GEOLOGICA ULTRAIECTINA

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## A fossil record in the Monllobat and adjacent Castigaleu Formations, in the drainage area of the present Rio Noguerra Ribagorzana, provinces of Huesca and Lérida, Spain

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# DE VERBREIDING VAN PYRENEES EROSIEMATERIAAL, AFGEZET DOOR EOCENE VLAKTEVLOED SYSTEMEN EN BIJBEHORENDE PUINWAAIER DELTA'S:

## Een fossiele registratie in de Monllobat en naastgelegen Castigaleu Formaties in het stroomgebied van de tegenwoordige Rio Noguerra Ribagorzana in de provincies Huesca en Lérida, Spanje

(met een samenvatting in het Nederlands)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD VAN DOCTOR
AAN DE RIJKSUNIVERSITEIT TE UTRECHT,
OP GEZAG VAN DE RECTOR MAGNIFICUS PROF. DR. J.A. VAN GINKEL
INGEVOLGE HET BESLUIT VAN HET COLLEGE VAN DEKANEN
IN HET OPENBAAR TE VERDEDIGEN
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door

SJOERD VAN DER MEULEN

geboren op 29 mei 1954 te Donkerbroek

PROMOTOR: PROF. DR. S.D. NIO
CO-PROMOTOR: PROF. DR. W. SCHLAGER

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Contents

Chapters 2-5 are publications of the author (as sole writer), which appeared earlier in geological journals.

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# SAMENVATTING

Dit proefschrift omvat vier artikelen over de sedimentologische opbouw, op grote zowel als op kleine schaal, van de Eocene Monllobat en Castigaleu Formaties in de Spaanse Pyreneeën. De betreffende artikelen worden voorafgegaan en afgesloten door respectievelijk een uitgebreide introduktie en door samenvattingen van de afzonderlijke artikelen, aangevuld met toepasselijke referenties van literatuur, die nog niet in de artikelen vermeld kon worden.

De beide bestudeerde Formaties bestaan uit continentale, respectievelijk (brak-) mariene afzettingen. Het uiterlijk van de Formaties wordt beheerst door plaatvormige lagen met een erg geringe dikte (maximaal 10 m.), vergeleken met de uitbreiding (100-den m!s - km!s). Het gehalte aan fijn materiaal is erg hoog (70-80% silt en klei), waarbij de plaatvormige habitus van de Monllobat Formatie nog eens onderstreept wordt door de regelmatige inschakeling van dikke caliche lagen.

Het onderzoek richtte zich in de eerste plaats op de samenhang tussen de grove en de fijne lagen. In de eerste plaats werd gekeken naar de verbinding met meanderende rivierafzettingen in dit gebied het meest duidelijk vertegenwoordigd door een in drie dimensies ontsloten binnenbocht afzetting. Het bovenvlak van deze sedimentlob vertoont heel duidelijk eèn stapsgewijs bijplaatsingspatroon, terwijl de restgeul door selektieve verwijdering van de fijne opvulling prachtig is uitgeprepareerd. Na vergelijking met de facies in verticale dwarsdoorsnedes van binnenbochtafzettingen kon daarom een erg gedètailleerd sedimentatie model opgesteld worden. Vergelijkingen van andere lagen met de kenmerken van dit model, samen met laterale laagvervolging, leverde overigens wel op, dat er erg weinig fijn materiaal met meanderende rivierafzettingen verbonden was, vooral ook omdat er over het geheel genomen erg weinig meanderende rivierafzettingen aanwezig bleken te zijn. Veel voorkomend zijn daarentegen juist verbindingen van vlakke fijne lagen met de top van symmetrische grind- en/of zandgeulopvullingen. Zowel deze symmetrische, als de asymmetrische, meanderende geulen bleken verder in het distale, verfijnende deel van de uitgestrekte, massieve conglomeraatplaten te liggen, die het skelet van de Monllobat Formatie vormen.

De afzetting van veel fijn materiaal direkt distaal van dikke conglomeraten is het duidelijkst geïllustreerd in de tweede gedètailleerde, ruimtelijke studie, waarin een conglomeratische puinwaaierdelta (fan-delta) over de voormalige kustlijn overgaat in zandsteen en silt-/kleisteen lagen. Vanwege de sterke aggradatie is het fijne materiaal zowel in het brakke, staande water, als op het land afgezet (Hjullströmtype delta). De afzettingen in dit voorbeeld blijken daarbij de typische strukturen van de Castigaleu zandsteenlagen te vertonen.

Het skelet van de Monllobat Formatie wordt gevormd door massieve, kilometersbrede conglomeraatplaten, zoals bleek uit grootschalige kartering en sectieopname in een gebied van ongeveer 10 x 15 km$^2$. De conglomeraatplaten liggen over het algemeen precies binnen de verticale grenzen van een achttal compartimenten, die vanaf de noordrand maximaal over driekwart van de bekkenbreedte te vervolgen zijn. Voornamelijk in het distale deel treedt binnen de platen een sterke gradering op samen met een loodrechte afbuiging van de paleostroomrichting. Verder zuidelijk ligt een smalle zone met het hoogste klei/silt gehalte, die de compartimenten van de zuidelijke bekken rand scheidt. Hier liggen kleine lobben met zuidelijke aanvoerrichtingen. Ook de stroomrichtingen in dit gebied buigen loodrecht af, in dit geval dus van noord naar west, terwijl in het noordelijke en centrale deel de afbuiging van zuidwest naar noordwest gaat. Dit onderstreept de duidelijk gescheiden ontwikkeling van de twee gebieden.

Tussen de top en basis van de Monllobat Formatie zijn drie hoofd- en vier subhorizons onderscheiden, waarop de conglomeraatplaten steeds liggen. Dunne lagen mariene mergel aan de basis van deze horizons dringen vanuit de Castigaleu Formatie over aanzienlijke afstanden (kilometers) de Monllobat Formatie binnen.

De acht noordelijke compartimenten zijn gevormd door acht individuele alluviale systemen, waarin uitgestrekte vlaktevloeden (sheetfloods) met grind, zand en fijner materiaal domineerden. Over de zuidelijke bekkenrand, een drempel, was er ook enige zuidelijke aanvoer. Distaal bogen de afgeremde vloeden haaks af in de richting van bekkenverbreding.

Het voorkomen van de mariene mergel - conglomeraatplaat horizons in de Monllobat Formatie werd veroorzaakt door de combinatie van versnelde bekkenbodemdaling - waardoor direkt een transgressie optrad - en versnelde achterland opheffing - waardoor enige tijd later extra grof en extra veel erosiemateriaal vrijkwam.

Het randgebied van de Castigaleu Formatie langs de Monllobat Formatie wordt gedomineerd door mergel en door zandsteenlagen met alleen in de laatstgenoemden fragmenten van oesterfossielen. De zandsteen facies met tabulaire grootschalige scheve gelaagdheid en parallelle laminatie, en lobben komt overeen met die van het in drie dimensies ontsloten détail voorbeeld van een puinwaaierdelta (fan-delta). Instroming van de vlaktevloeden in het ondiepe, brakmariene milieu zal daarom ook het onderzochte deel van de Castigaleu Formatie gevormd hebben.

De sterke overeenkomst tussen breuk- en diaklaasrichtingverspreiding en de verspreiding van de stroomrichtingen wijst op een bepaling van de afvoergradienten door beweging van grootschalige breukblokken. De distale haakse afbuiging tijdens afnemende stroomsnelheid wordt ook verklaard door de overgang van een richting dwars op de breukblokas (die in de richting van de bekkenas ligt) naar een richting evenwijdig aan die as (met een duidelijk lagere gradient). Synsedimentaire breukwerking op grotere diepte verklaart daarbij het voorkomen van sterke afwijkingen in facies, dikte en stroomrichtingen nabij breuken. Lens- en wigvormige lagen langs breuken (in het eerste geval vooral meanderende rivierafzettingen), een grotere synclinale met fan-deltas langs de randen, en lateraal uitgebreide facies verschillen aan weerszijden van het breukvlak (waarbij zelfs systeemgrenzen verschuiven nabij een van de grootste breuken) werden gevormd ten gevolge van deze breukwerking in de ondergrond. De meeste breuken hebben zich later door de sedimenten naar boven toe voortgeplant, onafhankelijk van de begrenzingen van de eerder gevormde anomalieën in de bovengrond.

Het grootste deel van het grind en zand bestaat uit Mesozoische micritische kalk, wat korte transportafstanden in het toenmalige milieu aangeeft. Variatie van het kwartsgehalte is daarbij een indicatie voor de mate waarin granitisch materiaal uit de kern van het achterland toegevoegd werd aan het carbonatische afbraakmateriaal van de gebergterand.

Recentere literatuur dan verwerkt in de artikelen in de hoofdstukken 2 - 5 onderbouwt de gemaakte interpretaties verder. Vlaktevloeden (sheetfloods) hebben (sub-)recent op puinwaaiers in zuidoost Spanje gelijksoortige facies gevormd als de Monllobat conglomeraat- en klei-/siltsteenplaten. Het fijnste materiaal ontstaat in grote massa's door bodemvorming in het bergachtige achterland van het afzettingssyteem. Het sedimentgehalte van de vlaktevloeden ligt tussen die van gewone stromen en modderstromen, waardoor relatief hoogenergetische, ondiepe, maar erg brede stromen kunnen bestaan. Grootschalige scheve gelaagdheid en parallelle laminatie zijn beschreven uit instromingsmilieu's van glaciale puinwaaierdelta's en worden gevormd door hoogturbulente restanten van vlaktevloeden, die aan de kustlijn veel van het vooral grofste materiaal al in lobben afgezet hadden.

## chapter I

# INTRODUCTION

Field research on the Eocene Monllobat and Castigaleu Formations in the Southern Pyrenees, specifically along the main and tributary valleys of the Rio Noguerra Ribagorzana, constitutes the basis of this Ph D study. The fieldwork was performed over a period of eight months in 1980, 1981 and 1982 - after a fieldwork period of four months in the scope of a M Sc thesis(Van der Meulen 1978) in 1975 and 1976.

Before the latter period three main elements of the sedimentary environment had been identified during previous, preliminary research (survey in Nijman & Nio, 1975), which distinguished the ancient Montañana delta (named after the village in the centre of the study area). The continental Monllobat Formation, the lower, most eastern delta part in the elongated Tremp-Graus Basin, was assumed to have been built by: 1 Meandering rivers, discharging along the basin axis from the east-southeast and especially characterized by a thick fine member. 2 Minor conglomerate influxes were pictured to enter the Basin perpendicular to the axis from the rising part of the Pyrenees (to the northeast). The adjacent more or less marine Castigaleu environment comprised element 3 Fluviomarine sedimentation of the meandering rivers, with a minor tidal influence out of the northwest-west.

During the M Sc research at least 75 % of the coarse members in the northeastern and central parts of the Monllobat Formation turned out to have been supplied from the northeast, perpendicular to the basin axis. Furthermore, these conglomeratic deposits of braided distal fan streams were connected with fines, which consequently were not solely deposited by meandering rivers, as postulated by Nijman & Nio (1975) after the generally applied model for alluvial fines deposition at that time (only 'minor conglomeratic influxes' were pictured in the palaeogeographical reconstruction of the subaerial area).

The following Ph D research consisted of two parts. The first half of the project was entitled: 'Investigation of the fine sediments in the distal and marginal zones of alluvial fan associations and the relation with river systems'. Three-dimensional exposures with surfaces of several hectares were studied, in which the relation of the former flood basin fines with coarse members is exceptionally well developed. Chapter 2 (publication of 1982) incorporates such a combination of fines and the point bar deposits along a meandering palaeochannel (reconstructed in great detail). Chapter 3 (publication of 1983) concerns the connection of fines with conglomeratic layers, deposited under sheetflood conditions in a fan-deltaic environment. In chapter 2 a first attempt is made to reconstruct the complete alluvial system. Occasional meandering rivers were pictured distally from graveliferous sheetflood plains, with marginal and distal sheetflood deposition of fines.

The second half of the project entitled : 'Deposits of Eocene mass transport systems along the northern margin of the Tremp-Graus Basin' involved large-scale mapping of the distributions of coarse members and sedimentary facies in the Monllobat and adjacent Castigaleu Formations. The emphasis was to be put on sediments originating from the rising Pyrenees to the northeast

The Monllobat Formation and a bordering zone of the Castigaleu Formation are very well exposed, base and top of the deposits are marked by regressions over at least 10 km, and the dip of the layers is mostly only a few degrees. These characteristics present major opportunities to trace the maximally 180 m thick deposits over kilometre-wide distances along the valley walls in the Ribagorzana drainage area.

In chapter 4 (publication of april 1986) several kilometre-wide cross-sections along valley walls present the distribution of the major coarse members of the Monllobat framework. These coarse members occur at distinct levels. Anomalous coarse member thickenings (s.l.) occur at fault sites. Chapter 5 (publication of june 1986) surveys the sedimentary facies, as recorded in logs at regular distances in the Formations.

It could be concluded, that northeasterly supplied sediment occupies about 75 % of the basin width, separated by a distal mudzone from a southerly supplied fringe of coarse member lenses along the southern basin margin. Conglomerate sheets contain the bulk of the northeasterly supplied coarse material. Foreset lobes of several types, symmetrical and scarce asymmetrical channel fills and occasional flat beds are the other coarse member types. Sheets lie at distinct levels within the Formations. Monllobat compartments are defined by the fixed positions of sheets, succeeding each other in vertical sequences. The eight compartments recognized in total, enclose the area of eight Eocene sheetflood systems on a bajada. Besides marginal mudzones the systems possess a distal mudzone and coarse tail parallel to the basin axis (in the continental area) or continue in the restricted marine Castigaleu environment. The Castigaleu coarse members comprise various types of lobes and flat layers, deposited by sheetflood outflow in standing water. No indications have been found for marine physical processes and of alluvial systems, discharging along the basin axis.

Synsedimentary subsurface faulting is expressed by anomalous coarse member development along several zones. The zones are often marked by steeply-inclined faults, with metres to tens-of-metres displacements, which all postdate coarse member sedimentation. Fault and joint directions, measured in order to obtain a more detailed picture of the basin floor geometry, indicate a direct relation with palaeocurrent directions and thereby with palaeogradients.

Quartz content in sandstone thin sections has been identified as the primary parameter in connection with hinterland and environmental reconstruction. As in the gravel fraction, extraclasts of Mesozoic micritic limestone predominate in the thin sections. However some of the samples contain also large amounts of homogeneous quartz and alkalic feldspar. In such sediments gravel is always relatively less abundant. These latter two features point to an inverse relation of the supply of Axial Pyrenean granite debris with the release of gravel from the Mesozoic limestone relief more close to the Basin. Obviously lowering of the limestone relief promoted the supply of axial, arkosic material. The gradual disappearrance of gravel in the western area adjacent to the arkosic San Esteban fan supports this interpretation. The sandstones underneath the base of the Monllobat Formation are arkosic, which is against the overall trend as expected during denudation of a mountain chain, and must be linked with the details of the Pyrenean orogenesis.

Results of three Ph D studies performed in the region have not been mentionned in chapters 2 - 5, because the publications presented here had gone to press before completion of this work (Atkinson, 1984, Fonnesu, 1984, and Cuevas Gozalo, 1985 a and b). Atkinson (1984) studied alluvial deposits all through the Basin with only a few Monllobat examples. Fonnesu (1984) studied the formations underneath the Monllobat and laterally equivalent Castigaleu Formations. This up to 1 km-thick sequence, exposed along the eastern margin of the Basin, is interpreted to be an association of tidally -reworked delta-front deposits in the lower parts and a regressive, braided sequence in the upper parts. Two major braided river deltas with S-SE and S-SW progradation directions are pictured. Apparently the feeder canyons of the eight Monllobat sheetflood systems in the study area originated after the major regression at the base of the Monllobat Formation. Nevertheless also in Monllobat times there remained a major fan-delta, built by the San Esteban alluvial fan to the west of the study area (Nijman & Nio, 1975). The progradation of the San Esteban fan surpassed the Monllobat bajada building and consequently this bajada received some arkosic San Esteban incursions (ch. 4, fig. 5).

The Castigaleu Formation of Nijman & Nio (1975) comprises both the top of Fonnesu's Figols sequence and the marine lateral equivalent of the Monllobat Formation. Restriction of the name Castigaleu Formation to the lateral equivalent of the Monllobat Formation is thought to be preferable, because there is a major, flat regression plane at the base and the palaeocurrents and sandstone mineralogies are significantly different below and above this plane.

In an area some 20 km to the west Cuevas Gozalo (1985 a and b) investigated the Capella Formation, which, together with the Castisent Formation, overlies the Monllobat Formation over a major regression plane. A distal alluvial fan environment with pronounced tidal influence has been suggested. Some reevaluation of these publications is considered necessary, even though Cuevas Gozalo does not present a direct comparison with the results on distal fan - fan-delta deposits in chapter 3 and only in a minor way with the meandering river deposits in chapter 2. The reasons for this consideration are the close resemblances of the coarse member geometries in both Formations and the assumption by Leo & Allen (1984), that an example of Monllobat meandering river deposits also reflects a tidal influence (discussion in ch. 5, p. 88). As in the latter case the evidence for tidal action is not very strong. The marine counterpart with completely tidal deposits is not described and neither are (restricted) marine fossils as oyster shells. Some vertical burrowing is thought to indicate intertidal conditions, but is also known from lacustrine environments (Bromley & Asgaard, 1979). Countercurrent and flood directions in crossbedding, on foreset lobes and in meandering river deposits respectively, do also not necessarily indicate a tidal influence. An alternative explanation for countercurrents on lobes is given in Chapter 3 (p. 52), and ebb tides can enforce fluvial crossbed building, however the main effects of flood tides are deceleration and obstruction of the alluvial flow (Leo & Allen, 1984).

In this consideration it should be noted, that the major regression plane in between the two Formations, differences in internal structures of the two coarse member associations, and the lower gravel contents of the Capella types may leave room for other interpretations. Finally it is remarked that former tidal influences on sedimentation may be largely obscured by the effects of other marine, and especially alluvial currents. However alluvial flow in the distal alluvial fan - fan-delta environment is highly energetic to an extent that tidal action can only rework sediments.

The approach to this study has been to record sedimentary facies, not only in vertical, but explicitly also in lateral sections. The construction of spatial models, based on three-dimensional exposures, on a scale of hundreds of metres as well as several kilometres, has been one of the main targets. The imposed task, but also the degree of exposure, and the type of sedimentary facies (uniform vertical facies development in most layers) are the reasons for this.

Miall (1985), after publishing major surveys of fluvial facies models has advocated the study of the spatial development of facies, because of some inadequassies of the vertical facies models. Furthermore, a more direct link with the build up of sediment bodies is established and the practical utility of the studies increases. This can also be inferred for the study presented here. With regards to the utility: Examination of coarse - fine member distribution in several-kilometre-wide field sections usually meets with major difficulties and high expense (Flores & Ethridge, 1981). These authors therefore recommend the application of computer simulation models for alluvial architecture. However only meandering river environments have so far been involved in such models.

