could be tested and refined by detailed sedimentological study.

Wide areas of well-exposed Oligo-Miocene sediments in the External Sierras south of the Montsech remain unstudied, and require primary description and interpretation. Further research and more detailed work in the study area awaits improved correlations and the erection of a chronostratigraphy: are the suggested correlations between the key localities justified, and, of greater interest, how much time does an individual fan body represent, how long does a palaeovalley take to fill, and what are the rates of diapir dissolution collapse?

REFERENCES

Albarede, F and Michard-Vitrac, A. (1978) Age and significance of the North Pyrenean Metamorphism. *Earth Planet. Sci. Lett.*, **40**, 327-332.

Allen, J.R.L. (1974) Studies in fluviatile sedimentation: implications of pedogenic carbonate units, Lower Old Red Sandstone. Anglo-Welsh outcrop. *Geol. J.*, **9**, 181-208.

Allen, P.A., Homewood, P. and Williams, G.D. (1986) Foreland basins: an introduction. In: *Foreland Basins* (Ed. by P.A. Allen and P. Homewood). *Spec. Publs. int. Ass. Sediment.*, **8**, 3-12.

Anadon, P., Marzo, M. and Puigdefabregas, C. (1985) The Eocene Fan-Delta of Montserrat (Southeastern Ebro Basin, Spain). In: *Excursion Guidebook, 6th European Regional Meeting, int. Ass. Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain, 109-146.

Anadon, P., Cabrera, L., Colombo, F., Marzo, M. and Riba, O. (1986) Syntectonic intraformational unconformities in alluvial fan deposits, eastern Ebro Basin margins (NE Spain). In: *Foreland Basins* (Ed. by P.A. Allen and P. Homewood). *Spec. Publs. int. Ass. Sediment.*, **8**, 259-271.

Anderson, E.M. (1942) *The Dynamics of Faulting.* 1st Ed. Oliver and Boyd, Edinburgh.

Atkinson, C.D. (1983) *Comparative sequences of ancient fluviatile deposits in the Tertiary South Pyrenean Basin, Northern Spain.* Unpublished Ph.D. Thesis, University of Wales, pp.360.

Atkinson, C.D. (1986) Tectonic control on alluvial sedimentation as revealed by an ancient catena in the Capella Formation (Eocene) of Northern Spain. In: *Paleosols their recognition and interpretation* (Ed. by V.P. Wright), pp.139-179. Blackwell Scientific Publications, Oxford.

Beaumont, C. (1981) Foreland Basins. *Geophys. J. R. astr. Soc.*, 65, 291-329.

Behrmann, R.B. (1949) Geologie und Lagerstatte des Olfeldes Reitbrook bei Hamburg. In:"*Erdol und Tektonik in Nordwestdeutschland*". Amt. Bodenf. Hanover Celle. pp.390.

Bernoulli, D. and Jenkyns, H.C. (1974) Alpine Mediterranean and Central Atlantic facies in relation to the early evolution of the Tethys. In: *Modern and*

Ancient Geosynclinal Sedimentation (Ed. by R.H. Dott and R.H. Shear). *Spec. Publ. Soc. econ. Paleont. Miner., Tulsa*, **19**, 129-140.

Bevan, T.G. and Hancock, P.L. (1986) A late Cenozoic regional mesofracture system in southern England and northern France. *J. geol. Soc. Lond.* **143**, 355-362.

Beverage, J.P. and Culbertson, J.K. (1964) Hyperconcentration of suspended sediment. *Proc. Am. Soc. Civil Engineers J. Hydraulics Division*, **90**, 117-128.

Blissenbach, E. (1954) Geology of alluvial fans in semi-arid regions. *Bull. geol. Soc. Am.*, **65**, 175-190.

Bluck, B.J. (1967) Deposition of some Upper Old Red Sandstone conglomerates in the Clyde area : a study in the significance of bedding. *Scot. J. Geol.*, **3**, 139-167.

Bluck, B.J. (1971) Sedimentation in the meandring River Endrick. *Scot. J. geol.*, **7**, 93-138.

Bluck, B.J. (1976) Sedimentation in some Scottish rivers of low sinuosity. *Trans. R. Soc. Edinburgh*, **69**, 425-456.

Bluck, B.J. (1979) Structure of coarse-grained braided alluvium. *Trans. R. Soc. Edinburgh*, **70**, 181-221.

Bluck, B.J. (1980) Structure, generation and preservation of upward fining, braided stream cycles in the Old Red Sandstone of Scotland. *Trans. R. Soc. Edinburgh*, **71**, 29-46.

Boillet, G. (1984) Some remarks on the continental margins in the Aquitaine and French Pyrenees. *Geol. Mag.*, **121**, 407-412.

Boothroyd, J.C. and Ashley, G.M. (1975) Processes, bar morphology and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska. In: *Glaciofluvial and Glaciomarine Sedimentation* (Ed. by A.V. Jopling and B.C. McDonald). *Spec. Publs. Soc. econ. Palaeont. Miner., Tulsa*, **23**, 193-222.

Boyer, S. and Elliott, D. (1982) Thrust systems. *Bull. Am. Ass. Pet. Geol.* 66, 1196-1230.

Bouma A.H. (1979) *Methods for the study of sedimentary structures.* R.E. Kreiger, Huntingdon, New York. pp.458.

Bown, T.M. and Kraus, M.J. (1981) Lower Eocene alluvial palaeosols (Willwood Formation, northwest Wyoming, U.S.A.)

and their significance for palaeoecology, palaeoclimatology and basin analysis. *Palaeogeog. Palaeoclim. Palaeocol.* **34**, 1-30.

Brewer, R. (1964) *Fabric and Mineral Analysis of Soils.* Wiley, New York. pp.470.

Bridge, J.S. and Leeder, M.R. (1979) A simulation model of alluvial stratigraphy. *Sedimentology* **26**, 617-644.

Brunet M.F. (1984) Subsidence history of the Aquitaine basin determined from subsidence curves. *Geol. Mag.*, **121**, 421-428.

Brunet, M.F. (1986) The influence of the evolution of the Pyrenees on adjacent basins. In: *The Geological evolution of the Pyrenees* (Ed. by E. Banda and S.M. Wickham). *Tectonophysics*, **129**, 343-354.

Bull, W.B. (1963) Alluvial fan deposits in Fresno County, California. *J. Geol.*, **71**, 243-251.

Bull, W.B. (1964a) *Geomorphology of alluvial fans in western Fresno County, California.* U.S. Geol Surv. Prof. Paper, **352-E**, 89-128.

Bull, W.B. (1964b) *Alluvial fans and near surface subsidence in western Fresno County, California.* U.S. Geol. Survey Prof. Paper, **437-A**. pp.70.