# chapter II

## THE SEDIMENTARY FACIES AND SETTING OF EOCENE POINT BAR DEPOSITS, MONLLOBAT FORMATION, SOUTHERN PYRENEES, SPAIN[1]

### II. 1. INTRODUCTION

Detailed descriptions are given of point bar deposits, whose general characteristics have been shown to resemble the features of ancient point bar deposits (scoured base, upward decrease in grain-size and energy of stratification types in combination with arcuate point bar geometry). PUIGDEFABREGAS & VAN VLIET (1978) pointed out this resemblance in a general article on point bar deposits in the S Pyrenees.

This paper describes and reconstructs a meander lobe and the morphology of the surrounding palaeochannel in three dimensions. In this channel a small part of the original, fine fill is present. In addition the vertical section of another meander lobe reveals sedimentary structures and grain-size distributions.

The terminology of NANSON (1980) for the subenvironments of the meandering river is used. The *point bar* is the sediment body against the convex bank of a river bend. The *meander lobe* is the total alluvial land-form within a meander loop and is used in a three-dimensional sense. The *flood plain* is the area adjacent to the river channel, which is subject to periodic flooding. Point bar and flood plain deposits constitute in the main the coarse and the fine member respectively of the meandering river environment. During the lateral migration of a meander the succession of point bar layers constitutes Epsilon Cross Stratification – E C S (ALLEN, 1963).

The purpose of this paper is to establish a three-dimensional meander lobe model which can be used to locate exposed point bar deposits. The interpretation of the origin of individual point bar layers and the arrangment of these layers are important elements of the model.

On the basis of a general sedimentological description a reconstruction has been made of the factors that determine the environment on a large scale. Finally the relationships between point bar sedimentation and general environmental conditions are investigated.

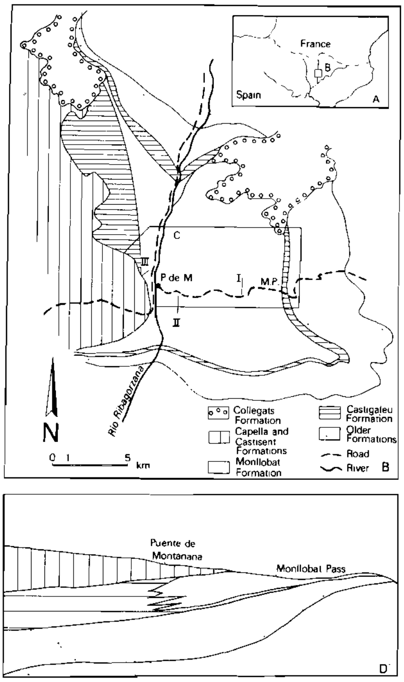


Fig. 1.
Location map.
A Situation of the study area in the S Pyrenees.
B The geological map with the locations of the sections and the outline of the study area (C).
D The vertical relation of the formations.

## II.2. GEOLOGY AND TOPOGRAPHY

The Monllobat Formation belongs to the lower part of the Eocene Montañana Group, which partly constitutes the fill of the Tremp-Graus basin of the Southern Pyrenees (NIJMAN & NIO, 1975). In this elongated basin – during the Eocene – continental depositional environments passed into marine environments in a W to NW direction. In this sense the Monllobat Formation passes laterally into the marine Castigaleu Formation (Fig. 1). The latter is also present with varying thickness at the base of the Monllobat Formation. The formations are concordantly overlain by the Castisent and Capella Formations. The Oligocene Collegats Formation discordantly overlies the Montañana Group.

Previous work on the Montañana Group was carried out by VAN EDEN (1970), NIJMAN & NIO (1975) and NIJMAN (1981). Fluvial deposits were studied by NIJMAN & PUIGDEFABREGAS (1978) and PUIGDEFABREGAS & VAN VLIET (1978).

The 170 m thick, Monllobat Formation is defined by the occurrence of multicoloured siltstone and mudstone with intercalated sandstone and conglomerate bodies (NIJMAN & NIO, 1975). The multicolouring and the frequently appearing carbonate nodules (caliche) are considered to be the products of palaeopedogenesis. The coarse material has been deposited in meandering and braided rivers (NIJMAN & NIO, 1975).

The Monllobat Formation is mainly exposed in the regions around Puente de Montañana. The results presented here are based on a study of the area outlined in Fig. 1. The outcrop conditions are good, there is a low tectonic dip and only a minor tectonic disturbance. Stratigraphic levels can be traced over distances of several kilometres. The study area is incised by several mountain streams. Individual vertical sections along valley walls attain lengths of 60-150 m.

## II.3. GENERAL SEDIMENTOLOGICAL DESCRIPTION

The point bar deposits that were studied in detail are outlined in Fig. 2. From sections I, via II to III, i.e. from E to W, (locations in Fig. 1) the conglomerate content decreases, whereas point bar deposits and intercalations of the Castigaleu Formation become increasingly important. Section I is located in an E-W valley just N of the road from the Monllobat Pass to Puente de Montañana (Fig. 1). In the

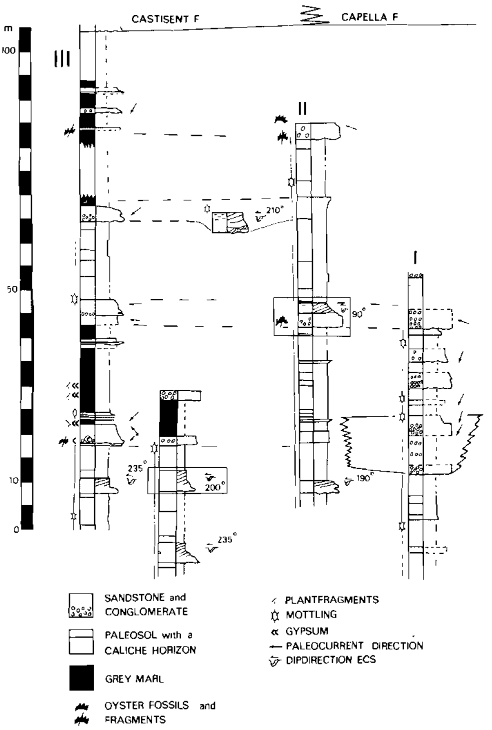


Fig. 2.
Sections through the Monllobat Formation (location in Fig. 1).

Fig. 3.
Upper part of the general section II above the meander lobe exposure. Nodular carbonate horizons stand out in the flat bedded sediment.

vicinity of this section coarse lithosomes, several kilometres wide and up to 10 m thick, are embedded in a matrix of fines. The cores of these bodies consist of massive or flatbedded conglomerates. In the margins (conglomeratic) sandstone is the major constituent. Several lithosomes of this type can be juxtaposed (Fig. 2). In S and W directions transitions towards flatbedded, fine sediment with occasional conglomerate lenses and towards layers with E C S can be observed.

In section II the extensive, coarse layers with oysters are assigned to the Castigaleu Formation. Coarse, fluviomarine deposits and grey marls constitute a major part of section III. In the area examined, deposits of the Castigaleu Formation show hardly evidence of reworking by tidal or wave action.

Apart from coarser deposits the Monllobat Formation contains 70-80% fine material (Fig. 3). The flat beds are extensive with a uniform thickness (dm-m). In the basically brown to ochre coloured beds vertical mottles are blue and occasionally blue-red or red. Within the beds the colour does not vary. Caliche appears as concretions or as a cement in fine sediment that has no clay fraction. The caliche is mainly confined to horizons that can be traced over large areas.

### II.3.1. *Palaeocurrent directions*

Conglomerate layers in general show SW palaeocurrent directions, perpendicular to the basin axis. In most conglomeratic and (very) coarse-grained sandstones directions range from SW to NW, while deposits with ECS have palaeocurrent directions to the NW, following the plunge of the basin axis. This palaeocurrent distribution agrees with the general pattern in the basin (NIJMAN & NIO, 1975).

### II.3.2. *Petrography*

Conglomeratic components consist of dark-blue or grey, micritic limestone and a minor amount of white vein quartz. There are some granitic pebbles. Sandstones often contain more than 90% detritic limestone grains with low percentages of quartz, chert and alkalic feldspar grains.

## II.4. DESCRIPTION OF TWO LARGE EXPOSURES OF POINT BAR DEPOSITS

### II.4.1. *Vertical section of a meander lobe*

Location: The exposure is situated in a valley wall above the Puente de Montañana-Aren road at the bifurcation towards Montañana (Figs 1 and 2, section III). Section III is situated in a shallow syncline with a high S/Sh ratio.

Setting: In the valley there is a 600 m wide section perpendicular to the accretion planes. Fig. 4 shows the excellent exposure of the E part of the section. At the base there is a mottled, flat-bedded layer with some small conglomerate lenses. The major part of the exposure is covered by multicoloured mudstone, but at the NE margin grey coloured channel fills and a conglomeratic layer are present.

Description: The initial part at the NE margin is relatively thick and coarse. The layers consist of (very) coarse sandstone with conglomerates and slumped mottled material. The succeeding, finer layers make up the major part of the exposure. With the fining the dip of the point bar layers increases. The dip direction is generally to the SW. At the SW margin there is a fine channel fill. A cut bank is also exposed (see PUIGDEFABREGAS & VAN VLIET, 1978, figure 8).

The point bar profile underneath the fine fill is not simply sigmoidal. A main channel is associated with a small channel cut in the top of the point bar sequence. The exposure at the SW margin is considered to contain one single point bar sequence instead of two separate layers as described by PUIGDEFABREGAS & VAN VLIET, 1978 (figure 5A). The base consists of large SW dipping foresets. Flat tabular sets in

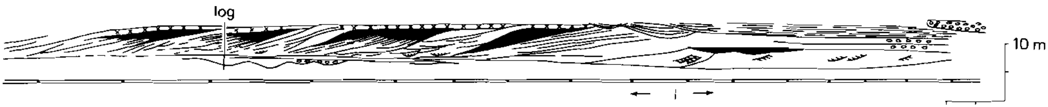


Fig. 4.
In the vertical section the point bar deposits are arranged in sets with variable mud contents (mud is designated in black). On top of the inclined beds of the upper lithofacies caliche nodules are developed in fine material. (Tracing from photographs).

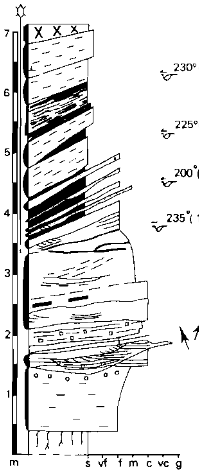


Fig. 5.
The log of the vertical section (location in Fig. 4).

Fig. 6.
Arcuate features in the top of a sandstone layer.

medium sandstone make up the middle part, while the top interval consists of a mottled interbedding of inclined mudstone and sandstone beds. Some layers with tabular sets are intercalated within the SW dipping beds at the top.

As shown in Fig. 4, point bar layers are arranged in sets (or bundles, PUIGDEFABREGAS & VAN VLIET, 1978) with varying mud content. Most boundary planes truncate preceding sets. Several sets show an extension of layers over the flat top.

The sedimentary facies found in the major part of the section is represented in the log (Fig. 5). A lower lithofacies is made up of coarse- and medium-grained sandstone with large-scale (up to 0.15 m thick) trough-shaped crossbedding in moderately inclined beds (dip up to 15°) In the lower lithofacies interval there is a blue-coloured horizon. On the NE margin this horizon has an extension towards the top of the relatively coarse-grained deposits. The upper lithofacies consists of steeply inclined (dip 11-26°) sandstone beds fining upwards to siltstone. Mud interbeds lie concordantly on top of these beds, which have a basal, erosional surface. The interval is strongly mottled.

Palaeocurrents from the lower interval are oriented slightly obliquely to the strike of the inclined beds or are directed up against the accretion planes.

### II.4.2. *Surface of a meander lobe*

Location: The exposure is situated (Figs 1 and 2, section II) in a valley on the S side of the Puente de Montañana – Tremp road (km pole 3).

Setting: Arcuate structures stand out on the 150 × 150 $m^2$ surface of a 1.5 to 2.5 m thick layer (Figs 6 and 8). The surrounding palaeochannel defines the extension of the layer: a meander lobe. The lobe is an isolated body at the top of two coarse layers, the lower being continuous and the upper one pinching out in a SW direction (Fig. 7). The lower layer shows

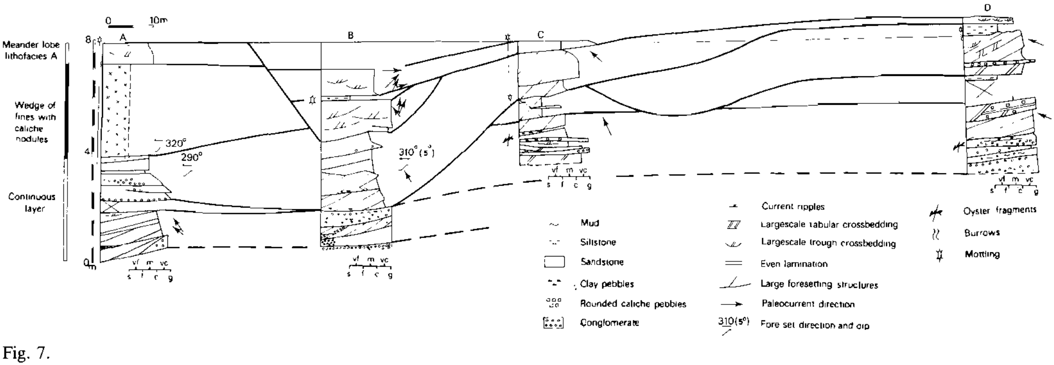


Fig. 7.
Logs located at the southern margin of the surface exposure of a meander lobe (Fig. 8). The major part of the logs is formed by sections of a continuous and of a wedging layer. At the top of logs A and B the lower point bar facies is exposed. At the base of this unit there is in log A a wedge of fine, mottled material with caliche concretions whereas in log B there is a lenticular channel fill with large trough-shaped sets.

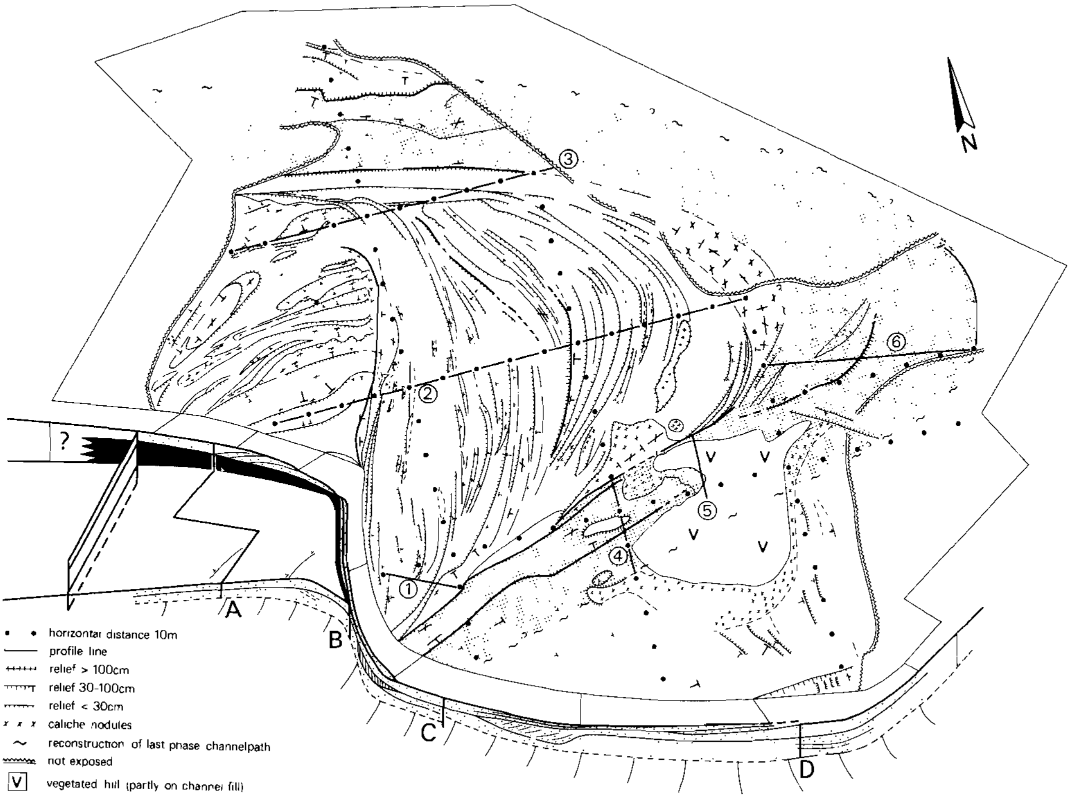


Fig. 8.
The geometry of the meander lobe and associated palaeochannel (section III) and the location of profile lines. (S part of the exposure).

large foreset structures. Some oyster fossils are present. The layers are embedded in a thick mudstone interval (Fig. 2, section II). Most of the fine sediment cover of the lobe has been removed by recent erosion. This interval of mottled mud is topped by a 0.5 m thick sandstone layer with oyster fossils.

Just to the W of the platform there is a vertical SE-NW fault. The SW layers have been displaced a few metres. Close to the fault several layers show thickening.

Description of the meander lobe (Fig. 8): On the basis of the dip directions of the accretion planes the lobe can be divided into two parts. A W unit with dip directions to the N and an E unit with dip directions to the E.

The completely mottled W unit consists of a medium-grained sandstone layer on top of which longitudinal ridges occur. The ridges are made up of inclined, medium- to fine-grained sandstone beds. Some beds drape over ridges. Underneath the flat base of the E unit a channel fill occurs with trough-shaped crossbedding (set thickness 0.5 m). The top of the fill consists of mottled mud (log B, fig. 7).

In the E unit two lithofacies units can be recognized, that correspond with the lower and upper lithofacies described earlier from the vertical section. The lower unit (coarse- to medium-grained sandstone with trough crossbedding and low angle accretion planes) thins towards the E. The mottled upper unit (steeply inclined medium- and (very) fine-grained sandstone and siltstone beds) thickens to the E. Mud interbeds only appear on the E margin. The upper lithofacies comprises convex-upward ridges which occupy only a part of the point bar. Palaeocurrent directions in the S margin of the exposure are to the NE along strike and up against lateral accretion planes.

The largely exposed palaeochannel is associated with the E unit. A distinct elevation of the base and the top of the E unit follows from the fact that to the N the base of the palaeochannel cuts the top of the W unit. Within the E unit several minor breaks in the accretion pattern are associated with elevations of the top of the lower and upper lithofacies. Nonmottled deposits of the lower lithofacies are found in

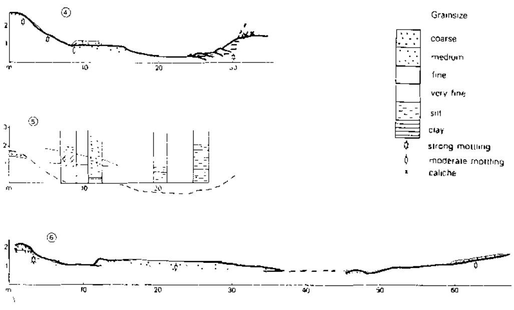


Fig. 9.
Profile lines 4 and 6 (Fig. 8) represent palaeochannel profiles with steep upper point bar slope (s), lower point bar platform (p) and cutbank (c). In profile 5 the caliche layer on top of the point bar wedges into the channel fill. Also at the cutbank caliche nodules are present.

several cases on top of mottled beds belonging to the upper lithofacies.

The width and depth of the palaeochannel vary strongly; 30 × 2 m at the site of profile 4, and 65 × 1 m at profile 6 (see Figs 8 and 9).

Three zones can be distinguished (Fig. 9 in the palaeochannel profiles:

- zone s, a steep slope with a strongly mottled deposit of very fine sandstone grading upwards into siltstone (corresponds with the upper lithofacies interval). The colour is brown with blue traces.
- zone p, a rather flat slope over a much larger distance (profile 6) or a platform with a steep downward slope at the edge (profile 4). Coarse- and medium-grained sandstone contains some oncolites at the top. The colour is blue-grey with faint brown mottles. (The interval corresponds to the lower lithofacies).
- zone c, a concave, erosive surface. The upper part is locally covered with laminated, fine sediment (profile 6). Coarse sediment fills scours in the deeper part of the channel.

Zones p and s are also found in the N part of the exposure. Zone p partly corresponds to the steep-edged platform of profile 4. The surface of the upstream part dips gently. The sediment is mainly coarse with some rounded caliche pebbles.

Nodular caliche is found on either side of the palaeochannel. The caliche layer on top of the point bar extends into the channel fill (profile 5). At the base the fine channel fill contains lenses of coarse sediment. The fill is browncoloured with a strong, (grey-)blue mottling.

## II.5.RECONSTRUCTION OF THE PALAEOCHANNEL FACIES AND DISCHARGE CONDITIONS

The palaeochannel is reconstructed in Fig. 12. The division of the point bar into two zones with greatly differing dip resembles the point bar morphology as described by BLUCK (1971) and BLUCK & FERGUSON (1981) for the meandering River Endrick. A large variation in channel width and depth (widening at the centre of a bend) is known from the Nueces River as described by GUSTAVSON (1978). In both cases major differences in grain-size distribution and sedimentary structures appear.

The alternation of sandstone, siltstone and mud beds adjacent to the palaeochannel and in the vertical section (Fig. 5) points to the existence of a discontinuous discharge. The following sequence of discharge events can be reconstructed, analogous to LEOPOLD ET AL. (1964):

- the channel profile was cut during a period of rising discharge. Inclined erosion planes were formed on the lower and upper point bar.
- during high discharge trough-shaped crossbedding developed on the lower point bar. At the upstream part of the platform, where the top is flat, high angle foresets accreted at the pool-facing margin. The head area was enriched by (very) coarse sediment and caliche pebbles. Fine sand grading upwards to silt was deposited on a part of the steeply inclined upper point bar.

Fig. 10.
The horizontal top and steep edge of the lower point bar platform (S part of the exposure).

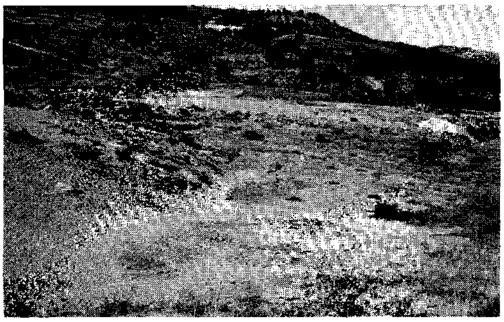


Fig. 11.
The steep upper point bar slope and the flat top of the lower point bar platform (E part of the exposure).

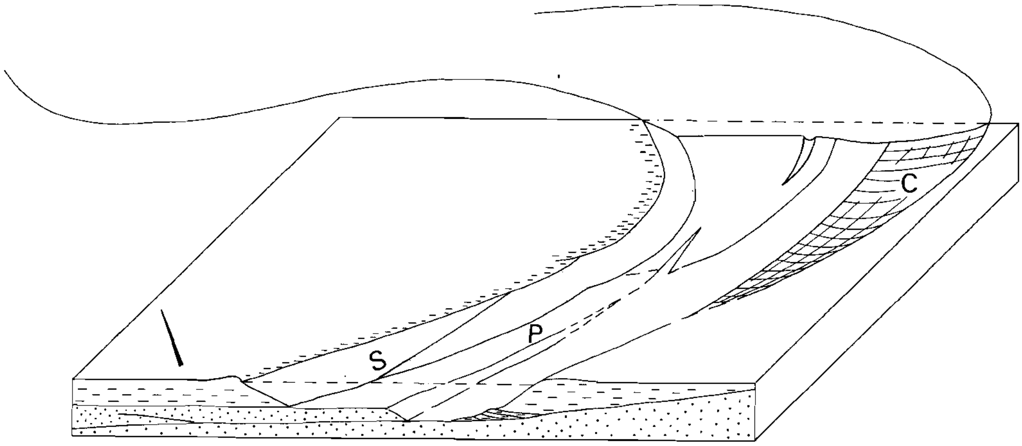


Fig. 12.
An impression of the E part of the palaeochannel without fine-grained fill. In the channel with variable width and depth the steep upper point bar slope (s), the lower point bar platform (p) and the cutbank (c) are indicated.

- during falling discharge mud drapes were deposited. Mud layers were preserved particularly on the upper point bar surface.

Sedimentary structures which correspond to a period in between peak discharge events have not been found. During this stage the upper interval was oxidized. The lower lithofacies interval remained saturated with water that contained amounts of organic matter and thus prevented the establishment of oxidizing conditions (MOODY STUART, 1966). With stagnant ground water, blue colouring could develop in the lower lithofacies interval.

The outer channel bank was predominantly an area of erosion, its outward migration involved the succession of point bar layers at the inner bank (forming ECS). Several types of stratification planes can be distinguished:

- 1 erosion planes (all over the point bar).
- 2 planes established by lithology changes (on the upper point bar).
- 3 boundary planes of sets (all over the point bar).

Type 3 planes are more distinct than planes of type 1 and 2. Type 2 planes are often difficult to trace in the lower lithofacies interval. PUIGDEFABREGAS & VAN VLIET (1978) describe a massive character of this interval in thick lithosomes with ECS. However, type 1 planes are also present in those cases.

Channel widths can be estimated easily from the vertical section since the horizontal projection of the ECS plane equals approximately 2/3rd of the total channel width (ALLEN, 1965). This rule applies well for the palaeochannel profiles 4 and 6, Fig. 9. Widths estimated from the well exposed part of the vertical section (Fig. 4) are at most 60 m. The amount of discharge can be calculated if channel dimensions and meander wave length of the rivers are known (LEEDER, 1973). The widening of the palaeochannel at the point bar centre, however, and the variable width/depth ratio are not accounted for in equations (see LEEDER, 1973). Only W×D is consistent. In addition to the difficulty of obtaining representative mean W and D values another problem is the facies variation in exposures. It is likely that during the formation of meander lobes mean W and D values varied considerably.

## II.6. DEVELOPMENT OF THE COARSE AND THE FINE MEMBER

The point bar deposits were preserved after a major lateral and a minor vertical channel migration. It is possible to reconstruct the migrating channel at the last stage of activity (Fig. 12). The combination of all the point bar features described leads to a point bar plan (Fig. 13B). An attempt has been made to reconstruct the path of currents, after comparing data with models from recent meandering rivers (BLUCK, 1971; BLUCK & FERGUSON, 1981; JACKSON, 1976).

Currents were deflected at the point bar head. A high angle slope at the upstream part of the platform was maintained by a combination of currents through the pool and over the platform into the pool. Fanning at the mid point bar was followed by a concentration of currents at the tail, with the flow directed up and against a moderately inclined surface. A helicoidal current pattern would probably not completely develop in the shallow channel. With the widening of the channel at the point bar centre important flow separation (analogous to LEEDER & BRIDGES, 1975) could take place at the upper point bar surface. The concentration of stream lines at the cutbank was highest, and thus erosion strongest, somewhat downstream of the centre of the bend.

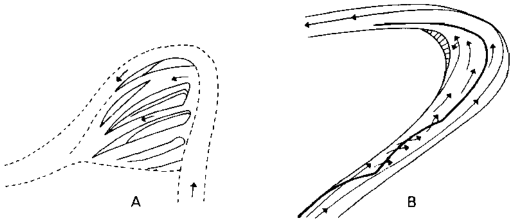


Fig. 13.
A. Reconstruction of the channel plan of the W unit. At high discharge flooding of the lateral bar occurred. B. Channel plan reconstructed on the basis of the palaeochannel described. Some current lines have been drawn in.

Sedimentary structures were formed during periods of rising and high discharge. Foresets were deposited in the accretion direction. Further palaeocurrent directions from the lower point bar interval were mainly along, or more or less up against, the point bar slope. Occasionally sets are rather tabular. Layers of this type may extend into the upper lithofacies interval, but never reach the top of the point bar. Therefore there is no scroll and swale relief (see Fig. 4). The upward directed sets originated at right angles from elongated scours in the lower point bar. At the apex of the point bar scours are directed slightly away from the upper point bar surface, at the point bar centre parallel to the face, and in the tail area towards the surface (see Fig. 15). Across current the upward directed sets will also show this trend; therefore it is possible to establish the location of a vertical section in terms of point bar geometry.

The erosional path of the channel is reconstructed in Fig. 14. Orthogonals to the accretion planes indicate the erosional path lines and the erosional axis, which is the longest erosional path line (HICKIN, 1974). The accretion direction of the W unit was probably both perpendicular to and in the current direction. Accretion of the E unit started in a straight reach with an irregular point bar. In the tail area there was a minor adjustment to the former accretion direction. With increasing lateral migration there was also a minor downstream migration.

With increasing curvature, extension of the point bar appeared in combination with symmetry changes. It is likely, that with increasing sinuosity the degree of flow separation also increased. A current pattern initiating a platform with flat top and high angle margin also only occurred during a high sinuosity phase. There is an analogy between this platform and platform development in high sinuosity, tidal channels, as described by BRIDGES & LEEDER (1976). However, the decreasing channel depth and changing point bar morphology towards the centre of the bend (Fig. 12) indicate that in this case the influence of low discharge flow can be ruled out.

The lithofacies found in the exposures shows far more variation than is accounted for in the analysis of point bar evolution just given. The W unit deviates most. The surface morphology in combination with the channel fill at the base of the E unit (log B, Fig. 7) resembles the description of a lateral bar from an outwash plain (BLUCK, 1974). The isolated setting and the lack of gravel in the W unit distinguish it from Bluck's description. A possible channel plan has been drawn on the basis of the resemblance (Fig. 13A).

In Fig. 15 transverse sections of the upstream (1), middle (2) and downstream (3) part of the meander lobe (Fig. 7) are reconstructed. The distribution of the coarse and fine member is indicated.

(1) The smallest part of the lobe. Only the lower lithofacies interval is exposed. Unstratified mud is assumed to have been present in the upper interval. The general aggradation is expressed by the draping of platform deposits over the top of earlier platforms.

(2) The largest part of the lobe. The upper lithofacies interval is well developed. The strike of the upper point bar layers shows only a minor variation.

(3) A narrow elongated part of the lobe with strongly curved strike-lines in the upstream part. The upper point bar deposits consist of mottled mud.

In Fig. 15 the general aggradation of the E unit is indicated by a large amount of mottled mud on top of the deposits. Several phases of aggradation can be distinguished in the E unit. The beginning of each phase is expressed by a break in the growth pattern (Fig. 14) and instantaneous elevation of the top level of the point bar deposits. There are major truncations at the pattern breaks, and mud appears in places where the channel withdrew partly from the point bar surface. It is probable that the shifts in the growth pattern are reflected in the formation of sets in vertical sections. Variation in the mud content of sets can be attributed to a point bar

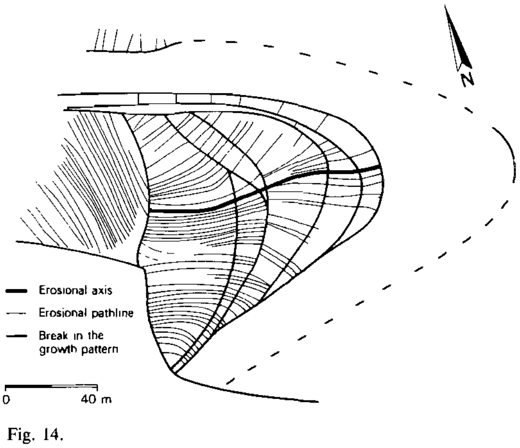


Fig. 14.
Orthogonals to the accretion pattern give the erosional pathlines and the erosional axis. The pattern of the W part is regular, while in the E part there are breaks in the growth pattern.

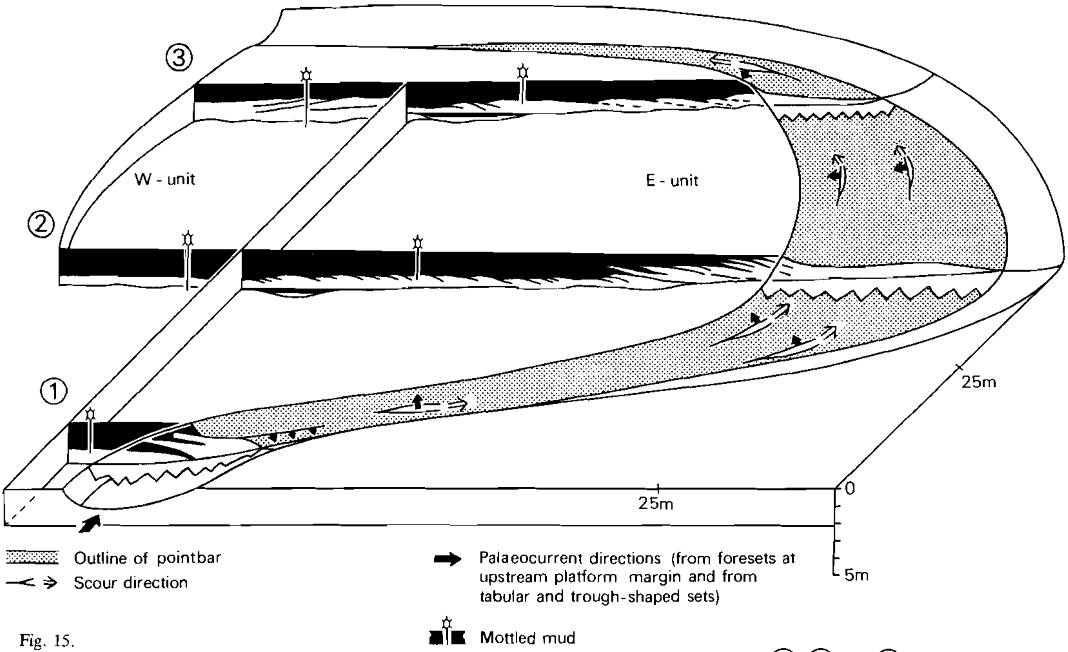


Fig. 15.

Transverse sections of the meander lobe constructed on the basis of profile lines over the surface (Fig. 8, ①, ②, and ③). Mud was deposited outside the channel during vertical accretion of the E unit. Palaeocurrent directions in two types of crossbeds give an indication of the location of the sets with respect to point bar geometry.

sedimentation process. It is likely that the site of maximum sedimentation in the flow separation zone moved over the point bar during formation of the meander lobe.

It can be concluded that the deposits originated in a river with major regime changes. After the initial, high energy phase a sudden transition took place towards a relatively low energetic river. With continuing activity there was a major lateral accretion in combination with a minor downstream migration and a distinct aggradation. River bed abandonment occurred due to avulsion.

The sedimentary facies of the vertical section described agrees largely with the section through the middle part of the meander lobe (Fig. 15; section 2). Palaeocurrent directions from the very well exposed NE part (Fig. 5) point to an origin somewhat downstream of the centre of the bend. In the SW margin a platform foreset facies has been found. Further directions derived from inclined tabular sets also point to an origin at the point bar apex area.

The dimensions obtained from both exposures are not directly comparable. The maximum width of the meander lobe in the vertical section is at least four times and the thickness twice larger than the meander lobe depicted in Fig. 15, whereas palaeochannel widths fall within the range of channel widths from the surface exposure.

## II.7. SETTING OF THE MEANDERING RIVER IN AN ALLUVIAL ENVIRONMENT

An attempt has been made to relate the origin and characteristics of the meandering river to the setting in an ancient alluvial environment. Factors which determine the development of alluvial systems are:

*Relief* (as a consequence of crustal movements and eustatic sea-level movements).

*Nature* of the material in which the system developed.

*Climate* (especially temperature and precipitation).

*Vegetation.*

JACKSON (1978) and READING (1978) present a review of the literature in this field.

In Fig. 16 the information of the general sedimentological description is compiled with emphasis on the grain size distribution. Debris from a NE uplifted area was brought into the basin by a small alluvial system. The transition of a wide, gravel-carrying stream to relatively fine-grained meandering rivers is known from Alaskan outwash fans (BOOTHROYD & NUMMEDAL, 1978). The transition is a consequence of a decrease in gradient, in this case accompanied by a deflection towards the NW plunge of the basin axis. Synsedimentary NW directed faults and flexures played an important role in this transition. In the distal part of the system extensive, fine layers were deposited by sheetfloods (analogous to HEWARD, 1978).

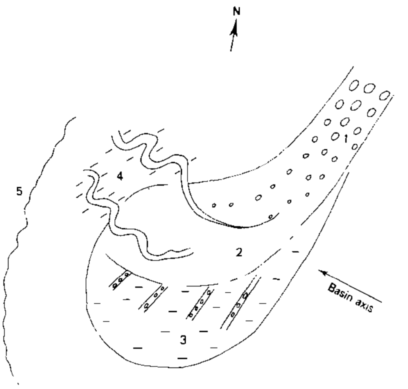


Fig. 16.
Sediments from the depositing part of an alluvial system grade downstream from conglomerate (1) to sandstone (2). Sheetfloods (3) and meandering rivers (4) deposited large amounts of fines. Marine conditions dominated in the W regions (5).

Long-term sedimentation was strongly influenced by tectonic conditions (see also NIJMAN, 1981). Deposition of thick conglomerates is linked to transgressions. Eustatic sea level rises lower gradients and would be associated with relatively finer deposits. Tectonic activity created and maintained gradients and supplied variable amounts of material. Relatively small-scale tectonic activity played an important role in the development of meandering rivers. The meander lobe was formed in a depression in front of a small deltaic build up (Fig. 8). The depression was situated in an active flexure close to the present vertical fault.

The extension of the meandering part of the system varied. The meander lobe of the surface exposure developed only after the system had deposited large amounts of (very) coarse sand and gravel in a shallow marine environment and meandering was limited to the final stage of alluvial system evolution. In the vertical section (Fig. 4) the meander lobe succeeds sheetflood deposits with minor amounts of gravel, while at the top a conglomerate layer wedges out into a scour that is filled mainly with non-mottled sediment.

The proximity of the sea might have created a tidal regime. However, sedimentary structures due to a tidal influence have not been found. The oxidation of the major part of the point bar deposits and the lack of tidal influences reflected in the deposits of the Castigaleu Formation in the study area seem to exclude a tidal influence on point bar deposition.