Bull, W.B. (1972) Recognition of alluvial-fan depsits in the stratigraphic record. In: *Recognition of Ancient Sedimentary Environments* (Ed. by J.K. Rigby and W.K. Hamblin). *Spec. Publs. Soc. econ, Palaeont. Miner., Tulsa*, **16**, 63-83.

Bureau de Recherches Geologiques et Minieres, ELF-Re, ESSO-REP, SNEAP (1974) *Geologie du Bassin d'Aquitaine.* Orleans: Bureau du Rescherces Geologiques et Minieres.

Busquets, P., Orti, F. Pueyo, J.J., Riba, O., Rosell, L., Saez, A., Salas, R. and Taberner, C. (1985) Evaporite deposition and diagenesis in the saline (potash) Catalan basin, Upper Eocene.
In: *Excursion Guidebook, 6th European Regional Meeting, int. Ass. Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain, 12-59.

Butler, R.W.H. (1982) Hangingwall strain: a function of duplex shape and footwall topography. *Tectonophysics*, **88**, 235-246.

Butler, R.W.H. and Coward, M.P. (1984) Geological constraints, structural evolution, and deep geology of the northwest Scottish Caledonides. *Tectonics*, **3**, 347-365.

Butler, R.W.H. (1985) Thrust tectonics; a personal view. *Geol.Mag.*, **122**, 223-232.

Butler, R.W.H. (1986) Thrust tectonics, deep stucture and crustal subduction in the Alps and Himalayas. *J. geol. Soc. Lond.*, **143**, 857-873.

Butler, R.W.H., Matthews, S.J. and Parish, M. (1986) The NW external Alpine Thrust Belt and its implications for the geometry of the Western Alpine Orogen. In: *Collision Tectonics* (Ed. M.P. Coward and A.C. Ries). *Spec Publ. geol. Soc. Lond.*, **19**, 245-260.

Buurman, P. (1980) Palaeosols in the Reading Beds (Paleocene) of Alum Bay, Isle of Wight U.K. *Sedimentology*, **27**, 593-606.

Cabrera, L. and Saez, A. (1987) Coal deposition in carbonate-rich lacustrine systems: the Calaf and Mequineza sequences (Oligocene eastern Ebro basin, NE Spain). *J. geol. Soc. Lond.* **144**, 451-461.

Cabrera, L., Colombo, F. and Robles, S. (1985) Sedimentation and tectonics interrelationships in the Paleogene marginal Alluvial systems of the SE Ebro basin. Transition from alluvial to shallow lacustrine environments. In: *Excursion Guidebook, 6th Regional Meeting, Int. Ass. Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain, 393-492.

Camara, P. and Klimowitz, J. 1985. Interpretacion Geodinamica de la Vertiente Centro-Occidental Surpirenaica. *Estudios Geolgicos* Vol.41; no.5-6. Madrid (in press).

Choukroune, P. and Seguret, M. (1973) Tectonics of the Pyrenees; role of compression and gravity. In: *Gravity and Tectonics* (Ed. by K. Dejong and R. Scholten). Wiley, New York. 141-156.

Coleman, W. and Spring, M. (1987) *Cornish Tunes.* Yertiz Publications, Gunnislake, Cornwall. pp.10.

Collinson, J.D. (1986) Alluvial sediments. In: *Sedimentary Environments and Facies, 2nd. Ed.* (Ed. by H.G. Reading) Blackwell Scientific Publications, Oxford. 20-62

Coneybeare, C.E.B. (1976) *Geomorphology of oil and gas fields in sandstone bodies.* Elsevier, Amsterdam. pp.344.

Cooper, M.A. (1981) The internal geometry of nappes: criteria for models of emplacement. In: *Thrust and Nappe*

Tectonics. (Edited by McClay, K.R. and Price, N.J.). *Spec. Publs. geol. Soc. Lond.*, 9. Blackwell Scientific Publications, Oxford. 225-234.

Costello, W.R. and Walker, R.G. (1972) Pleistocene sedimentology, Credit river, Soutern Ontario: a new component of the braided river model. *J. sedim. Petrol.*, 42, 389-400.

Coward, M.P. (1983) Thrust tectonics: thin skinned or thick skinned, and the continuation of thrusts to deep in the crust. *J. Struc. Geol.*, 5, 113-123.

Coward, M.P. and Butler, R.W.H. (1985) Thrust tectonics and the deep structure of the Pakistan Himalayas. *Geology* 13, 417-420.

Crumeyrolle, P. and Mutti, E. (1986) Stratigraphy and sedimentology of Santa Liestra sequence shelf depositional systems (South-Pyrenean Eocene basin, province of Huesca, Spain). *C. R. Acad. Sci. Paris, t. 303, Serie II,* 7, 581-584

Crusafont, M., Riba, O. and Villena, J. (1966) Nota preliminar sobre un nuevo yacimiento de vertebrados Aquitanienses en Santa Cilia (Rio Formiga; Provincia de Huesca) y sus consequencias geologicas. *Notas y Communicaciones del Instituto de Geologico y Minero de Espana*, 83, 7-14.

Cuevas Gozalo, M., Donselaar, M.E. and Nio, S.D. (1985) Eocene clastic tidal sedimentation in the Tremp-Graus basin. (Provs. Lerida and Huesca). In: *Excursion Guidebook, 6th European Regionnal Meeting, Int. Ass. Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain. 217-266.

Curnelle. R., Dubois, P. and Sequin, J.C. (1982) The Mesozoic-Tertiary evolution of the Aquitaine Basin. *Philos. Trans. R. Soc. London, Ser. A*, 305, 63-84.

Curnelle, R. and Marco, R. (1983) Reflection profiles across the Aquitaine basin (North Pyrenean basin). In: *Seismic expression of structural styles vol.1* (Ed. by A.W. Bally). *Am. Assn, Petrol. Geol. Studies in Geology Series*, 15.

Daignieres, M., Gallart, J., Banda, E. and Hirn, A. (1982) Implications of the seismic structure for the orogenic evolution of the Pyrenean range. *Earth Planet. Sci. Lett.*, 57, 88-110.

Dahlstrom, C.D.A. (1969) Balanced cross sections. *Can. J. Earth Sci.*, 6, 743-757.

Darwin, C. (1839) *The Voyage of the Beagle.* Extract taken from *The Voyage of Charles Darwin* 2nd Ed. (1979) by C. Ralling. British Broadcasting Corporation, London

Davis, D.M. and Engelder, T. (1985) The role of salt in fold-and-thrust belts. *Tectonophysics,* **119**, 67-88.

Deramond, J., Fisher, M., Hossack, J., Labaume, P., Seguret, M., Soula, J-C., Viallard, P. and Williams, G. (1984) *Thrusting and Deformation.* Field guide. Paul Sabatier University, Toulouse, France.