The type of multicolouring of the fine sediment and the caliche development agree with the characteristics of Pseudogley soil formation as given by BUURMAN (1980). This implies a seasonal, hot climate with marked dry and wet periods. A discontinuous discharge has been interpreted in the analysis of point bar deposits. The large width and the shallowness of the braided stream and the connection of coarse layers with extensive flat layers of mottled fine material also point to an alternation of high discharge periods (with sheetfloods) and low discharge periods (with little or no sedimentation). However, HASELDONKX (1973) took the occurrence of *Nypa* pollen as an indication of a humid, tropical climate without marked periods of aridity. The two different interpretations of the climate may be the consequence of the sedimentary situation. The sedimentary activity of an alluvial system in a small drainage area with only scarce vegetation can be determined by precipitation during storm conditions (MCGOWEN, 1971). Because of the thin vegetation cover in sheetflood areas (analogous to the situation described by HEWARD, 1978) the soil will have been extensively subjected to weather conditions. The coastal plain vegetation (HASELDONKX, 1973) will have been restricted to the margins of the sedimentation areas.

Variation in sedimentary characteristics in vertical sections cannot be due to changes in the climate and/or vegetation type. The palaeosol character does not change significantly in the Monllobat Formation and there is no evidence of a change in vegetation type.

## II.8.THE RELATIONSHIPS OF THE PALAEO-ENVIRONMETAL CONDITIONS AND POINT BAR SEDIMENTATION

With decreasing gradient and loss of energy in the alluvial system (expressed by the lack of gravel) a transition from unconfined to more or less confined flow took place. A high bank stability (causing a high frictional resistance to flow) was very important in this respect. The bank stability was caused by margins of synsedimentary flexures and the presence of large amounts of fine material (with some vegetation). A high suspended load/bed load ratio in the channel was also important in establishing the transition. The thick mud beds intercalated in the upper point bar deposits point to a high suspension load. Thick mud drapes can develop on point bars during stagnant periods. In the sketched situation this could take place during storm conditions, when rapid run-off through the more proximal parts of the system caused flooding of the low gradient, coastal plain. The major part of the bed load (especially the gravel) remained in the proximal parts of the system.

The initial coarse phase of the meander lobe in the top surface exposure was the first reaction of the system to a newly developed, favourable gradient through a flexure. The high discharge was through the channel and over the lateral bars. A period of active sedimentation of the channel lowered the gradient and consequently a lower energy type of river developed. Lateral accretion in this river was characterized by an alternation of phases with a regular accretion and (short) phases of aggradation, accompanied by major shifts in the accretion pattern.

Prior to avulsion, sedimentation took place in a shallow meandering river during occasional high discharges. In the centre of the highly sinuous meander bend the channel was twice as wide as at the inflection point. The width increase was associated with a distinct decrease in the depth.

## II.9. CONCLUSIONS

1 A point bar in a palaeochannel shows a well-established grading (fining upwards) in a vertical sense only. Nevertheless, there can be a large variation in sedimentary facies in transverse sections of meander lobes. Therefore, an analysis of point bar sediments has to be combined with a reconstruction of general environmental conditions.

2 Large-scale environmental conditions have put distinct marks on point bar sediments. In this respect the most striking features are the reconstructed discharge events through a channel with highly variable W/D ratio, the setting of point bar layers in isolated meander lobes with a coarse initial part, and the setting of meander lobes in associations with sheetflood deposits.

3 Reconstructions of point bar sedimentation processes and the environmental setting lead to a three dimensional model. The location of vertical sections in this model can be established by comparing sedimentary facies and palaeocurrent directions.

# chapter III

# INTERNAL STRUCTURE AND ENVIRONMENTAL RECONSTRUCTION OF EOCENE TRANSITIONAL FAN-DELTA DEPOSITS, MONLLOBAT-CASTIGALEU FORMATIONS, SOUTHERN PYRENEES, SPAIN

## III.1. INTRODUCTION

The Eocene Monllobat Formation is situated in the Southern Pyrenees, Spain (Fig. 1 A, B); it contains a large amount of mottled mudstone (70–80%). This mud is thought to have been deposited by meandering rivers as well as by sheetfloods in the distal parts of shallow, wide, graveliferous systems (Van der Meulen, 1982). Distally the systems also were in contact with a shallow marine basin. The continental Monllobat Formation interfingers with the marine Castigaleu Formation (Fig. 1C); The latter also consists mainly of mudstone, but there is no mottling. Further, only small amounts of conglomerate are found.

A stratigraphic interval of limited extent is described to illustrate the transition of conglomerates to sand- and mudstone. A part of the mud is mottled. For the rest blue-grey mud is found, occasionally with marine fossils. The chiefly three-dimen-

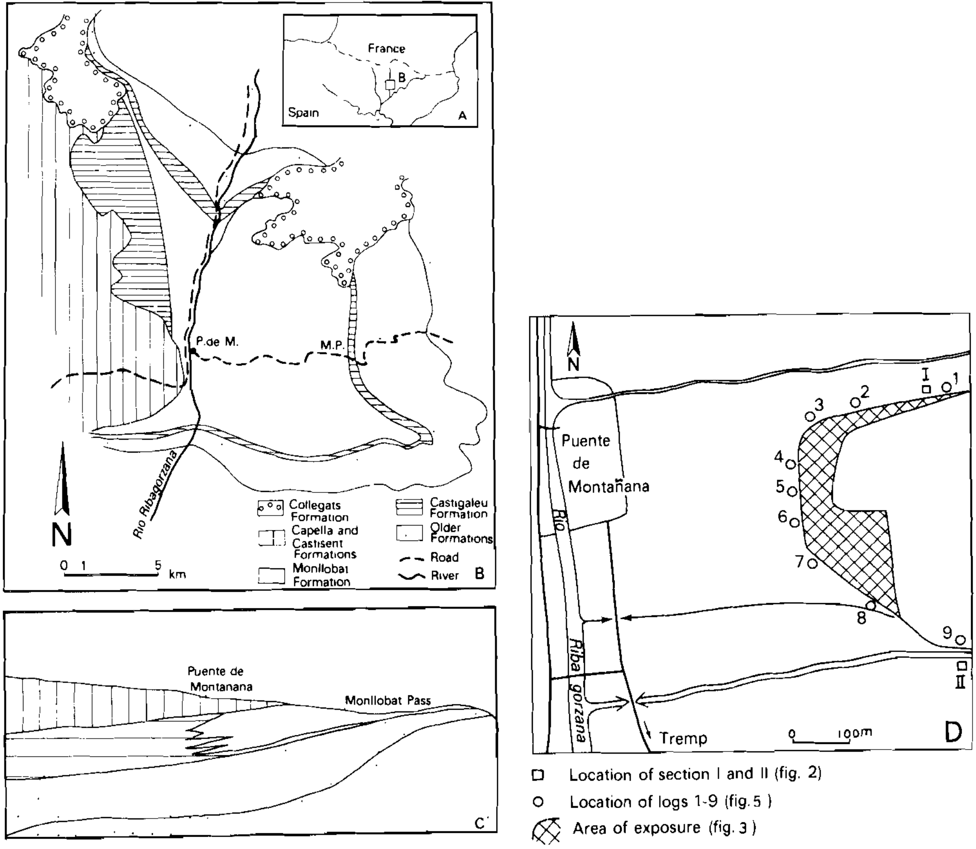


Fig. 1. Locality (A) and geological (B) map, an E–W section of the area (C) and a sketch map (D) of the investigated interval.

sionally exposed deposits, with low tectonic dip, are situated near Puente de Montañana (Fig. 1D). The morphology of the layers, the sedimentary structures in vertical sections and the bedforms on top surfaces were studied to reconstruct rivermouth processes. Also the stratigraphic setting has been considered. The setting of the Monllobat Formation in the Tremp–Graus Basin is given by Van Eden (1970), Nijman and Nio (1975), Puigdefabregas and Van Vliet (1978) and Nijman (1981).

## III.2. DESCRIPTION OF THE EXPOSURES

A three-dimensional exposure and a vertical section are described. The exposures are situated between two tributary valleys of the Ribagorzana valley (Fig. 1D). Towards the north some conglomerate and sandstone layers occur approximately on the same stratigraphic level. To the south of the interval, poorly exposed mud occurs on this level. Sections I and II (Fig. 2) also show decreasing mud contents in this direction (locations in Fig. 1D). To the south the interval is bounded by a series of vertical faults, directed from SE–NW. There is an uplift of a few metres at the southwestern side. The tectonic dip is 1–2°.

### III.2.1. *Petrography*

Conglomerates of the proximal parts consist mainly of clastic limestone pebbles, with minor amounts of quartz pebbles. Several large blocks (up to 0.5 m thick) of fine mottled material are intercalated. In the distal parts intraformational mud clasts and rounded caliche pebbles are found. Sandstones consist of clastic limestone grains with minor amounts of quartz and alkalic feldspar grains.

The major part of the mud is mottled (see Fig. 5). In addition blue-grey mud is found. At the top of the northern part of the three-dimensional exposure a sudden transition of blue-grey to mottled mud is found (Fig. 3). However, the western longitudinal rise shows a profile with a gradual change of the colour of the sandstone. Normally grey-coloured sandstone passes upwards through the following range of colours; blue–blue with brown mottles–brown (sandstone with burrows)–brown with vertical blue mottles. The type of mottling agrees with the description by Buurman (1980) of Pseudogley type soils. Non-mottled mud of the three-dimensional exposure is found in laterally non-persistent layers (Fig. 5). Large amounts of plant debris are locally present. At the base of the interval mottling gradually disappears upwards (log 3, Fig. 5). A thin lignite layer is intercalated in the blue-grey mud. In the basal parts of the mottled mud, thick caliche beds are found (logs 3 and 9, Fig. 5). The fine matrix may contain quantities of coarse-grained sand and some pebbles.

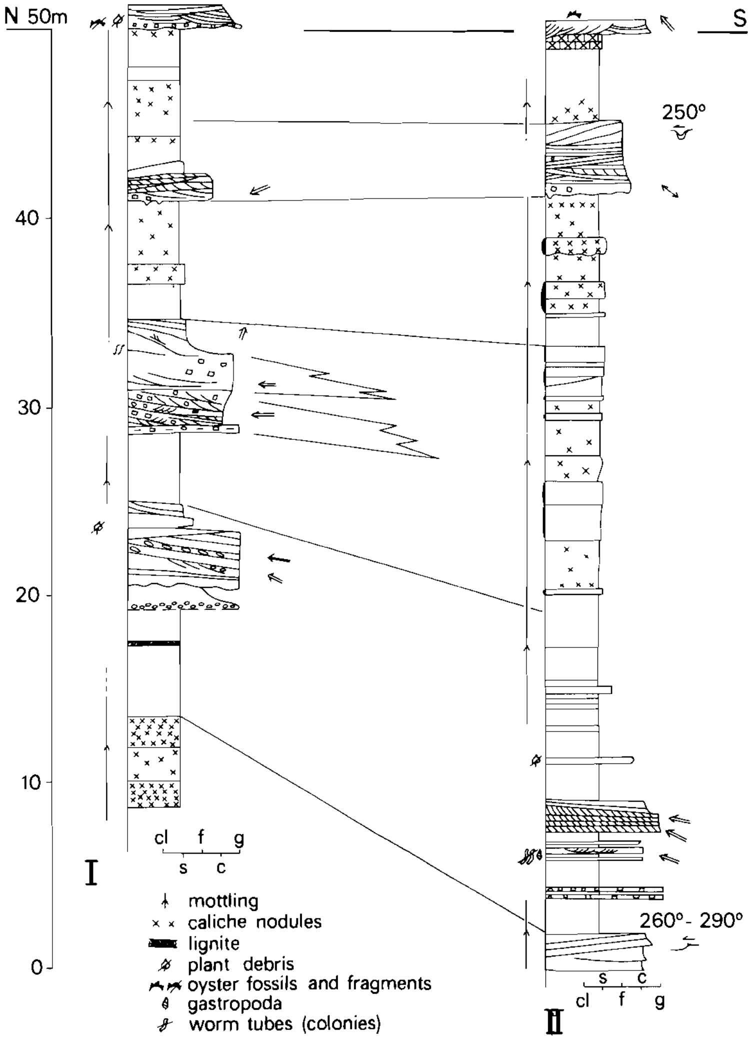


Fig. 2. Two general sections of the Monllobat Formation near Puente de Montañana (see Fig. 1D for locations). The interval which has been studied in detail constitutes the basal parts. Note the thickening of mud sequences towards the south.

### III.2.2. *Division of the exposures into lithofacies units*

Four lithofacies units have been distinguished on the basis of the geometry of the layers and the type of lithology (Figs. 3 and 4).

*Unit 1* consists of two southwards dipping layers with nearly flat lower ends (Fig. 3). The lower layer makes up the major part of the northern transverse section. The upper and lower surface are undulating (Figs. 3 and 4).

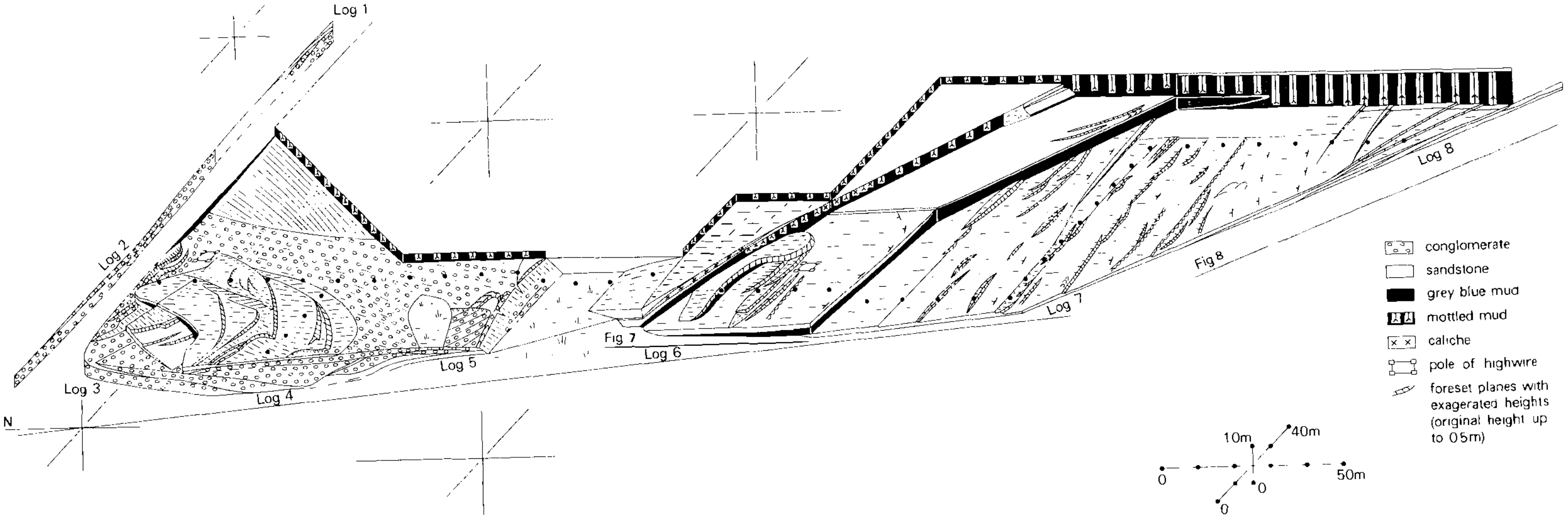


Fig. 3. The three-dimensional exposure shows a marked lateral facies transition in the major palaeocurrent direction. A conglomeratic facies passes laterally into a mud facies with some sandstone layers. The locations of logs (Fig. 5) and figures of details (Figs. 7 and 8) are also given in this figure.

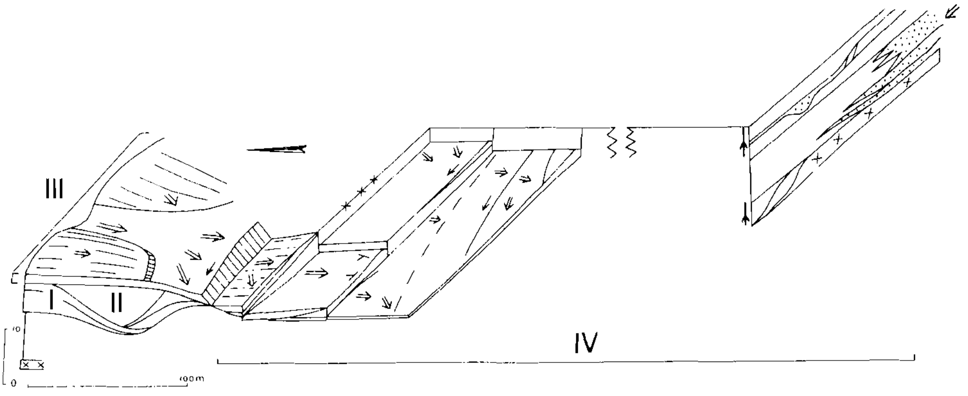


Fig. 4. A general view of the exposures in the investigated interval. To the north the exposure is three-dimensional (Fig. 3). To the south a vertical section is completely exposed (Fig. 10). The thickness has been exaggerated three times. Palaeocurrent directions are indicated on the top surfaces.

In the eastern thickening, large-scale foresetting is directed to the west (log 1, Fig. 5). The western thickening consists of massive conglomerate with some imbrication of flat pebbles (log 3, Fig. 5) (directed to the southwest). The eastern part of the lobe is made up of conglomerate layers, some of which contain large amounts of rounded caliche pebbles (log 2, Fig. 5). Directions from imbrication and tabular cross-beds are to the west. In the longitudinal section massive and banded conglomerates of the two inclined (10–15°) layers pass downwards into flat sandstone beds with upcurved ends. These beds are massive (base of log 4, Fig. 5) or tabular cross-bedded and cross-laminated (Fig. 6a) with northwestern palaeocurrent directions. Large tabular foresets (Fig. 6c) make up the slightly inclined (5–10°) top of the upper layer. The sandy conglomerates show some imbrication. The longest axis of elongated pebbles parallels the strike. Flat pebbles dip in an upstream direction when related to the foreset. However, the dip is downstream with respect to a horizontal face.

*Unit II* consists of southwards as well as northwards-dipping layers (Fig. 3). The bases of the southwards dipping layers are tangential to a northwards dipping truncation plane in the middle part of the unit. Northwards dipping layers underneath the truncation plane possess a flat upper part.

The southwards dipping layers show foreset structures in conglomerates and tabular cross-bedding in (conglomeratic) sandstone. Directions are to the southwest and the northwest, respectively. The beds underneath the truncation plane are coarsening and thickening upwards. Near to this plane parallel-laminated, inclined (8–14°) beds are topped by an erosive conglomerate bed. Distally tabular cross-bedded sandstone (southwestern direction) succeeds a mud-/sandstone interbedding (northwestern direction, log 5, Fig. 5).

*Unit III* consists of a flat layer with two north–south aligned thickenings. A steeply inclined (20°) face bounds the unit to the south (Fig. 6b).

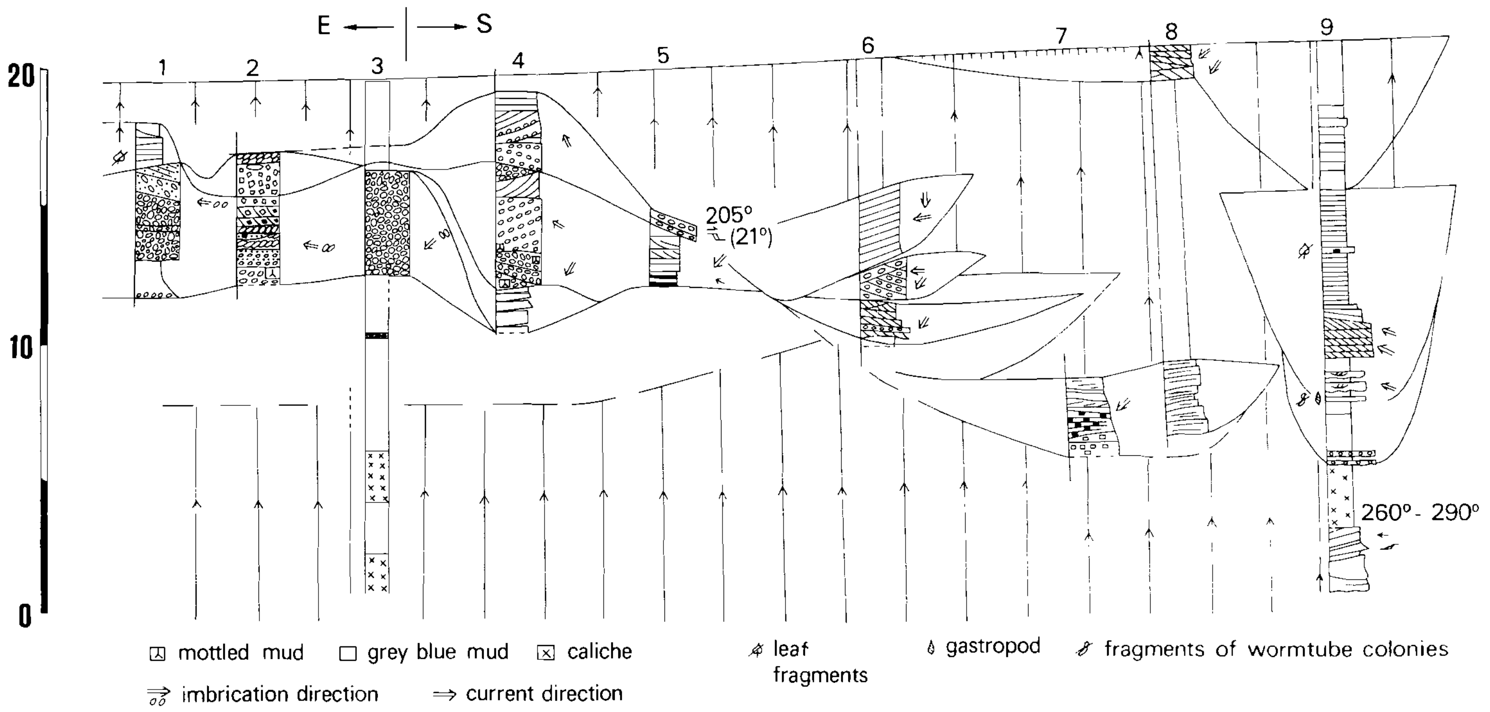


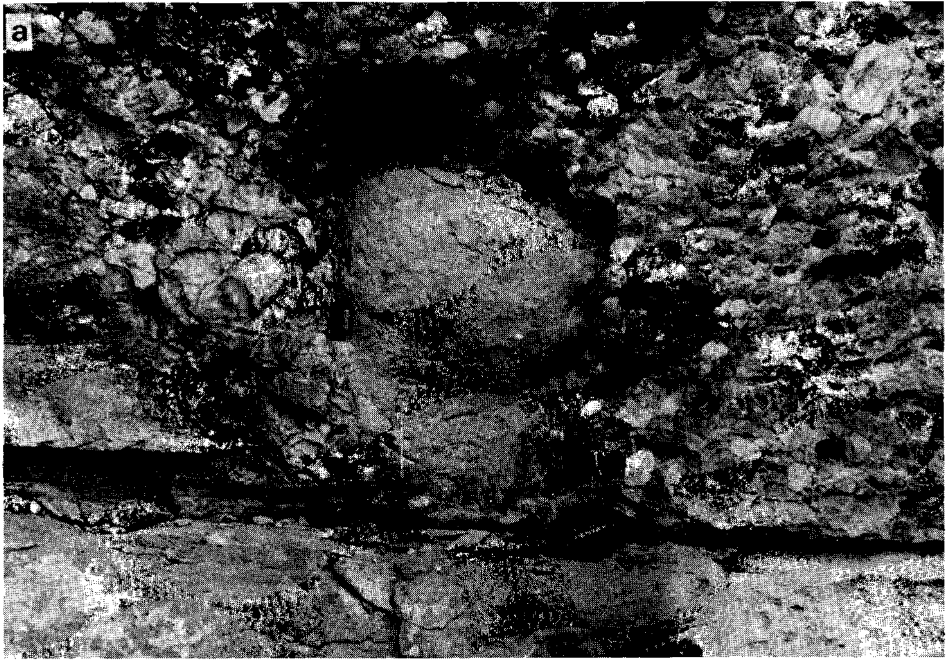
Fig. 5. The location of the logs of the investigated interval is given in Fig. 3. Most striking is the occurrence of blue-grey mud, which is extensively developed in the lower parts, but is limited to pockets in the upper parts.

Most of the western rise is mottled. Mudstone is intercalated between the arcuate, inclined sandstone beds (Fig. 3). The top of the eastern rise is also mottled. Conglomeratic sandstone with tabular cross-beds makes up the core, which is covered by a convex sandstone layer. Massive conglomerates connect the two thickenings. To the south these conglomerates are tabular cross-bedded (top of log 5, Fig. 5). The maximum 4.5 m high southern boundary face shows small slump scars at the top and a conspicuous fining upwards at the base from gravel to medium sand. Weathering of the southwestern part of the exposure has revealed stacking and overriding of tabular sets with arcuate foresets (Fig. 3). The large, inclined face is only a superficial feature.

Palaeocurrent directions of the largest features are to the south-southwest (see also Fig. 4). Close to the crest of the southern boundary face trough cross-beds in (conglomeratic) sandstone is directed to the west (see also Fig. 3). Deviating 340° and 310° directions are found on the lee of the western rise of unit I.

*Unit IV* consists of several coarse layers embedded in fine sediment. Most layers wedge to the south. A deep-mud-filled depression separates this unit from units I, II and III.

The western longitudinal section of the three-dimensional exposure shows four subunits, of which only the upper three are well exposed (Fig. 7, location in Fig. 3).



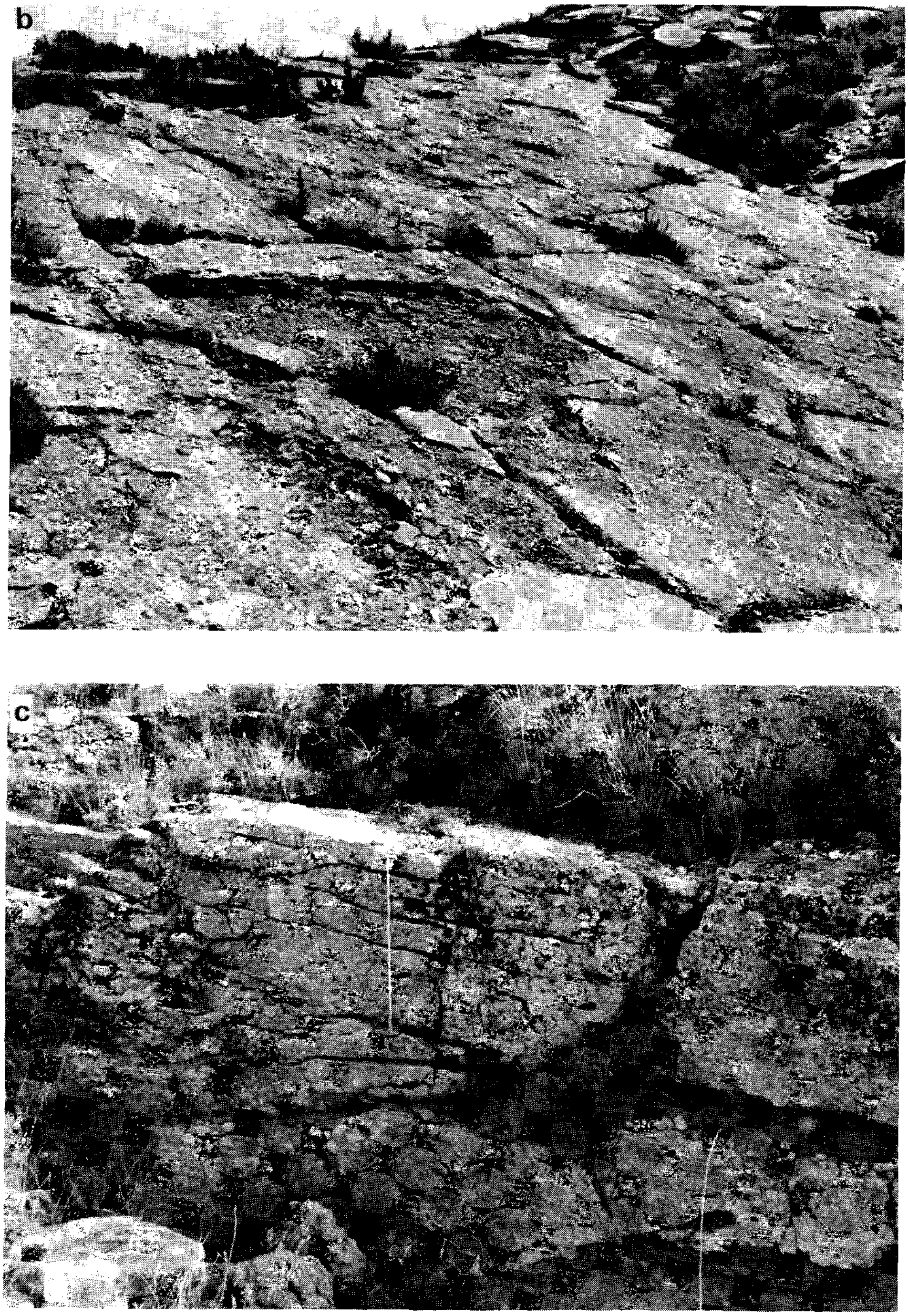


Fig. 6. a. Tabular cross-laminated sandbeds of the basal parts (bottomset) are incised by an (inclined) conglomeratic layer (foreset). A large, intraformational block (mottled, fine material) is intercalated. b. View of the maximally 4.5 m high foreset face, which bounds unit III to the south. c. Large cross-beds (scale is 0.5 m) from the rather flat top of unit I.

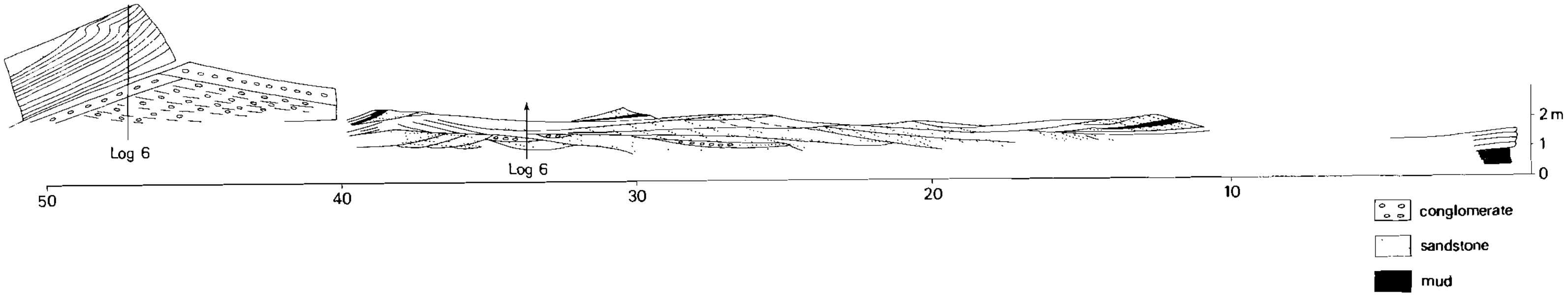


Fig. 7. Detail of Fig. 3. The lower layer is made up of climbing, flat and descending tabular sets. Intercalation of mud layers at the top and distally in combination with a concave lower surface cause the transition to a mud facies. A stoss-side bed and foreset beds make up the middle layer. The upper layer consists of parallel lamination with convolutions and a trough-shaped crossbed at the top.

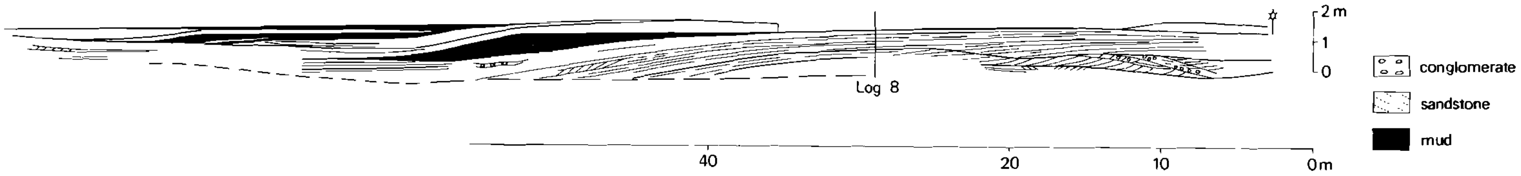


Fig. 8. The southeast part of the oblique section of the lower-most division of unit IV. Parallel lamination is the main sedimentary structure of the inclined beds. Some mud layers are intercalated. To the northwest of this location tabular sets appear in the section with a transition towards trough cross-bedding in the northwestern part of the section. Location in Fig. 3.

Massive, fine-grained beds of the lower exposed subunit succeed blue-grey mud (Fig. 7 and log 6, Fig. 5). Laterally tabular cross-bedding is arranged in climbing, horizontal and descending sets. Climbing sets at the top of the section continue as climbing bedforms on the top surface. Directions are to the south-southwest.

The next subunit contains conglomerates, which contrasts with the composition of unit IV as a whole. Large foresets and a stoss-side bed are preserved. The subunit thins to the south and terminates in a curved foreset face with southern-southwestern directions. To the west side there is a pronounced northwestern deflection.

The upper subunit dips to the north, the top is flat. Parallel lamination is the main sedimentary structure with trough cross-bedding (Fig. 7) and tabular sets, both directed to the west. Convolute lamination appears at the top (Fig. 9a). The transition to a flat top part also goes together with thinning. This subunit again terminates in a curved foreset face (see Fig. 3).

The lowermost subunit of the three-dimensional exposure is well exposed in an oblique section (location in Fig. 3). Several inclined layers can be distinguished. From the southeast to the northwest the dip of the layers decreases. In the southeastern part parallel lamination drapes a convex upwards core (Fig. 8 and the base of log 8, Fig. 5). The base and top of bent, inclined layers is flat. With decreasing dip, tabular and large trough cross-bedding is successively found.

The base of the subunit is made up of intraformational lags (log 7, Fig. 8). The flat top part can be traced over the convex-upwards top surface (Fig. 3). A series of climbing, rather straight-crested bedforms (0.1–0.3 m thick) and inclined, parallel laminated beds (0.1 m thick) form reliefs up to 0.5 m high. Erosive trough-shaped sets (0.02–0.1 m thick) are directed at right angles to the straight-crested bedforms.

Two trends accompany the westwards lowering of the top surface.

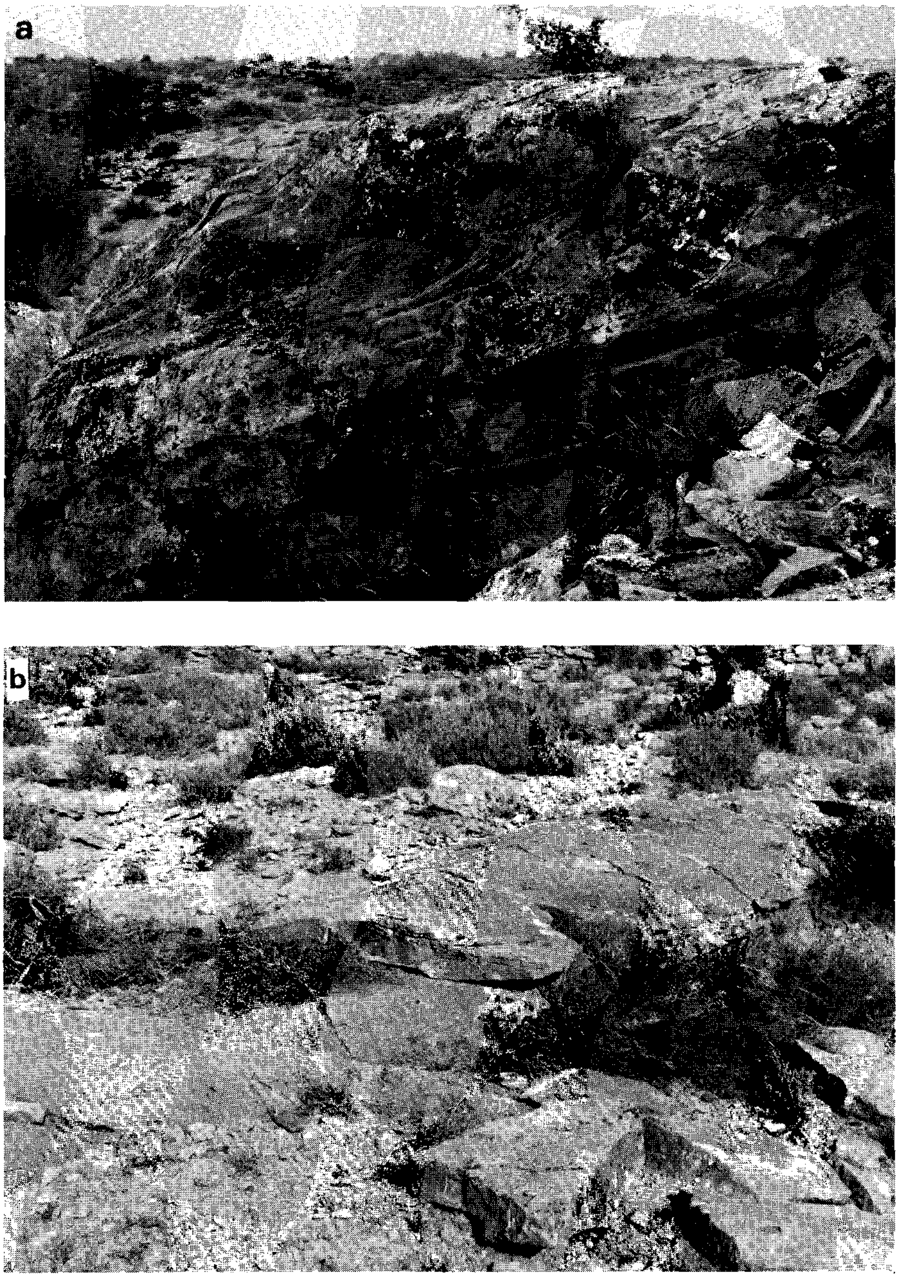
(1) In the eastern parts layers are erosive to each other; to the west well-developed coarsening-upwards sequences are found and mud is intercalated between the layers. The eastern deposits are relatively coarser.