Deramond, J., Graham, R.H., Hossack, J.R., Baby, P. and Crouzet, G. (1985) Nouveau modele de la chaine des Pyrenees. *C.R. Acad. Sci. Paris, Ser. II,* 1213-1216

Dijk, C. van, Kruif, S. de, Oost, A.P., Postma, A.D., Pragt, J., Rumpff, D., Donselaar, M.E. and Boer, P.L. de (1985) Vertical stacking of facies boundaries on a moving and folding thrust sheet. *Abstr. 6th European Regional meeting, int. Ass. Sediment.* Lleida, Spain. 133-136.

Ditzel, J. (1966) *Sedimentologie van het Boven-Krijt in de omgeving van Torre de Tamurcia (Noord Spanje).* Internal report, Geol. Inst., Univ. Leiden, The Netherlands, pp78.

Doeglas, D.J. (1962) The structure of sedimentary deposits of braided streams. *Sedimentology,* **1**, 167-190.

Donselaar, M.E. and Nio, S.D. (1982) An Eocene tidal inlet/washover type barrier island complex in the S.Pyrenean marginal basin, Spain. *Geol. en Mijnbou,* **61**, 343-353.

Elliott, D. (1976) The energy balance and deformation mechanisms of thrust sheets . *Phil. Trans. R. Soc. Lond. A.* **283**, 289-312.

Elliott, D. (1983) The construction of balanced cross-sections. *J. Struct. Geol.* **5**, 101.

Elliott, D. and Johnson, M.R.W. (1980) Structural evolution in the northern part of the Moine thrust belt of NW Scotland. *Trans. R. Soc. Edin., Earth Sci.,* **71**, 69-96.

Elliott, T. (1986) Siliciclastic Shorelines. In: *Sedimentary Environments and Facies 2nd Ed.* (Ed. by H.G. Reading). Blackwell Scientific Publications, Oxford. 155-188.

Eisbacher, G.H., Carrigy, M.A. and Campbell, R.B. (1974)

Palaeodrainage patterns and late orogenic basins of the Canadian Cordillera. In: *Tectonics and Sedimentation* (Ed. by W.N. Dickenson). *Spec. Publ. Soc. econ. Palaeont. Miner., Tulsa,* 22, 143-166.

Esteban M. and Klappa C.F. (1983) Subaerial exposure environment. In: *Carbonate Depositional Environments* (Ed. by P.A. Scholle, D.G. Bebout and C.H. Moore). *Mem. Am. Ass. petrol. Geol.* 33. 1-54.

Eynon, G. and Walker, R.G. (1977) Facies relationships in Pleistocene outwash gravels, Southern Ontario: a model for bar growth in braided rivers. *Sedimentology,* 21, 43-70.

Feuillee, P., Villanova, M. and Winnock, E. (1973) La dynamique des fosses a "turbidites" et de leur content sedimentaire dans le systeme pyrenean. *Bull. Soc. Geol. Fr.,* 7, 61-76.

Filbrandt, J.B., Underhill, J.R. and Williams, G.D. (1985) Tectonic and foreland basin evolution in the Hellenides. In: *Programe and abstracts, Foreland Basins Symposium.* (Ed. by P. Homewood, P. Allen and G. Williams). Fribourg, Switzerland.

Fisk, H.N. (1944) *Geological investigation of the alluvial valley of the lower Mississippi River.* Mississippi River Commission, Vicksburg, Mississippi. pp.78.

FitzPatrick, E.A. (1984) *Micormorphology of Soils.* Chapman and Hall, New York. pp.433.

Freytet, P. (1973) Petrography and palaeoenvironment of continental carbonate deposits with particular reference to the Upper Cretaceous and Lower Eocene of Languedoc (southern France). *Sedim. Geol.,* 10, 25-60.

Friend, P.F. (1978) Distinctive features of some ancient river systems. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary,* 5, 531-543.

Friend, P.F., Slater, M.J. and Williams, R.C. (1979) Vertical and lateral building of river sandstone bodies, Ebro basin, Spain. *J. geol. Soc. London,* 136, 39-46.

Friend, P.F., Marzo, M., Nijman, W. and Puigdefabregas, C. (1981) Fluvial sedimentology in the Tertiary South Pyrenean and Ebro basins, Spain. In: *Field guides to modern and ancient fluvial systems in Britain and Spain* (Ed. by T.Elliott). *Int. Fluv. Conf, Dept. of Geol., Uni. of Keele.* Quadrant Partners, Swansea.

Friend, P.F. (1983) Towards the field classification of alluvial architecture or sequence. In: *Modern and ancient fluvial Systems* (Ed. by J.D. Collinson and J. Lewin). *Spec. Publs. int. Ass. Sediment.*, 6, 345-354.

Gallart, J., Daignieres, M., Banda, E., Sutinach, E. and Itisn, A. (1980) The East Pyrenean domain: lateral variations at crust-mantle level. *Ann.Geophys.*, 36, 141-158.

Gallart, J. Banda, E. and Daignieres, M. (1981) Crustal structure of the Palaeozoic Axial Zone of the Pyrenees and transition to the North Pyrenean Zone. *Ann. Geophs.*, 37, 457-480.

Gallart, J., Daignieres, M., Ganepain-Beynieux, J. and Hirn, A. (1985) Relationship between structure and seismicity in the western Pyrenees. *Ann. Geophys.*, 3, 239-248.

Galloway, W.E. (1981) Depositional architecture of Cenozoic Gulf Coastal Plain fluvial systems. In: *Recent and Ancient nonmarine depositional environments: models for exploration* (Ed. by F.G. Ethridge and R.M. Flores). *Spec. Publ. Soc. econ. Paleont. Miner., Tulsa*, 31, 127-155.

Garrido-Megias, A. (1968) Sobre la estratigrafia de los conglomerados de Campanue (Santa Liestra) y formaciones superiores del Eoceno (extremo occidental de la cuenca Tremp-Graus, Pirineo Central, provincia de Huesca). *Acta. Geol. Hisp.*, 3, 39-43.

Garrido-Megias, A. and Rios Aragues, L.M. (1972) Sintesis geologica del Secundario y Tercerio entre los rios Cinca y Segre (Pirineo central de la vertiente sur pirenaica, provincias de Huesca y Lerida). *Bol. Geol. y Min.*, 83, 1-47.

Geddes, A. (1960) The alluvial morphology of the Indo-Gangetic plain: its mapping and geographical significance. *Inst. British Geogr. Trans.*, 28, 253-276

Gole, C.V. and Chitale, S.V. (1966) Inland delta building activity of Kosi River. *J. Hydraul. Div., Proc. Am. Soc. Civ. Engrs.*, 92, 111-126.

Goudie, A.S. (1973) *Duricrusts in the Tropical and Sub-Tropical Environment*. Calendon Press, Oxford. pp.174.

Goudie, A.S. and Pye, K. (1983) *Chemical Sediments and* *geomorphology. Precipitates and Residua in the Near-Surface Environment* Academic Press, London. pp.439.