(2) Crescent-shaped bedforms are linguoid to the east and barchanoid to the west (Fig. 3). Palaeocurrent directions are to the south, but at the western margin strong northwestern deflection is found. Linguoid forms possess low-angled, lobate foresets (dip 8–10°). The stoss side dip 10–14°. The foreset dips of barchanoid forms have a very wide range (13–30°). Stoss-side dips vary from 8–15°. The two bedform types are represented in Fig. 9b,c.

The succeeding subunit at the western side also shows the second trend on a convex-upwards top surface. The basal surface is concave upwards in the southern as well as in the eastern direction and wedges rapidly. To the east a flat layer has the same southwards extension.

The upper two subunits at the western margin dip to the north. However, the rather flat upper ends are convex upwards in a W–E direction (Fig. 3). In this direction the upper subunit wedges against the lower. Further to the east the lower one passes over into a thin caliche layer at the centre of the exposure. On the same level 0.2 and 0.5 m high foreset faces are found on the eastern side (Fig. 3). The foresets are made up of mottled silt.

A thin subunit at the southeastern top of the three-dimensional exposure can be traced into the tributary valley to the south of Puente de Montañana (Figs. 3 and



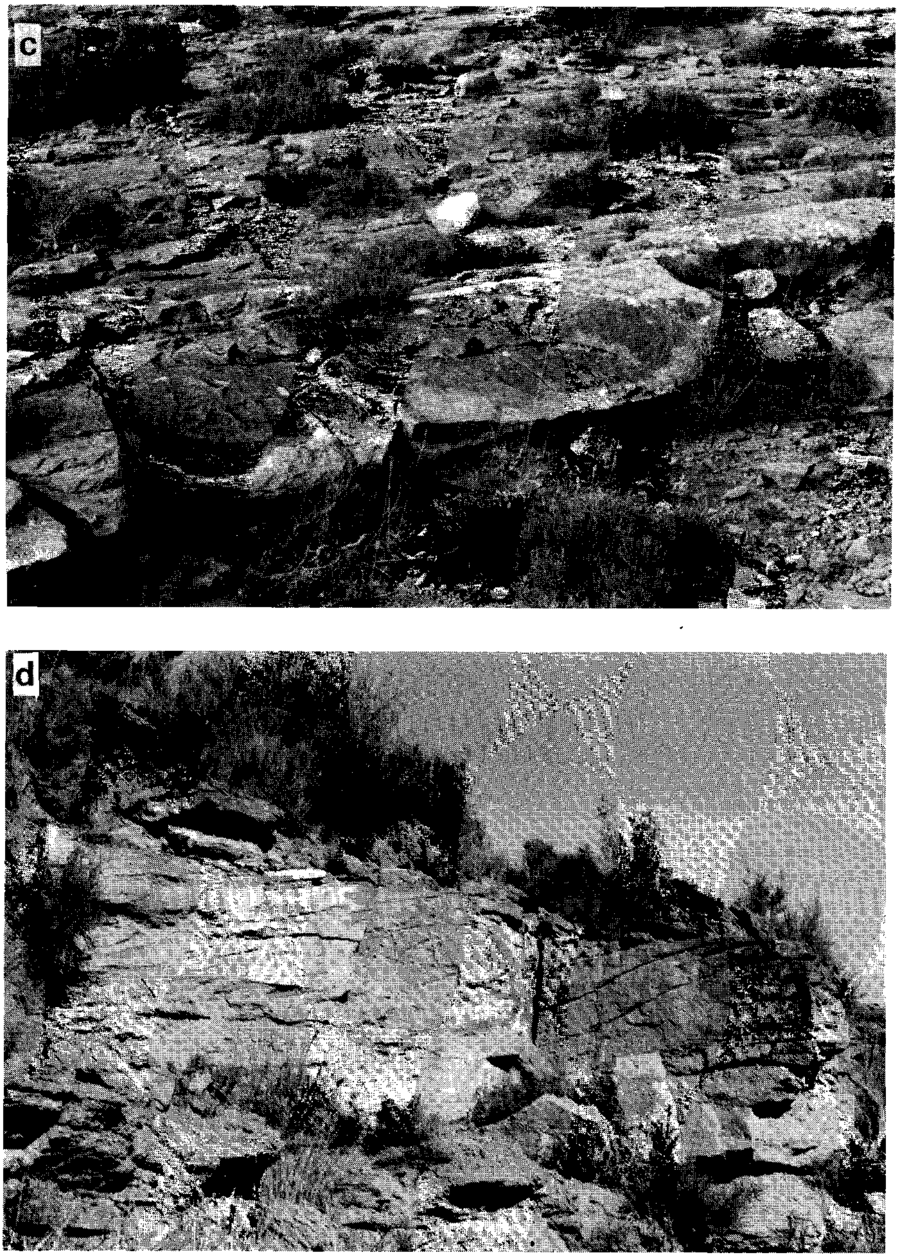


Fig. 9. a. Convoluted parallel lamination from the northwestern part of unit IV (upper layer in Fig. 7). b. The rather low-angled foresetting constitutes a lobate set with slightly erosive base (part of a lingoid bedform, southeastern top of the lowermost division, Fig. 3). c. High-angled foresetting of a barchanoid bedform (southwestern top surface of the lowermost division of unit IV, Fig. 3). d. A transition of trough-shaped to tabular sets of cross-bedding in the lowermost subunit of the southern vertical section (Fig. 10).

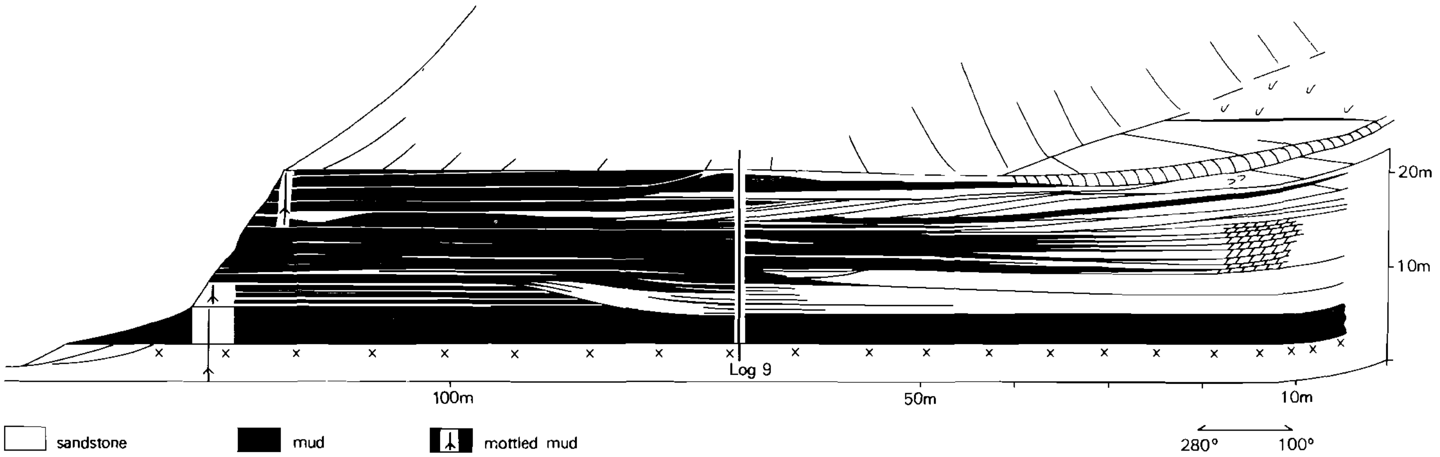


Fig. 10. The vertical section of the southern, distal parts of the investigated interval (see Fig. 4 for position with respect to the three-dimensional exposure). The lower layer with caliche contains a large amount of sand and fine gravel in a matrix of fines. The non-mottled interval contains three sandstone layers, which wedge in the palaeocurrent direction (to the northwest). Wedging is due to concave lower and convex upper surfaces and/or thickening of intercalated mud layers. The upper mottled interval is the downstream continuation of the upper sandstone layer in the southeastern part of Fig. 3.

1D). The subunit thickens in this direction. Tabular cross-bedded sandstone (log 8, Fig. 5) passes into a completely mottled silt/mudstone interbedded with minor sandstone intercalations (log 9, Fig. 5). In the tributary valley, flat sandstone layers also are embedded in mudstone. The down-current extension of three non-mottled subunits (Fig. 10) decreases successively upwards, as was the case in the three-dimensional exposure. However, the palaeocurrent directions are to the northwest, instead of to the south. Tabular cross-bedding is the main sedimentary structure. The lower subunit shows a lateral transition of large trough-shaped sets to tabular cross-bedding (Fig. 9d). At the base there is an intraformational lag interbedded with parallel-laminated, fine sediment. This material shows an alternation of very fine sand and mud laminae with leaf imprints. Fossils (amongst them oysters) are also found in the base of the non-mottled interval on the opposite valley wall.

The described section (Fig. 10) is parallel to the northwestern palaeocurrent direction. Therefore the details of a transition to a complete mudstone sequence can easily be distinguished. A combination of convex-up or flat top surfaces and concave-up basal surfaces causes the wedging of sandstone layers in a grey-blue mudstone sequence. This transition can also be the direct result of increasing thickness of intercalated mud layers together with thinning of cross-bedded sets.

In the mottled top part, the thickening of intercalated mud layers also plays a role in the southwards thickening of the continuous subunit at the top of the interval. Intercalated sandstone layers terminate in a foreset plane (Fig. 10). Only a coarse-grained sandstone lens with flat siltstone wings can be traced further to the south. Siltstone layers contain nodular caliche.

### III.2.3. *Sequences*

The described characteristics lead to coarsening-upwards sequences in the mainly conglomeratic, proximal parts (unit I, II and III). The decreasing extent of sandstone layers up to the top of unit IV formed a majority of the fining-upwards sequences. Proximal as well as distal sequences are topped by a thin fining-upwards layer. The distal top lies somewhat higher than the proximal top.

The presence of large-scale erosion planes in unit I and II involves occasional truncation of the basal sandstone parts. Inclined layers of unit I and II are mostly bounded by erosion planes. However, part of the lower layer of unit I shows a flat top draped by mud. The mud layer is overlain by mud pebbles and conglomerates on an erosion plane. Intercalation of mud layers is also found in the lowermost division of unit IV in the three-dimensional exposure.

## III.3. ENVIRONMENTAL RECONSTRUCTION

Several features indicate that the deposits were formed in a fan-delta environment.

(1) The marine fossils in the distal base of the interval point to the presence of marine standing water. Large quantities of fine material are mottled under subaerial conditions.

(2) The morphology of the layers and the direction of the largest sedimentary structures originated during a progradation process towards the marine environment.

(3) The environmental setting on a large-scale involves sediment supply by a small alluvial system transporting immature clastics from the rising Pyrenees (reconstruction in Van der Meulen, 1982). Lithofacies unit III represents the deposits and abandonment relief of a shallow, braided stream.

Wescott and Ethridge (1980) distinguish three subenvironments of the fan-delta; the fan-delta plain, transition zone and submarine fan-delta. The described deposits are the results of sedimentation processes in the transition zone of a fan-delta.

### III.3.1. *Evolution of the sedimentary relief*

Three major phases can be distinguished, corresponding to the sedimentary relief (Fig. 11).

*Phase 1*: Sediments were deposited on a southwards dipping fan-delta face, which was undular in a transverse direction. The coarsest sediments are found where deposits are thickest. Gravel grades to sand downwards over the face. At the base there was a west-northwest deflection of currents. Mud was deposited at some distance from the fan-delta face.

*Phase 2*: The rise in front of the face was draped by sand beds deposited from southwest directed currents. These deposits have been partly eroded during continuing progradation.

*Phase 3*: Deposition was mainly to the south of the previously formed sediments of unit I and II. The relatively thin unit III on top of units I and II was formed simultaneously. This is indicated by the similar palaeocurrent directions from the southern boundary face of unit III and the transverse bar buildups of unit IV adjacent to the face. The unit IV lithofacies is largely different from unit II facies at the same level underneath the inclined face (compare Fig. 7 and log 5 of Fig. 5).

The development of unit IV can be divided into three subphases, however, the sedimentary processes leading to unit III cannot be unraveled. Deposits of unit III consist of convex-up, longitudinal bars, two of which are now exposed. The western rise of unit I formed the nucleus for the accretion of sand and mud beds. In the course of phase 3, the stream bed aggraded and longitudinal bars became more closely spaced.

In front of unit III, sediments were deposited by sheet flow over the southern area of the three-dimensional exposure during phase 3a. During phase 3b two sandstone layers were formed, separated by a small mud area. With the progradation of the central gravel bar in the upstream parts, the mud zone developed into a mud rise on

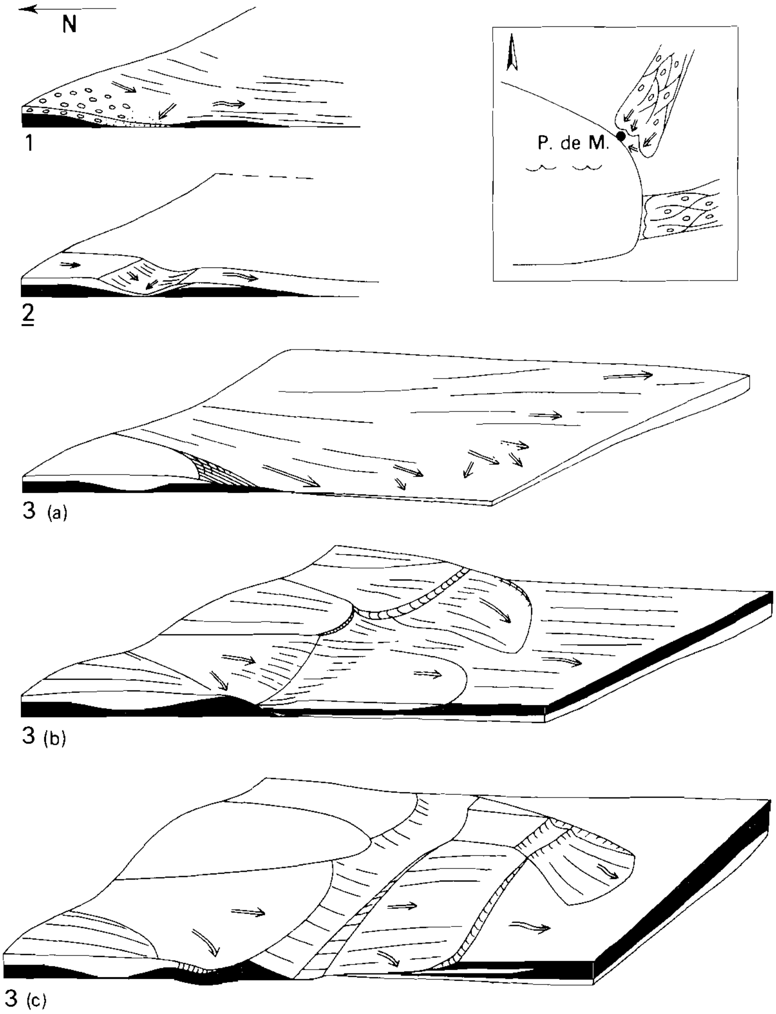


Fig. 11. A reconstruction of the evolution of the sedimentary relief during three successive sedimentation phases.

the leeside of the longitudinal bar. During phase 3c a channel was scoured on the eastern side and erosively based, coarse layers draped the western side.

Concurrently strongly deflected currents deposited sand layers further to the south. This situation is sketched in the inset of Fig. 11.

Finally a flat sandstone layer, that grades to mud in the palaeocurrent direction, was deposited by sheet flow over most of the transitional fan-delta deposits.

### III.3.2. *Reconstruction of discharge conditions*

Sedimentation of the gravel layers of units I and II was preceded by erosion. Occasionally a noneroded, flat surface, covered by a mud layer was preserved. Mud pebbles in the succeeding conglomerates indicate erosion after a period of time sufficient for mud compaction.

McGowen (1970) demonstrated that alluvial sediment supply to a recent fan-delta was restricted to storm periods with large amounts of precipitation. It is likely that the layers of unit I and II, bounded by large erosion planes, were also formed during rather short sedimentation events. Flow energy must have been very high. During waning flow conditions, mud was deposited.

The intercalation of mud layers in the lowermost division of unit IV points to a development due to a succession of several sedimentation events. The upper divisions of this unit do not show intercalated mud layers and will have been formed during one event. The limited extent of the subunits agrees with such a conclusion. As stated before, unit III does not show an alternation of facies successions on a large scale. However, a small-scale alternation of sand and mud beds is found in the western rise of the unit.

## III.4. INTERPRETATION OF HYDRODYNAMIC CONDITIONS AND SEDIMENT MOVEMENT

### III.4.1. *Interpretation of rivermouth processes*

Sandstone layers in front of the fan-delta face are of limited lateral extent (Fig. 11). The sedimentary relief of the shallow, braided river evidently was high enough to cause jet flow. In the proximal parts, the coarsest sediments are found where deposits are thickest, due to concave-up basal surfaces and convex-up top surfaces. This lateral variation points also to a certain degree of confinement of the main outflow.

Sediments reworked by marine agents as described by Wescott and Etheridge (1980) from the transition zones of fan-deltas have not been found. The major part of the distal sequence of the three-dimensional exposure consists of mottled mud. The pseudo-gley type of soil formation developed under mainly subaerial conditions. The geometry and sedimentary structures are considered to be solely the result of outflow processes at the mouth of an alluvial stream.

Farquharson (1982) described flat mouth bar-type deltas and a Gilbert-type delta (well-developed foresets), intercalated in a Mesozoic alluvial sequence. The deltas were formed in shallow lakes, which involves rather thin units (up to 14 m thick). The thickness of deposits described in this paper also falls within this range. Further there was also a lack of reworking processes.

Farquharson (1982) made the interpretation that the two delta types originated during hyperpycnal and homopycnal inflow, respectively.

*Hyperpycnal inflow* (more dense than receiving basin waters Bates, 1953); formed a plane jet underneath the basin water. Bottom friction caused deceleration and consequent mouth bar sedimentation (Wright, 1977).

*Homopycnal inflow* (as dense as receiving basin) formed a turbulent, axial jet. Flow deceleration and consequent sedimentation was due to inertial forces.

*Hypopycnal inflow* (less dense than receiving basin) did not play a role in the freshwater basin; this type of inflow forms a plane jet on top of (salt) basin waters. Expansion decelerates the flow from which suspended sediment settles. In our case, the large amounts of gravel in the proximal parts and the geometry and structures of the distal parts rule out this possibility.

The delta type which was described changes from a Gilbert- to a mouth bar-type in the progradation direction. Farquharson (1982) suggests that the single Gilbert-type delta might be formed during a period with decreased suspension load. A similar variation of the sediment type did not occur during fan-delta progradation, as is pointed out by the large mud contents of lithofacies unit IV. The decreasing slope does accompany a disappearance of gravel from the sequence.

It can be concluded that, although fan-delta deposits originated during jetflow, inflow models from the deltaic environment cannot be directly applied.

### III.4.2. *Analysis of the three-dimensionally exposed units III and IV*

A starting point for the reconstruction of hydrodynamics and sediment transport is found in the three-dimensionally exposed units III and IV (phase 3, Fig. 11).

(1) Unit III records a period of progradation of a braided stream in the area. Aggradation of the stream bed accompanied progradation. Gravel was unable to pass over the top of the western rise. Curved, low-angled sand and mud beds accreted on the lee. Transverse bars with flat tops developed during continued enlargement. The eastern rise was mainly built by transverse gravel bars with high-angled slip faces. However, most of the gravel passed through the central depression. The continuous sand covers point to sediment settling all over the rises, prior to abandonment.

Superficially erosive features indicate flow over the top of the rises into the depression. Channelized flow in the depression also caused some erosion of the rises and directed a small gravel bar up against the western rise (see also Fig. 3). The maximum depth of the channel relative to the top of the rises is 2 m.

The western orifice for jetflow was formed by the downstream terminations of the rises (Fig. 11, phase 3b, c). Gravel sedimentation and consequent shallowing accompanied flow expansion. With increasing flow expansion, transverse gravel bars with linguoid shapes were formed. Stacking and overriding of successive gravel bars during the whole of phase 3 has caused the transition of a low-angled to a high-angled fan-delta (foreset) face.

(2) Unit IV: the most extensive division of this unit appears at the base (phase

3a). Marine fossils further distally point to marine conditions nearby. The top of this subunit and the mud sequence on top are mottled. This sequence involves a transition from inflow in a standing body of water to inflow strongly obstructed by a mottled mud rise. Therefore it is concluded that frictional forces must have played a role of increasing importance during phase 3.

The erosive base and intraformational lags of the lowermost subunit are indicative of strong friction, even during phase 3a. During phase 3b coarse deposits, enveloped by grey-blue mud, filled scour hollows. During phase 3c deposits with erosive bases were draped against and over the mottled mud rise in front of the fan-delta face.

The geometry of the lowermost subunit developed in standing water. However, the lobate shape pointing upstream differs from the lunate shape of a mouth bar, formed due to frictional resistance to hyperpycnal inflow (Bates, 1953; Wright and Coleman, 1974). Lunate mouth bars consist of a transverse bar linked with subaqueous levees. The sedimentary structures are also much larger than those described by Farquharson (1982) from mouth bars.

The geometry of the subunit does resemble the build-up of a sieve lobe as described by Rachocki (1981, fig. 34) from a small scale alluvial fan. The lobe developed in the proximal area where sheet floods suffered extreme losses of transport capacity due to infiltration. The extent of the surface layers upwards decreases as in our example. Also the extension on the margins is further than on the top, where also some erosion occurred. The downstream end of the layers represents the "frozen" front of successive sheetfloods (Rachocki, 1981). The composition of the sieve lobe and the subunit are not the same. Gravel and sand are found respectively. However, large-scale climbing bedforms indicate also very rapid loss of transport capacity.

The climbing bedforms on the surface are crescentic. McBride (1962) ascribed a transition of continuous straight-crested to isolated crescentic bedforms in the cross-laminated division of a turbidite to sedimentation during waning flow. Relatively small trough-shaped sets, directed at right angles to the climbing bedform relief, have been described from the three-dimensional exposure. It is likely that climbing bedforms developed first during waning flow. Subsequently low-velocity flow became deflected and confined by the sedimentary relief, building up erosively based trough-shaped sets. The transition of linguoid bedforms on the convex-up top to barchanoid forms at the low margin is a consequence of erosion of the crest of the former structure. Erosion did not occur at the lower margin. Internal structures show a deviating trend. The transition from parallel lamination to tabular cross-beds was promoted by increasing depth towards the margin. However, flow became confined at the margin during the earliest stages and large trough crossbeds were formed. Distally the trough cross-bedding may have passed into tabular cross-bedding of a smaller size; a transition, which was observed in the lateral equivalent to the south (Fig. 9d).

It can be concluded that there are several indications of rapid sedimentation. The geometry and the large-scale bedforms on top of the lowermost subunit point to rapid sedimentation from sediment-laden currents. The transverse bar build-ups and convex layers close to the foreset face also involve rapid sedimentation. The degree of turbulence of the inflow cannot be reconstructed on the basis of the exposures, but it is likely that inertial forces also played a role during the sedimentation process, by creating strong turbulence, keeping large amounts of sediment in suspension.

### III.4.3. *Application of a model of recent highly energetic inflow*

An example of modern highly energetic inflow with resembling relief is found where a hydro-electric station debouches in the Rio Ribagorzana (some 4 km to the south of Puente de Montañana). Shooting flow enters an abandoned reach under a high angle (Fig. 12). Inertial forces cause high turbulence and consequent deceleration in a pit in front of the orifice. Slowed-down flow is deflected towards a shallow channel and enters the main stream. However, part of the inflowing waters is directed back to the orifice forming a return current around a large-scale eddy (Fig. 12). The water–air surface is depressed on the location of the eddy, with an adjacent rise of highly turbulent waters. The surface of the inflow lies higher than the Rio Ribagorzana level. Waters in front of the orifice consist completely of previously discharged waters, because of constant mixing and refreshing.

The model represents very well the deflection of flow due to the perpendicular to oblique entry of the marine basin by the alluvial streams (inset of Fig. 11). It is also clear that high-velocity flow entering a shallow environment from a shallow river

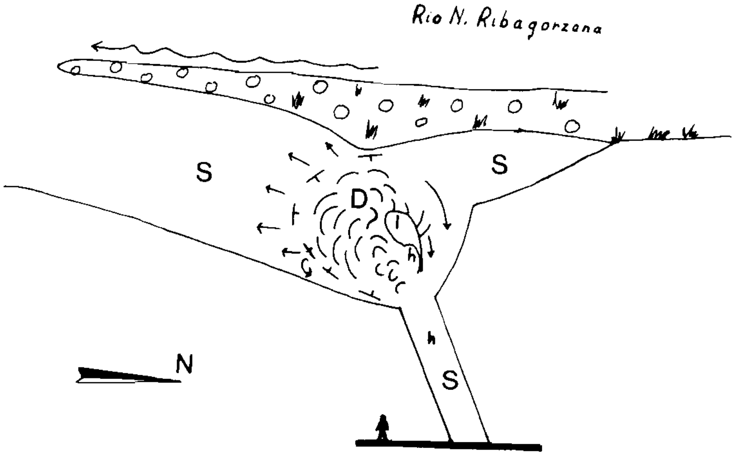


Fig. 12. Inflow in an abandoned reach of the Ribagorzana River from a hydro-electric station. Shallow (*S*) and deep (*D*) parts of the sedimentary relief are distinguished. Further relatively high (*h*) and low (*l*) positions of the level of the inflow are indicated.

must have brought about large-scale turbulence due to inertial forces. Therefore the preceding analysis of three-dimensionally exposed units III and IV can be expanded.

Phase 1: Gravel sedimentation on the steeply inclined foreset (unit III) face was due to avalanching of overriding, transverse bars (phase 3). Expansion at the orifice caused sufficient flow deceleration to prevent gravel transport over a moderately inclined foreset. During phase 1 such transport was possible. Inertial forces decelerated the flow over the foreset face, causing a fining downwards to sand. Concave-upwards sand beds may point to some erosion at the base and consequently to the operation of frictional forces. The large-scale turbulence caused by inertial forces close to the orifice broke down at some distance from the foreset face with consequent dumping of the suspension load.

Phase 2: During further progradation parallel-laminated sand beds draped the initial suspension-load deposits and a rise developed (Fig. 13). On the flat top of the rise, seawards-directed tabular cross-bedding was formed. The same type of deposit forms the northwestern top of unit IV.

Landwards dipping, low-angled lamination in combination with flat-bedded sand may represent berm-flat backshore zones and the backshore zone, respectively (Wescott and Ethridge, 1980). However, palaeocurrent directions from the flat

Fig. 13. The distal parts of units II and III. To the right (south) log 5, Fig. 5 is situated. Northwards dipping, parallel-laminated layers possess flat upper ends with tabular cross-bedding (unit II). Unit III truncates unit II.

tabular sets indicate seawards transport. Further unit IV deposits of this type are embedded in mottled mud and there is no connection with the marine environment.

The deposits inclined against the main palaeocurrent direction resemble backsets. Davis (1890) described backsets in combination with downstream foresets situated in front of a receding glacier. Jopling and Richardson (1966) produced backsets in a hydraulic jump in a shooting flow. Addition of upstream backsets and downstream foresets could be achieved by raising the baselevel. In the case of unit IV raising of the baselevel was accompanied by heightening of the orifice at unit III. Therefore backset sedimentation was due to flow deceleration caused by friction at the sediment rise in combination with mainly vertical expansion in front of the orifice. Unit II backsets coarsen upwards and the top is truncated due to continued progradation.

Phase 3: The indications of rapid sedimentation from unit IV, such as large-scale climbing bedforms, can be better explained when the effects of inertial forces are considered. Large amounts of suspended material became available for sedimentation in the zone where large-scale turbulence broke down in scour pits. The coarser fractions were stored in transverse bars. A more gradual decline of turbulence will have taken place during phase 3a when there was a connection with a marine environment.

A highly turbulent flow keeping large amounts of sediment in suspension may easily pass into a turbidity current, provided that the submarine slope is steep enough. Then an originally fluid gravity flow can pass over into a self-propelled sediment gravity flow (Middleton and Hampton, 1976). The flow was then driven by gravity acting on the sediment instead of on the flow as a whole.

Laminated intercalations of very fine-grained, detrital limestone in between intraformational lags in the deepest, distal parts (phase 3a) may represent respectively the deposits of the tail and the traction carpet of a differentiated turbulent flow. Sneh (1979) describes large amounts of laminated, detrital limestone from the distal parts of a fan-delta.

The relevance of the model for highly energetic inflow is pointed out by another exposure from the northwestern parts of the Monllobat Formation. Mottled sand- and siltstone layers make up layers with a large-scale curvature adjacent to a conglomeratic rise. On the lee of this rise cross-bedding is directed towards the rise. In a minor way such a facies was also present in the northwestern thickening of unit III. The arcuate ridges of cross-bedding (Fig. 14) are lying against low gravel ridges. The ridges are the continuations of the inclined beds on the gravel platform.

The conglomeratic layer can be traced over several kilometers in a transverse direction. To the east the next conglomeratic rise is situated at a distance of a few hundreds of meters. Downcurrent there is a complete transition to graded sandstone beds and mottled fine material. The southern log shows a regressive, fining-upwards sequence. Blue-grey sandstone beds are found at the base. Large foresets with a flat

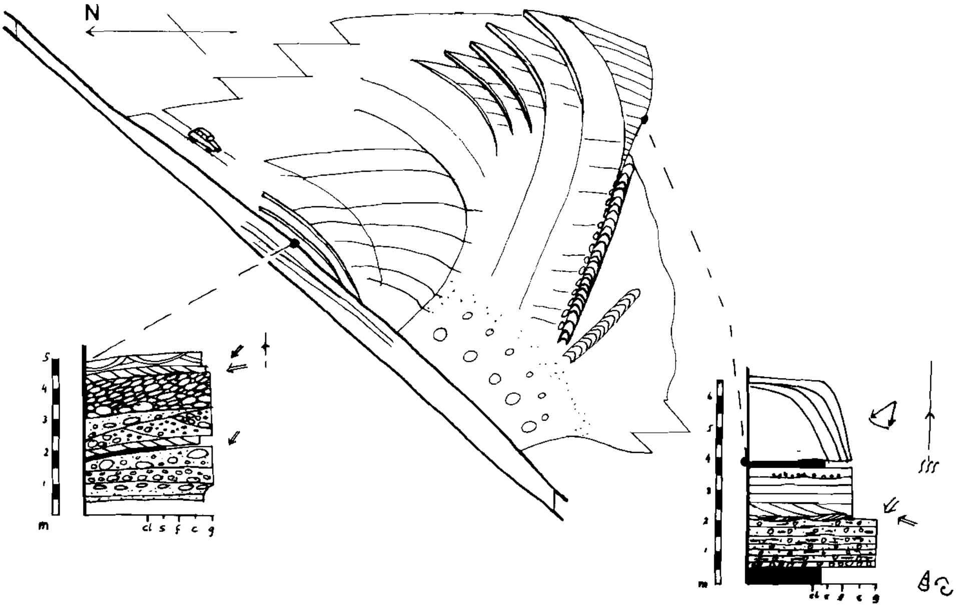


Fig. 14. A conglomeratic sheet passes downstream (to the south-southwest) over in sand- and mudstone. Alongside a massive conglomerate thickening arcuate, inclined layers are found. The continuations of these layers on the lee of the conglomeratic thickening show cross-bedding, developed by return flow.

top grade upwards from sand to silt. The completely mottled terminal foresets on the location of the log are covered by mottled mud. The proximal log shows a highly differentiated conglomeratic facies. The convex-upwards layers pass laterally over into massive conglomerates of the western rise.

## III.5. SEQUENCES

Rapid aggradation exceeding initial water depth took place because of the supply of large amounts of sediment. The differences between proximal and distal sequences were the consequence of a decrease of flow energy and pronounced fining of the sediments in the last stages of fan-delta progradation, when gradients were lowest and finest sediment was supplied. The response of the sedimentary environment has been described in detail above.

## III.6. THE SETTING OF FAN-DELTA DEPOSITS IN STRATIGRAPHIC SECTIONS

Both at the base and the top of sections I and II there are indications (oyster fossils) of the proximity of marine deposits. The stratigraphic interval as a whole represents a period of increased activity of alluvial streams in the area. However, within the interval a gradual decreasing activity can be observed, with extensive caliche formation and a transgression at the top. Such a succession agrees with the model for sedimentary cycles of the Monllobat Formation (Nijman, 1981). Maximal progradation of the alluvial systems will be due to maximal gradients during the transgressive phase.

Point-bar deposits close to the top of section II form a deviation of this sequence. A fan lobe passes downstream over into meandering river deposits as occasionally can be observed within the Monllobat Formation (Van der Meulen, 1982). There are two indications that caliche formation was related to alluvial sedimentation and that there was no distinct link with the occurrence of transgressions, as suggested by Nijman (1981).

(1) Caliche layers are regularly distributed over section II. Similar alternations are found in the distal parts of the upper division of unit IV (Fig. 10) and of the fan lobe succeeding the three-dimensional exposure. The caliche formation may very well have been the result of the evaporation of ground-water discharge through the alluvial deposits. McGee (1897) described migration of rainwater through deposits after rates of precipitation insufficient to cause run off through the channel.

(2) A prominent caliche layer has been developed prior to a transgressive phase—especially at the base of the sections I and II. Instead of being developed during a period of minor sedimentation (Nijman, 1981) this layer may represent the first indication of a steepening of relief due to increased tectonic activity. Subsidence of the basin linked with uplift of the hinterland may have involved deposition of transgressive sediments within the period of time necessary for hinterland erosion. A major regressive phase with deposition of large amounts of newly eroded, coarse detritus succeeded the period of sedimentation of fines. Such a response to tectonic activity has been mentioned by Rust (1979).

From section I to II increasing thickness is combined with decreasing quantities of sandstone and gravel. This may very well have been caused by synsedimentary tectonic activity. Another relation of sedimentation pattern and tectonics is found close to the northwest directed joints in the southern tributary valley. Southwest and northwest directed faults are parallel to the primary and secondary (maximally deflected) palaeocurrent directions, respectively.

The deposits described in detail constitute the base of the sections. A conglomeratic sheet is split up into several lenses in a mud matrix (Fig. 15). The sediments were formed at the mouth of a shallow, braided river during a single fan-delta progradation process. The braided stream made up a part of a small alluvial system, several of which fringed the rising Central Pyrenees. The exposed deposits represented in Fig. 14 belonged to another system situated to the west. The systems transported detritus from the mountain chain to an elongated marine basin. At the time of deposition the eastern termination of the basin was situated in the vicinity of Puente de Montañana. The climate was tropical with marked dry and wet periods (Van der Meulen, 1982).

Wescott and Etheridge (1980) described two fan-delta types, representing end

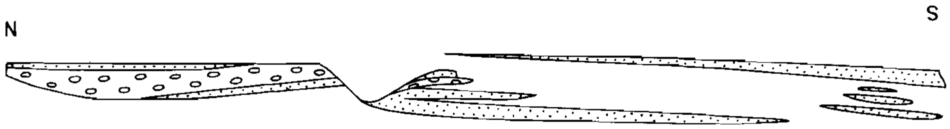


Fig. 15. A generalised N–S section through the investigated interval showing the morphologies of the conglomerate and the sandstone facies.

members of a spectrum. In this case we are dealing with the end member; more completely developed, subaerial fans that prograde into shallow marine water bodies. In contrast, the other end member comprises truncated subaerial fans that prograde onto relatively steep submarine slopes. The description of a Pliocene fan-delta from southeast Spain by Postma (1983) seems to agree with this model.

## III.7. CONCLUSIONS

The sedimentary relief, inflow conditions and sediment movement in an Eocene transitional fan-delta environment have been reconstructed. The study of three-dimensionally exposed deposits was the basis of the reconstructions. Inflow took place from a shallow, braided stream. The low sedimentary relief allowed for jetflow from orifices of limited extent, which involved the formation of interconnected or separated lobes. Tabular cross-bedding and parallel lamination are the main sedimentary structures. Large-scale climbing bed forms indicate rapid sedimentation.

A large range of grain sizes (including gravel) is found in the shallow, braided channels. The sediment was transported by highly-energetic flow during high discharge periods. Inflow models for the deltaic environment cannot be applied, because of very high flow energy, the large range of grain sizes and the availability of large amounts of sediment. After comparison with a model of recent highly energetic inflow, the following interpretation of inflow hydrodynamics and sediment movement was made:

(1) Gravel was deposited close to the orifice. Flow expansion and inertial forces caused the formation of Gilbert-type deltas with moderately inclined foreset faces. During abandonment stages, flow expansion led to a steeply inclined foreset face.

(2) Frictional and inertial forces were in variable degree responsible for flow deceleration; this caused sand and mud sedimentation on a flat slope or in scour pits.

(3) Backsets formed due to expansion and frictional forces.

Gravel graded to sand downwards over the moderately inclined fan-delta face. Therefore a coarsening-upwards sequence of proximal deposits developed during progradation. In the course of progradation, large amounts of sand and mud were deposited in front of the fan-delta. After a phase of deep incision of these deposits, gravel supply diminished and aggradation of the stream channel took place. Large amounts of mud were deposited in front of the fan-delta face. The extent of incised sand lobes decreases upwards and therefore fining-upwards sequences are found in the distal parts. The top of the sequences is formed by a thin, alluvial fining-upwards sequence. Aggradation exceeded initial water depth.

The deposits originated in the transition zone of a fan-delta, where the distal parts of a wide, braided system contacted a restricted part of a marine basin. The basin was elongated in a direction perpendicular to the main alluvial palaeocurrent direction and therefore deflection of the inflow could take place. There was no reworking by marine agents. Only during periods of accelerated tectonic activity major amounts of gravel were deposited in the transition zone. There was then a rapid regression of the marine environment.