Goudie A.S. (1983) Calcrete. In: *Chemical sediments and geomorphology, precipitates and residua in the near-surface environment.* (Ed. by A.S. Goudie and K. Pye). Academic Press. 93-131.

Graham, R.H. (1985) In: *A structural and sedimentological traverse through the Tertiary foreland basin of the External Alps of South-east France* (Elliott, T., Apps, G., Davies, H., Evans, M., Ghibaudo, G. and Graham, R.H.) In: *Excursion Guidebook, Foreland basins Symposium* (Ed. by P.Allen, P. Homewood and G. Williams). Fribourg, Switzerland.

Grimaud. S, Boillet, G., Colette, B., Mauffret, A., Miles, P.R. and Roberts, D.B. (1982) Western extension of the Iberian-European plate boundary during the Early Cenozoic (Pyrenean) convergence: a new model. *Marine Geology,* 45, 63-77.

Grimaud, S., Boillot G., Colette, B., Mauffret, A., Miles, P.R. and Roberts, D.B. (1983) Western extension of the Iberian-European plate boundary during the Early Cenozoic (Pyrenean) convergence: a new model. Reply (after comment). *Marine Geology,* 53, 238-239.

Guimera, J. (1984) Paleogene evolution of deformation in the Northeastern Iberian Peninsula. *Geol. Mag.,* 121, 413-420.

Gussow, W.C. (1968) Salt diapirism: importance of temperature, and energy source of emplacement. In: *Diapirism and Diapirs* (Ed. by J. Braunstein and G. O'Brien). *Spec. Publs. Am. Ass. Petrol. Geol.,* 8, 16-52.

Hancock, P.L. (1986) Joint swarms indicative of halokinetic growth folding in the Jaca molasse basin, SW Pyrenees. In: *Sedimentation and tectonics in the Southern Pyrenees.* Abstracts of thematic meeting, University College Cardiff

Hardenbol, L. and Berggren, W.A. (1978) A new Paleogene Numerical Time Scale. In: *Contributions to the Geological Time Scale.* (Ed. by G.V. Cohee, M.F Glaessner and H.D. Heldberg) *Am. Ass. Petrol. Geol. Studies in Geology,* 6. 231-234.

Harding, T.P. and Lowell, J.D. (1979) Structural styles, their plate tectonic habitats and hydrocarbon traps in petroleum provinces. *Bull. Am. Ass. Petrol. Geol.,* 63, 1016-1058.

Harms, J.C., Southard, J.B. and Walker, R.G. (1975) *Depositional environments as interpreted from primary sedimentary structures and stratification sequences,* pp161. *Lecture notes: Soc. econ. Paleont. Miner., Short Course,* 2, Dallas.

Harms, J.C. Southard, J.B. and Walker, R.G. (1982) *Structure and sequence in clastic rocks. Lecture Notes: Soc. econ. Paleont. Miner., Short Course,* 9.

Hayward, A.B. (1983) Coastal alluvial fans and associated marine facies in the Miocene of S.W. Turkey. In: *Modern and Ancient Fluvial Systems* (Ed. by J.D. Collinson and J. Lewin). *Spec. Publs. int. Ass. Sediment.,* 6, 323-336.

Hein, F.J. and Walker, R.G. (1977) Bar evolution and development of stratification in the gravelly, braided Kicking Horse River, British Columbia. *Can. J. Earth. Sci.,* 14, 562-570.

Heward, A.P. (1978a) Alluvial fan and lacustrine sediments from the Stephanian A and B (La Magdalena, Cinera-Matallana and Sabero) coalfields, northern Spain. *Sedimentology,* 25, 451-488.

Heward, A.P. (1978b) Alluvial fan sequence and megasequence models: with examples from Westphalian D - Stephanian B coalfields, northern Spain. In: *Fluvial Sedimentology* (Ed. by A.D. Miall) *Mem. Can. Soc. Petrol. Geol., Calgary,* 5, 669-702.

Hirst, J.P.P. (1983) *Oligo-Miocene alluvial systems in the northern Ebro Basin, Huesca Province, Spain.* Unpublished Ph.D Thesis, University of Cambridge, 247pp.

Hirst, J.P.P. and Nichols G.J. (1986) Thrust tectonic controls on Miocene alluvial distribution patterns, southern Pyrenees. *Spec. Publs int. Ass. Sediment.,* 8, 247-258.

Hiscott, R.N. and Middleton, G.V. (1979) Depositional mechanics of thick-bedded sandstones at the base of a submarine slope, Tourelle Formation (Lower Ordovician) Quebec, Canada. In: *Geology of continental slopes* (Ed. by L.J. Doyle and O.H. Pilkey). *Spec. Publ. Soc. econ. Paleont. Miner., Tulsa,* 27, 307-326

Hossack, J. (1979) The use of balanced cross-sections in the calculation of orogenic contraction: A review. *J. Geol. soc. London,* 136, 705-711.

Instituto Geologico y Minero de Espana (Huesca 1980 2nd Ed.) *Mapa Geologico de Espana E. 1:200,000.* Servicio de Publicaciones Minesterio de Industria y Energia.

Sheets: 22; 23; 24; 32; 33; 34.

Iwaniw, E. (1984) Lower Cantabrian basin margin deposits in NE Leon, Spain - a model for valley-fill sedimentation in a tectonically active, humid climatic setting. *Sedimentology* 31, 91-110.

James, S. (1984) A tectono-sedimentary study of the Pyrenean Mountain Front near Foix. *Abstr. Tectonics Studies Group Annual Meeting, Swansea,* 42.

Jenny, H. (1941) *Factors of Soil Formation.* McGraw-Hill Book Co., New York.

Jenyon, M.K. (1985) Salt movement. In: *The Role of Evaporites in Hydrocarbon Exploration* (Ed. by J.C.M. Taylor). *Joint Association for Petroleum Exploration courses* -*(UK).*

Joffee, A.F. (1928) *The physics of crystals* (Ed.by L.B. Loeb) McGraw-Hill New York, pp.198.

Jones, P. (1982) Oil and gas beneath east-dipping underthrust faults in the Alberta Foothills. In: *Rocky Mountain Assn. of Geologists. Geol. Study of the Cordilleran Thrust Belt Field Conference. (Denver, 10/24-27/28) Guidebook,* 1, 1-45.

Jordan, E. (1977) *Whistle and sing: songs, airs and dances of Ireland.* 4th Ed. Whistle and Sing, Portadown, N. Ireland. pp.97.

Julivert, M. (1978) The areas of alpine cover folding in the Iberian Meseta (Iberian cahin, Catalanides etc.) In: *Geological Atlas of Alpine Europe and adjoining Alpine areas.* (Ed. by M. Lemoine) Amsterdam, Elsevier. pp.93-112.

Kent, P.E. (1979) The emergent Hormuz salt plugs pf Southern Iran. *J. Petrol. Geol.,* 2, 117-144.

Klappa, C.F. (1978) Biolithogenesis of Microcardium: elucidation. *Sedimentology,* 25, 489-522.

Klappa, C.F. (1980) Rhizoliths in terrestrial carbonates: classification, recognition, genesis and significance. *Sedimentology,* 27, 613-629.

Kochel, R.C. and Johnson, R.A. (1984) Geomorphology and sedimentology of humid-temperate alluvial fans, Central Virginia. In: *Sedimentology of Gravels and conglomeates* (Ed. by E.H. Koster and R.J. Steel). *Mem. Can. Soc. Petrol. Geol., Calgary,* 10, 109-122.

Kraus, M.J. (1984) Sedimentology and tectonic setting of

Early Tertiary quartzite conglomerates, northwest Wyoming. In: *Sedimentology of Gravels and Conglomerates* (Ed. by E.H. Koster and R.J. Steel). *Mem. Can. Soc. Petrol. Geol., Calgary* 10, 203-216.

Kulander, B.R. and Dean, S.L. (1986) Structure and tectonics of central and southern Appalachian Valley and Ridge and Plateau Provinces, West Virginia and Virginia. *Bull. Am. Ass. Petrol. Geol.*, 70, 1674-1684.

Kupfer, D.H. (1968) Relationship of internal to external structure of salt domes. In: *Diapirism and Diapirs* (Ed. by J. Braunstein and G.D. O'Brien). *Spec. Publs. Am. Assoc. Petrol. Geol., Tulsa*, 8, 79-89.

Lemoine, M., Bas, T., Arnaud-Vanneau, A., Arnaud, H., Dumont, T., Gidon, M., Bourbon, M., de Graciansky, P.-C., Rudkiewicz, J.L., Megard Galli, J., and Tricart, P. (1986) The Continental margin of the Meszoic Tethys in the Western Alps. *Marine and Petroleum Geology*, 3, 179-199.

Lemoine, M. (1985) Structuration Jurassique des Alpes occidentales et plainspastique de la Tethys ligure. *Bull. Soc. Geol. Fr., Ser, I*, 1, 126-137.

Lowe, D.R. (1972) Implications of three submarine mass-movement deposits, Cretaceous, Sacramento Valley, California. *J. Sedim. Petrol*, 42, 89-101.

Lowe, D.R. (1979) Sediment gravity flows: their classification and some problems of application to natural flows and deposits. In: *Geology of continental slopes* (Ed. by L.J. Doyle and O.H. Pilkey). *Spec. Publs. Soc. econ. Paleont. Miner., Tulsa*, 21, 75-82

Lowe, D.R. (1982) Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. *J. Sedim. Petrol.*, 52, 279-297.

Lucas, C. (1985) *Le Gres Rouge du versant nord des Pyrenees. Essai sur la geodynamique de depots continentaux du Permian et du Trias.* Ph.D Thesis, Toulouse. pp.265.

Masselink, T. (1965) *De sedimentologie van enkele formaties uit het Boren-Krijt bij Pobla de Segur (Zuid Pyreneeen).* Internal report, Geol. Inst., Univ. Leiden, The Netherlands. pp442.

Mattauer, M. (1968) Les traits structuraux essentiels de la chaine des Pyrenees. *C.R. Acad. Sci. Paris* 300, Ser. II, 71-74.

Mattauer, M. and Henry, J. (1974) The Pyrenees. In:

Mesozoic-Cenozoic Orogenic Belts; data for orogenic studies (Ed. by A.M. Spencer). *Spec. Publ. geol. Soc. London*, Scottish Academic Press. pp.814.

McLean, R.J. and Jerzykiewicz, T (1978) Cyclicity, tectonics and coal: some aspects of fluvial sedimentology in the Brazeau-Paskapoo Formations, Coal Valley area, Alberta, Canada. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 441-468.

Miall, A.D. (1977) A review of the braided river depositional environment. *Earth. Sci. Rev.*, 3, 1-62.

Miall, A.D. (1978) Lithofacies types and vertical profile models in braided river deposits: a summary. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 597-604.

Middleton, G.V. and Hampton, M.A. (1973) Sediment gravity flows: mechanics of flow and deposition. In: *Turbidites and Deep-Water Sedimentation* (Ed. by G.V. Middleton and A.H. Bouma). *Soc. econ. Palaeont. Miner. Pacific Section*, Short Course, 1-38.

Middleton, G.V. and Hampton, M.A. (1976) Subaqueous sediment transport and deposition by sediment gravity flows. In: *Marine Sediment Transport and Environmental Management* (Ed. by D.J. Stanley and D.J.P. Swift). Wiley, New York, 197-298.

Misch, P. (1934) Der Bau der Mittleren Sudpyrenaen. *Abh. Ges, Wiss. Gottingen, Math.-Phys. Kl III Folge*, 12, 1-168.

Montadert, L., de Charpal, O., Roberts, D., Guennoc, P. and Sibuet, J.C. (1979) Northeast Atlantic passive continental margin: rifting and subsidence processes. In: *M. Ewing Series 3: Deep Drilling Results in the Atlantic Ocean: Continental Margins and Paleoenvironment.* (Ed. by M. Talwini, W. Hay and W.B.F. Ryan) pp.154-186. Washington D.C., American Geophysical Union.

Moreløy, C.K. (1987) Lateral and vertical changes of deformation style in the Osen-Roa thrust sheet, Oslo Region. *J. Struct. Geol.*, 9, 345-352.

Muehlberger, W.R. and Clabaugh, P.S. (1968) Internal stucture and petrofabrics of Gulf Coast salt domes. In: *Diapirism and Diapirs* (Ed. by J. Braunstein and G.D. O'Brien). *Spec. Publs. Am. Assoc. Petrol. Geol.*, 8, 90-98.

Mukherji, A.B. (1976) Terminal fans of inland streams in

Sutlej-Yamuna plain, India. *Geomorph.*, 20, 190-204.

Mutti, E., Rosell, J., Guibaudo, G. and Obrador, A. (1975) The Paleogene of the Ager basin and the Upper Cretaceous Aren Sandstone in its type area. In. *The Sedimentary Evolution of the Paleogene South Pyrenean Basin* (Ed. by J. Rosell and C. Puigdefabregas). *Excursion Guidebook IX[th] International Congress, int. Ass. Sediment.* Nice, France, A.

Mutti, E., Remacha, E., Sgavetti, M., Rosell, J., Valloni, R., and Zamorano, M. (1985a) Stratigraphy and Characteristics of the Eocene Hecho Turbidite systems, South-Central Pyrenees. In: *Excursion Guidebook, 6[th] European Regional Meeting, int. Ass. Sediment.* (Ed. M.D.Mila and J. Rosell). *Llerida, Spain*, 521-576.

Mutti, E., Rosell, J., Allen, G., Fonnesu, F. and Sgavetti, M. (1985b) The Eocene Baronia tide dominated delta/shelf-system in the Ager basin. In: *Excursion Guidebook, 6[th] Regional Meeting, int. Ass. Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain, 577-600.

Nagtegaal, P.J.C., Van Vliet. A. and Brouwer, J. (1983) Syntectonic coastal offlap and concurrent turbidite deposition: the Upper Cretaceous Aren sandstone in the South-Central Pyrenees, Spain. *Sediment. Geol.*, 34, 185-218.

Nemec, W. and Muszynski, A. (1982) Volcaniclastic alluvial aprons in the Tertiary of Sofia district (Bulgaria). *Annalles Societatis Geologorum Poloniae*, 52, 239-303.

Nemec, W. and Steel, R.J. (1984) Alluvial and coastal conglomerates: their significance and some comments on gravely mass-flow deposits. In: *Sedimentology of Gravels and Conglomerates* (Ed. by E.H. Koster and R.J. Steel). *Mem. Can. Soc. Petrol. Geol., Calgary*, 10, 1-31.

Nemec, W., Steel, R.J., Porebski, S.J. and Spinnangr, A. (1984) Domba conglomerate, Devonian, Norway: process and lateral variability in a mass flow-dominated, lacustrine fan-delta. In: *Sedimentology of Gravels and Conglomerates* (Ed. by E.H. Koster and R.J. Steel). *Mem. Can. Soc. Petrol. Geol., Calgary*, 10, 295-320

Nichols, G.J. and Ori, G.G. (1983) Sedimentation along the compressive margin of the Ebro basin (NE Spain). Abstr. 4[th] European Regional meeting, int. Assn. Sediment. Split, Yugoslavia, 117-119.

Nichols, G.J. (1987) Syntectonic alluvial fan sedimentation, southern Pyrenees. *Geol. Mag.* 124, 121-133.

Nickel, E. (1982) Alluvial-fan-carbonate facies with evaporites, Eocene Guarga Formation, Southern Pyrenees, Spain. *Sedimentology*, 27, 761-796.

Nijman, W. and Puigdefabregas, C. (1974) Stratigraphical and facies relationship of the Paleogene South Pyrenean Basin. In: *The Sedimentary Evolution of the Paleogene South Pyrenean Basin* (Ed. by J.Rosell and C. Puigdefabregas). *Excursion Guidebook IXth International Congress, int. Ass. Sediment.*, Nice, France, C, p2.

Nijman, W. and Nio, S.D. (1975) The Eocene Montanana Delta. In: *The Sedimentary Evolution of the Paleogene South Pyrenean Basin* (Ed. by J. Rosell and C. Puigdefabregas). *Excursion Guidebook IXth International Congress, int. Ass. Sediment.*, Nice, France, B.

Nijman, W and Puigdefabregas,C. (1978). Coarse grained point bar structure in molasse-type fluvial system, Eocene Castisent sandstone formation, S.Pyrenean basin. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 487-510.

Nio, S.D. and Siegenthaler, C. (1978) *A Lower Eocene estuarine-shelf complex in the Isabena valley.* Comparative Sedimentology Division, Utrecht, 18, 1-44.

Padgett, G. and Ehrlich, R. (1978) An analysis of two tectonically controlled integrated drainage nets of Mid-Carboiferous age in southern west Virginia. In: *Fluvial Sedimentology* (Ed. A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 789-800.

Parish, M. (1984) A structural interpretation of a section of the Gavarnie nappe and its implications for Pyrenean geology. *J. Struct. Geol.*, 6, 247-255.

Parkash, B., Sharma, R.P. and Roy, A.K. (1980) The Siwalik Group (molasse) - sediments shed by collision of continental plates. *Sedim. Geol.*, 25, 127-159.

Parkash, B. Awasthi, A.K. and Gohain, K. (1983) Lithofacies of the Markanda terminal fan, Kurukshetra district, Haryana, India. In: *Modern and Ancient Fluvial Systems* (Ed. by J.D. Collinson and J. Lewin). *Spec. Publs. int. Ass. Sediment.*, 6, 337-344.

Parker, T.J. and McDowell, A.N. (1951) Scale models as a guide to interpretation of salt doming. *Am. Assn. Petrol. Geol. Bull.* 35, 2076-2086.

Peybernes, B. and Souquet, P. (1984) Basement blocks and tecto-sedimentary evolution in the Pyrenees during Mesozoic times. *Geol. Mag.*, 121, 397-405.

Pinet, B., Montadert, L., Curnelle, R., Cazes, M., Marillier, F., Rolet, J., Tomassino, A., Galdeano, A., Patriat, Ph., Brunet, M.F., Olivet, J.L., Schaming, M., Lefort, J.P., Arrieta, A. and Riaza, C. (1987) Crustal thinning on the Aquitaine shelf, Bay of Biscay, from deep seismic data. *Nature*, 325, 513-516.

Platt, J.P. and Leggett, J.K. (1986) Stratal Extension in Thrust Footwalls, Makran Accretionary Prism: Implications for thrust tectonics. *Bull. Am. Ass. Pet. Geol.*, 70, 191-203.

Price, R.A. (1981) The Cordilleran foreland thrust and fold belt in the southern Canadian Rocky Mountains. In: *Thrust and Nappe Tectonics* (Ed. by K.R. McClay and N.J. Price). *Spec. Publs. geol. Soc. London*, 9, 427-448.

Puidefabregas, C.T. (1975) *La Sedimentacion Molasica en la Cuenca de Jaca.* Mono. del Inst. de Estudios Pirenaicos, 104. pp188.

Puigdefabregas, C., Rupke, N.A., and Sole, J. (1975) The sedimentary evolution of the Jaca Basin. In: *The sedimentary Evolution of the Paleogene South Pyrenean Basin* (Ed. by J. Rosell and C. Puigdefabregas). *Excursion Guidebook, IXth International Congress, int. Ass. Sediment.*, Nice, France, C.

Puigdefabregas, C.T. and Vliet, A. van (1978) Meandering stream deposits from the Tertiary of the Southern Pyrenees. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 469-485.

Puigdefabregas, C., Samso, J.M., Serra-Kiel, J. and Tosquella, J. (1985) Facies analysis and faunal assemblages of the Roda Sandstone Formation, Eocene of the Southern Pyrenees. *Abstr. 6th European Regionnal meeting, int. Ass. Sediment.,* Lleida Spain. 639-642.

Puigdefabregas, C., and Souquet, P. (1986) Tectosedimentary cycles and depositional sequences of the Mesozoic and Tertiary from the Pyrenees. In: *The Geological Evolution of the Pyrenees* (Ed. by E. Banda and S.M. Wicham). *Tectonophysics*, 129, 173-204.

Puigdefabregas, C., Munoz, J.A. and Marzo, M. (1986) Thrust belt development in the eastern Pyrenees and related depositional sequences in the foreland basin. In:

Foreland Basins (Ed. by P.A. Allen and P. Homewood). *Spec. Publs int. Ass. Sediment.*, 8, 229-246.

Retallack, G.J. (1983) *Late Eocene and Oligocene Palaeosols from Badlands National Park, South Dakota.* Geol. Soc. Am. Spec. Paper, 193, pp.82.

Retallack, G.J. (1986) Fossil soils as grounds for interpreting long-term controls on ancient rivers. *J. Sedim. Petrol.* 56, 1-18.

Riba, O. (1976a) Syntectonic unconformities of the Alto Cardener, Spanish Pyrenees: a genetic intepretation. *Sediment. Geol.* 15, 213-233.

Riba, O. (1976b) Tectogenese et Sedimentation: deux modeles de discordance syntectoniques pyreneennes. *Bull. Bur. Rech. Geol. min. 2eme ser.*, I, 4 387-405.

Riba Arderiu, O., Reguant Serra, S. and Villena Morales, J. (in prep) Ensayo de sintesis estratigrafica y evolutiva de la cuenca Terciaria del Ebro.

Rosell, J. and Puigdefabregas, C. (1975) *The sedimentary Evolution of the Paleogene South Pyrenean Basin.* Excursion Guidebook XIth Int. Congress int. Ass. Sediment., Nice, France.

Rust, B.R. (1972) Structure and process in a braided river. *Sedimentology*, 18, 221-245.

Rust, B.R. (1975) Fabric and structure in glaciofluvial gravels. In: *Glaciofluvial and Glaciolacustrine Sedimentation* (Ed. by A.V. Jopling and B.C. McDonald). *Spec. Publ. Soc. econ. Paleont. Miner.* 23,238-248.

Rust, B.R. (1978) A classification of alluvial channel systems. In: *Fluvial Sedimentology* (Ed. by A.D. Miall). *Mem. Can. Soc. Petrol. Geol., Calgary*, 5, 187-198.

Schumm, S.A. (1968) Speculations concerning palaeohydrologic controls of terrestrial sedimentation. *Bull. Geol. Soc. Am.*, 79, 1573-1588.

Schumm, S.A. (1976) Episodic erosion: A modification of the geomorphic cycle. In: *Theories of Landform Development* (Ed. by W.N. Melhorn, and R.C. Flemal). *Geomorphology*, SUNY, Binghampton, New York, 69-85.

Schumm S.A. (1977) *The Fluvial System.* John Wiley and Sons, New York. pp338.

Schuum S.A. (1981) Evolution and response of the fluvial system, sedimentologic implications. In: *Recent and* *ancient nonmarine depositional environments: models for exploration* (Ed. by F.G. Etheridge and R.M. Flores). *Spec. Publ. Soc. econ. Paleont. Miner., Tulsa*, 31, 19-29.

Scotese, C.R. and Summerhayes, C.P. (1986) Computer Model of Palaeoclimate Predicts Coastal Upwelling in the Mesozoic and Cenozoic. *Geobyte Summer*, 28-42.

Seguret, M. (1972) *Etude tectonique des nappes et series decolles de la partie centrale du versant sud des Pyrenees; caractere synsedimentaire, role de la compression et de la gravite.* These Fac. Sci., Montpellier, Publ. USTELA, Serie Geol. Struct., v.2, 155pp.

Seguret and Daignieres, M. (1985) Coupes balancees d'echelle crustale des Pyrenees. *C.R. Acad. Sci. Paris, Ser. II*, 301, 341-346.

Seguret, M. and Daignieres, M. (1986) Crustal scale balanced cross-sections of the Pyrenees; discussion. In: *The geological evolution of the Pyrenees* (Ed. by E. Banda and S.M. Wickham). *Tectonophysics*, 129, 308-318.

Seilacher, A. (1967) Bathymetry of trace fossils. *Mar. Geol.*, 413-428.

Seilacher, A. (1978) Use of trace fossil assemblages for recognising depositional environments. In: *Trace Fossil Concepts* (Ed. by P.B. Basan). *Soc. econ. Paleont. Miner., Short Course*, 5, Oklahoma.

Selzer, G. (1934) Geologie der subpyrenaishen Sierren in Oberaragonien. *N. Jb. Min. Geol. Palaont.*, 71B, 370-406.

Sgavetti, M., Muttti, E., Rosell, J. and Legarreta, L. (1984) Tectonically controlled Depositional Sequences in the Upper Cretaceous Aren Sandstone, South-Central Pyrenees, Spain. *Abstr. 5th int. Ass. Sediment. European Regional Meeting*, 406-407.

Shell (1975) Unpublished map of the the South Pyrenean basin.

Simo, A. and Puigdefabregas, C. and Gili, E. (1985) Transition from shelf to basin on an active slope, Upper Cretaceous, Tremp area, Southern Pyrenees. In: *Excursion Guidebook, 6th European Regional Meeting, int. Ass Sediment.* (Ed. by M.D. Mila and J. Rosell). Llerida, Spain, 63-108.

Simo, A. (1986) Carbonate Platform depositional sequences, Upper Cretaceous, South Central Pyrenees (Spain). In: *The Geological Evolution of the Pyrenees* (Ed. by E. Banda and S.M. Wickham). *Tectonophysics,* 129, 205-231.

SNPA (1972) Carte Geologique des Pyrenees. *Societe Nationale des Petroles d'Aquitaine.*

Sole-Sugranes, L. (1972) Nota sobre una discordacia en el Eoceno medio del Prepirineo Oriental. *Acta geol. hispanica,* 7, 1-6.

Sole-Sugranes, L. (1978) Gravity and compressive nappes in the central southern Pyrenees (Spain). *Am. J. Sci.,* 278, 609-637.

Souquet, P. (1967) *Le Cretace Superieur sud-pyreneen en Catalogne, Aragon et Navarre.* These, Faculte des Science de Toulouse. pp.529.

Souquet, P. (1984) Les cycles majeurs du Cretace de la palaeomarge iberique dans les Pyrenees. *Strata., Univ. Toulouse III,* 1, 47-70.

Speksnijder, A. (1985) Anatomy of a strike-slip controlled sedimentary basin, Permian of the southern Pyrenees, Spain. *Sedim. Geol.,* 44, 179-224.

Stapp, R.W. (1967) Relationship of lower Cretaceous depositional environment of oil accumulation, northeastern Powder River Basin, Wyoming. *Am. Assoc. Pet. Geol. Bull.,* 51, 2044-2055.

Steel, R.J. (1974) New Red Sandstone flood plain and piedmont sedimentation in the Hebridean Province Scotland. *J. Sedim. Petrol.,* 44, 336-357.

Steel, R.J. and Thompson, D.B. (1983) Structures and textures in Triassic braided stream conglomerates ("Bunter" Pebble Beds) in the Sherwood Sandstone Group, North Staffordshire, England. *Sedimentology,* 30, 341-367.

Surlyk, F. (1978) Submarine fan sedimentation along fault scarps on tilted fault blocks (Jurassic-Cretaceous boundary. East Greenland) *Bull. Gronlands. geol. Unders.,* 128, 1-108.

Talbot, C.J. (1979) Fold trains in a glacier of salt in southern Iran. *J. Struc. Geol.,* 1, 5-18.

Taylor, J.C.M. (1984) Late Permian - Zechstein. In: *Introduction to the Petroleum Geology of the North Sea* (Ed. by K.W. Glennie). Blackwell Scientific Publications, 61-83.

Thode, H.C., Kleerskoper, H. and McElcheran, D. (1951) Isotope fractionation in the bacterial reduction of sulphate. *Research, London*, **4**, 581-582.

Thode, H.C., Wanless, R.K. and Wallouch, R. (1953) The orgin of native sulphur deposits from isotope fractionation studies. *Geochim. et Cosmochim. Acta*, 5, 286-298.

Townsend, C., Roberts, D., Rice, A.H.N., Gayer, R.A. (1986) The Gaissa Nappe Finnmark, north Norway: an example of a deeply eroded external imbricate zone within the Scandinavian Caledonides. *J. Struct. Geol.*, **8**, 431-440.

Trenchmann, C.T. (1955) *The British Association for the Suppression of Science; or, A new explanation of mountain uplift based on lunar gravitation and oceanic pressures.* Ord and Printing: West Hartlepool. pp64.

Truc, G. (1978) Lacustrine Sedimentation in an evaporitic environment: the Ludian (Paleogene) of the Mormoiron Basin, southeastern France. In: *Modern and Ancient Lake Sediments* (Ed. by A. Matter and M.E. Tucker). *Spec. Publs int. Ass. Sediment.*, 2, 187-202.

Trusheim, F. (1960) Mechanism of salt migration in Northern Germany. *Bull. Am. Assoc. Petrol. Geol.*, **44**, 1519-1540.

Tunbridge, I.P. (1981) Sandy high-energy flood sedimentation - some criteria for recognition, with an example from the Devonian of S.W. England. *Sedim. Geol.*, 28, 79-95.

Turner, P., Hirst, J.P.P. and Friend, P.F. (1984) A palaeomagnetic analysis of Miocene sediments at Pertusa, near Huesca, Ebro Basin, Spain. *Geol. Mag.*, **121**, 279-290.

U.S.D.A. (United States Department of Agriculture; 1975) *Soil taxonomy. A basic system of soil classification for making and interpreting soil surveys.* Castle and sons Publications Ltd. Tunbridge Wells, Kent pp754.

van Bemmelen, R.W. (1955) Tectogenese par gravite. *Ann. Soc. Geol. Belg.* **64**, 95-123.

van Buchem, F.S.P., van den Hoek Ostende, H., Oude Nijhuis, A., Ziogas, A., van Gelder and de Boer, P.L.. (1985) Tectonic control of sedimentation in the Eocene Sobrabe basin, S. Pyrenees. *Abstr. 6th European Regional Meeting, int. Ass. Sediment.* Lleida, Spain, 527-530.

van Hoorn, B. (1970) Sedimentology and Paleogeography of an Upper Cretaceous Turbidite Basin in the South Central Pyrenees, Spain. *Leidse Geologische Mededelingen,* 45, 73-154.

Verschure, V.H., Hebeda, A., Boelrijk, O. Priem, C.J. and Ave Lallement, H.G. (1967) K-Ar age of a hornblende from a hornblendite vein in the alpine type ultramafic mass of the Etang de Lhers (Ariege) French Pyrenees. *Leidse Geol. Meded.,* **42**, 59-60.

Washington, P.A. and Moreley; and Root, S.I. and Moreley, C.K. (1987) A classification of thrust fronts: Discussions and replys. *Bull. Am. Ass. Petrol. Geol.,* **71**, 737-744.

Wasson, R.J. (1977) Last-glacial alluvial fan sedimentation in the Lower Derwent Valley, Tasmania. *Sedimentology,* **24**, 781-799.

Wasson, R.J. (1979) Sedimentation history of the Mundi-Mundi alluvial fans, western New South Wales. *Sedimentary Geology,* 22, 21-51.

Watts, N.L. (1980) Quaternary pedogenic calcretes from the Kalahari (Southern Africa): mineralogy, genesis and diagenesis. *Sedimentology,* 27, 661-686.

Weaver, C.E.and Beck, K.C. (1977) Miocene of the S.E. United States: a model for chemical sedimentation in a peri-marine environment. *Sedim. Geol.,* **17**, 1-234.

Weimer, R.J. (1984) Relation of Unconformities, Tectonics and Sea Level Changes, Cretaceous of Western Interior, U.S.A. In: *Interregional Unconformities and Hydrocarbon Accumulation* (Ed. by J.S. Schlee). *Mem. Am. Ass. petrol. Geol.* **36**, 7-35.

Weimer, R.J. (1986) Relationship of Unconformities, Tectonics and Sea Level Changes in the Cretaceous of the Western Interior, United States. In: *Paleotectonics and Sedimentation in the Rocky Mountain Region, United States* (Ed. by J.A. Peterson). *Mem. Am. Ass. petrol. Geol.* **41**, 397-422.

Wendlant, E.A., Shelby, T.H. Jr. and Bell, J.S. (1946) Hawkins field, Wood County, Texas. *Am. Ass. petrol. Geol. Bull.* **30**, 1830-1856.

Wentworth, C.K. (1922) A scale of grade and class terms for clastic sediments. *J. Geol.,* **30**, 377-392.

Weirsma, D.J. (1965) *Sedimentologiese aspekten van het Boven krijt bij Pobla de Segur (Zuid Pyreneeen).*

Internal report, Geol. Inst., Univ. Leiden, The Netherlands. pp83.

Williams, G.D. and Fisher, M.W. (1984) A balanced section across the Pyrenean orogenic belt. *Tectonics*, 3, 773-780.

Williams, G.D. (1985) Thrust tectonics in the south central Pyrenees. *J. of Struct. Geol.* 7, 11-17.

Wright, V.P. (1982) Calcrete paleosols from the Lower Carboniferous, Llanelly Formation, South Wales. *Sedim. Geol.*, 33, 1-33.

Wright, V.P. (1986) In: *Paleosols* (Ed. by V.P Wright). Blackwell Scientific Publications, Oxford.

Zwart, H.J. (1979) The Geology of the Central Pyrenees. *Leid. Geol. Meded.* 50, 1-74.

A

YERTIZ!

PUBLICATION

original artwork by

W. Coleman