# Sedimentary stratigraphy of Eocene sheetflood deposits, Southern Pyrenees, Spain

## IV.1. Introduction

The Eocene Monllobat Formation is part of the Montañana Group of the Tremp–Graus Basin. This elongate basin lies at the southern flank of the Pyrenees (Fig. 1). The east–west trending Sierra de Monsech separates the Tremp–Graus Basin from the Ager Basin to the south. The continental Monllobat Formation consists of material eroded during uplift of the Pyrenees to the north and sediment transported out of the Ager Basin over the Monsech area (van Eden, 1970; Nijman & Nio, 1975; Nijman, 1981). The marine Castigaleu Formation was the dominant deposit in the west and northwest of the study area (Fig. 1).

Recently north–south flowing rivers have cut deep valleys in the Tremp–Graus Basin fill. One of these rivers, the Rio Noguerra Ribagorzana, and its tributaries have provided a number of well exposed sections of the two Formations. Sheets of conglomerate and sandstone, several kilometres wide, stand out in the thick mudstone and siltstone sequences of the valley walls. Lenticular conglomerate and sandstone bodies occur especially along the distal margins of the sheets, but are also found at fault and flexure zones all over the area.

In this paper the architecture (i.e. the geometry and relationships of the sediment bodies; Collinson, 1978*a*; Miall, 1983) is described and interpreted. The emphasis lies on the architecture of Monllobat deposits, supplied from the axial Pyrenees.

## IV.2. The setting of the Monllobat Formation

The marine Castigaleu Formation underlies, with a sharp contact, the continental Monllobat Formation and interfingers with it. The Formations are distinguished on the basis of mudstone colours (Nijman & Nio, 1975), mainly blue and red mottled brownish in the case of the Monllobat and blue-grey in the case of the Castigaleu Formation. Also caliche nodules are typical of the Monllobat fine members and fossil oysters are widely distributed in the Castigaleu sandstones.

Conformably below the Castigaleu Formation lies the Ager Formation. Above the Castigaleu and Monllobat Formations is the Castisent Formation which interfingers in this area with the Capella Formation; both originated in an alluvial environment. The Campodarbe Group lies on an incised contact, which has especially high relief at the northern margin of the basin fill (Fig. 1).

Outcrop conditions of the Monllobat Formation are generally very good, especially of the conglomerate and sandstone coarse members. In contrast, the Castigaleu coarse members, consisting of sandstone only, are badly exposed except close to the Monllobat Formation.

The dip of the strata is lowest (1–2°) along the basin axis running near Puente de Montañana, and highest (up to 10°) in a narrow zone along the northeastern and southern basin margins (Fig. 2). The dip is 4–6° in the other parts. The Formations thin strongly at the southern margin and the Monllobat Formation thins also to the east at Monllobat.

## IV.3. Previous sedimentological research

Earlier interpretations of the Monllobat Formation comprised meandering rivers on a deltaic plain with minor gravel incursions from Pyrenean alluvial fans (van Eden, 1970; Nijman & Nio, 1975). Puigdefabregas & van Vliet (1978) described some of the deposits in an article on meandering river deposits of the southern Pyrenees. During continued research it emerged that Pyrenean supply has played a considerable role in building the Monllobat Formation (van der Meulen,

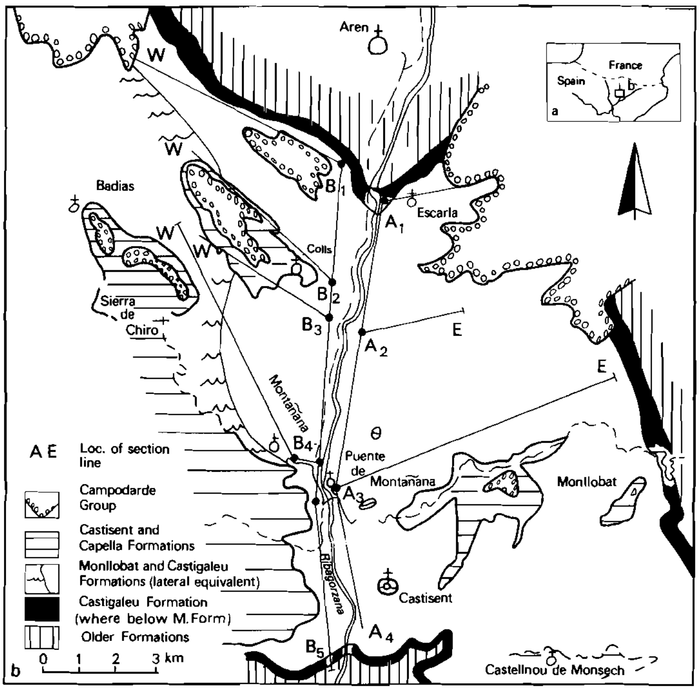


Figure 1. (*a*) Position of the study area. (*b*) Map of the eastern part of the Tremp–Graus Basin with the location of the section lines.

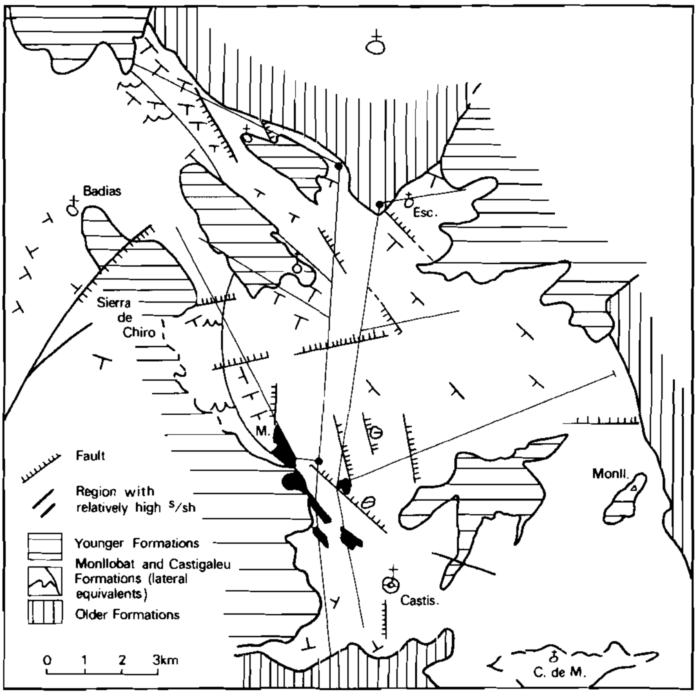


Figure 2. Distribution of fault and dip directions. The blackened area between Castisent (Castis.) and Montañana (M.) is a synclinal zone with high sand/shale ratios.

unpub. M.Sc. thesis, Univ. Leiden, 1978; Vreeburg, unpub. M.Sc. thesis, Univ. Leiden, 1978; Frikken, unpub. M.Sc. thesis, Univ. Leiden, 1980; compilation and additional data in Nijman, 1981). The association of fine member sequences (70–80% of the deposits), both with meandering channel deposits and Pyrenean-derived conglomerates, was also established (van der Meulen, unpub. M.Sc. thesis, Univ. Leiden, 1978; van der Meulen 1982, 1983).

## IV.4. The lateral arrangement of the Monllobat deposits

### IV.4.1. Description of the deposits

Most of the coarse members accumulated in sheets in the northern and central parts. The sheets are up to a few kilometres in width and their thickness is mostly 4–5 m with maxima of 10 m. Symmetrical channel fills with cross-sectional sizes of a few to tens of metres are up to 5 m thick. Both sheets and channel fills are composed of conglomerate and/or sandstone. Asymmetrical channel fills (meander lobes made up of point bar accretion) and lobes consisting of foresets are of comparable size and composition. The widths of cross-sections of these bodies are up to a few hundreds of metres, and thicknesses range between 2 and 7 m. The composition is dominated by sandstone and siltstone and the sequences are fining-upwards.

The conglomeratic parts of sheets generally have a flat base and top and consist of massive conglomerate. Well-established channel fills are very scarce. Sandstone margins possess mostly rather flat channel fills. Sheets embedded in Monllobat mudstone can split into symmetrical channel fills downstream. Small channel fills especially have extensive wings of siltstone and occur in well bedded parts of the fine sediments. In a few cases the symmetrical channel fills have been found upstream from sheets. Asymmetrical channel fills and foreset lobes are situated in the distal parts or downstream from sheets.

Most of the Monllobat fine sediments are brown or ochre and possess multiple vertical blue and/or red traces. There is a regular intercalation of nodular carbonate caliche layers. Less extensively distributed yellow mudstone has blue or grey traces and contains small amounts of gypsum caliche.

Sheet margins interpreted as fan-deltas are composed of well defined top-, fore- and bottomsets (van der Meulen, 1983) or of tabular crossbedded sandstone. Underlying blue-grey mudstones and oyster fossils in sandstones indicate a connection with the main Castigaleu Formation. In some cases there are no marine fossils and the fan-delta bodies on top of some blue-grey mudstone are surrounded by mottled mudstone of the Monllobat Formation. The sections of the Castigaleu Formation close to the Monllobat Formation contain sandstone layers which are flat bedded (decimetre-scale). Several beds are parallel laminated. A detailed account of sedimentary structures and coarse member build-ups is presented elsewhere.

### IV.4.2. Lateral subdivision of the Monllobat Formation

The good outcrop conditions, the low dip of the layers and the well established base and top of the Monllobat Formation favour the construction of a diagram showing the distribution of the coarse members in the river valley sections (Fig. 3). The well exposed part of the Castigaleu Formation is also incorporated. The positions of the sections are indicated in Figure 1, and it is clear that sections $B_2W$, $B_3W$ and $B_4W$ contain sediments of both the Monllobat and laterally equivalent Castigaleu Formations. On Figure 1, the position is shown of the lateral boundary of the Monllobat fine sediments (to the east) and the Castigaleu fine sediments (to the west).

Coarse member sheets generally lie at particular levels and without lateral overlap. Three sheets in the eastern area illustrate this type of lateral arrangement. Two of the sheets are completely present in section $A_3E$ (level 3). Only a tip of the third sheet is indicated at the eastern margin of the section; however this conglomeratic sheet is exposed in a kilometres-wide section at Monllobat (location in Fig. 1, information from van der Meulen, unpub. M.Sc. thesis, Univ. Leiden 1978; Frikken, unpub. M.Sc. thesis, Univ. Leiden, 1980):

Two major sheets (both accompanied by lenses) in section $B_1W$ at level 2 are separated by a fine member zone. This zone is located centrally on the crest of a gentle anticline (expressed at level 1, Fig. 3) and it is remarkable that, at level 2, the fine sediment is yellow to the northwest and brown and ochre to the southeast of the anticline crest. Another major sheet at level 2 wedges out close to $B_1$ in the north–south section $B_1B_5$. At Monllobat there is also a major sheet at level 2.

The latter sheet, consisting of sandstone, is completely overlapped by the position of the earlier mentioned conglomerate sheet at level 3. The somewhat smaller sandstone sheet lies off centre under the central and western parts of the conglomerate sheet only. In the western parts of the $B_3W$ section three conglomerate sheets (at levels 3a, 4 and 4a respectively) overlap each other. The same goes for sheets at levels 1a and 3 in the eastern parts of section $A_3E$ and at levels 3a and 4 in the $A_3A_4$ section.

The general palaeocurrent distributions for the eastern (Figs. 4a, b) and western areas (Figs. 4c, d) show a modal direction for the conglomerate facies to the southwest. Crossbedded sandstones also have western and northwestern components. In the conglomeratic cores of the sheets the palaeocurrent directions are to the southwest; in the sandstone margins they are to the west and northwest. These trends involve perpendicular and oblique cross-sections of sheets in the present day tributary valleys and more or less longitudinal sections in the Ribagorzana Valley, especially in the northern parts.

By plotting the sheet positions, together with sheet margins, and the palaeocurrent directions, eight

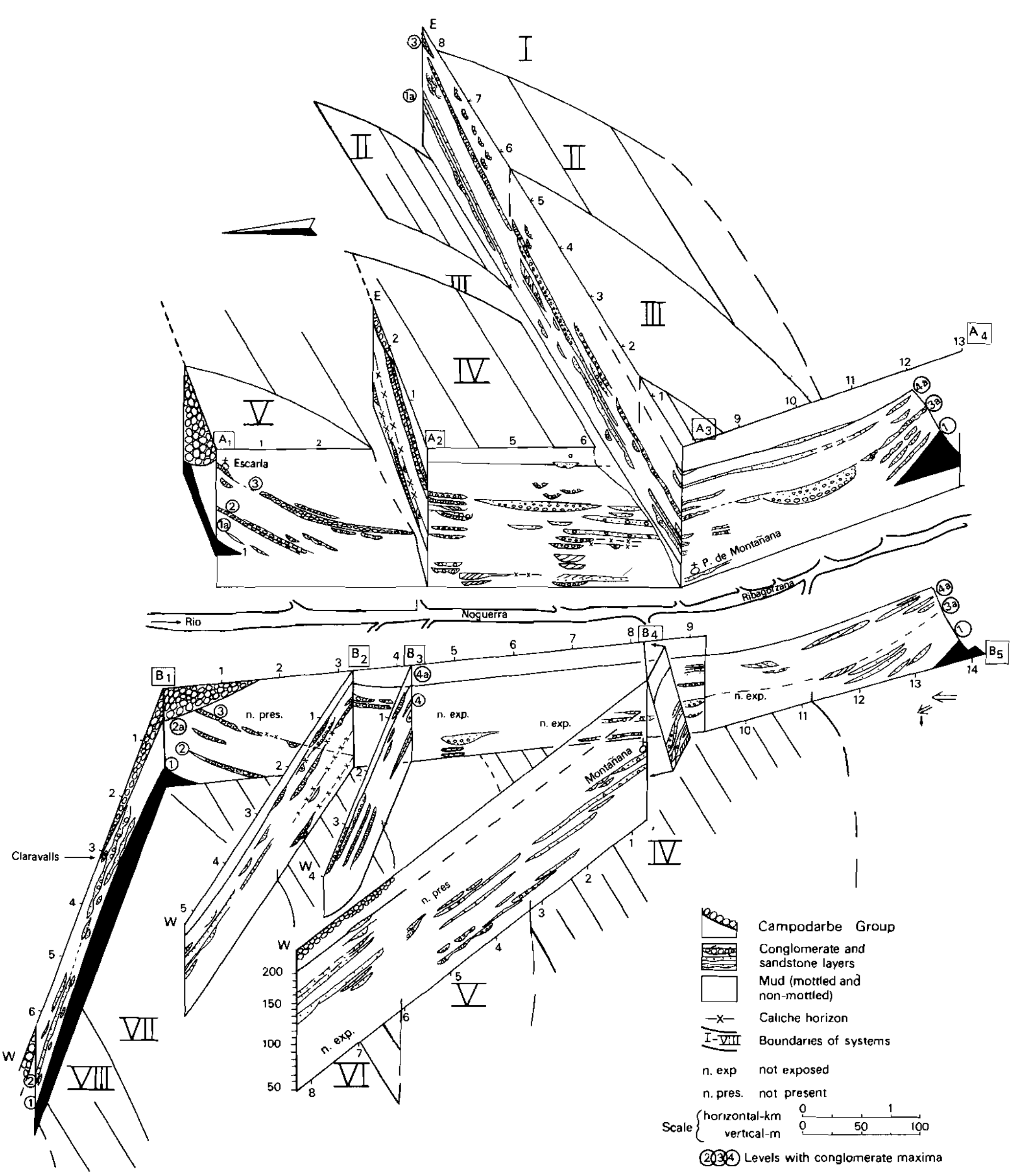


Figure 3. The distribution of conglomerate and sandstone in sections along the Ribagorzana Valley and some of the tributary valleys (locations in Fig. 1). The dip of the layers approximates the real dip except for an exaggeration at the margins of the formations. Several zones (systems I–VIII) as well as stratigraphic levels (1–4a) with concentrations of the coarse member are indicated. Displacements at faults have been corrected.

parallel systems of deposits can be defined. These systems (I–VIII, Fig. 3) form laterally distinct but contemporaneous parts of the Monllobat Formation. The systems are defined primarily by the location of sheets in the same vertical succession, by lack of overlap of laterally adjacent sheets and by the fixed overall palaeocurrent pattern. The mudstone colour can be an additional criterion.

The palaeocurrent pattern shows that there was a gradual bend in the systems in the distal areas, where the conglomerate to sandstone transition takes place (Fig. 3). In these areas there is also no interfingering of the systems even when sections of conglomerate and sandstone sheet parts succeed one another. This is due to the relatively small size of sandstone sheet parts which then lie off centre with respect to the larger

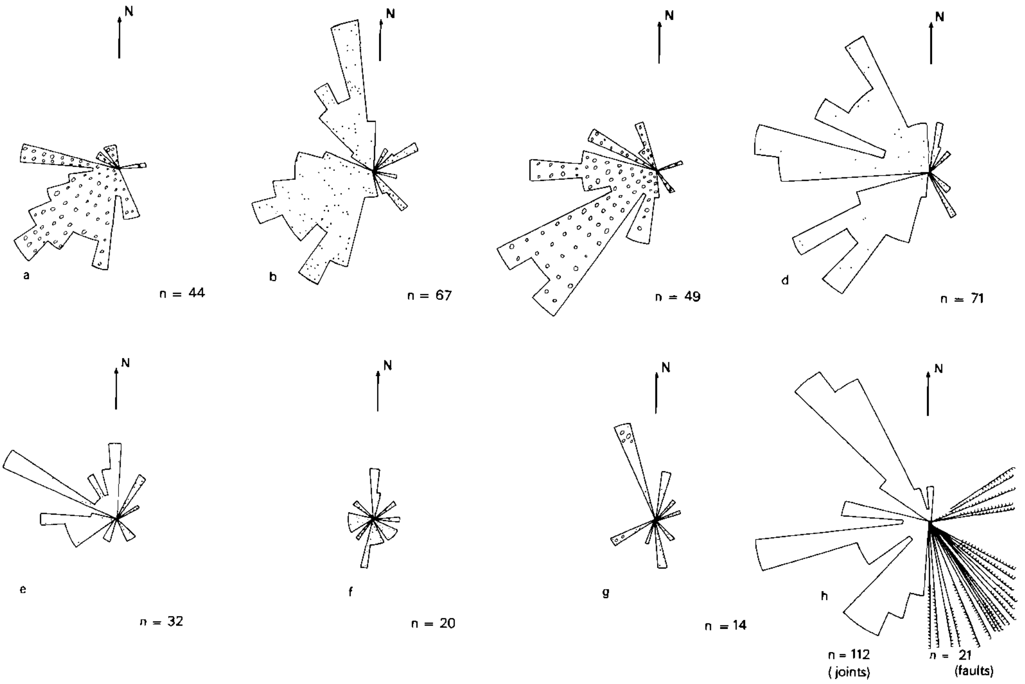


Figure 4. The distribution of palaeocurrent directions. (*a*) and (*b*): western area; (*c*) and (*d*): eastern area; (*e*): southern Ribagorzana Valley; (*f*): area to the north of Badias; (*g*): area along fault (80–260° trend); (*h*): joint directions in the western quadrants and fault directions in the eastern quadrants.

conglomerate sheets. In the clusters of coarse member lenses that accentuate sheet positions, the conglomerate lenses lie underneath or above the southeastern margins. The sandstone lenses are situated near the centre or to the west.

The positions of the systems I–VIII have been projected onto a horizontal plane (Fig. 5). The downstream boundaries of sheets and clusters of lenses are indicated within the plan areas of the systems for levels 1a–5. Although not entirely uniform, there is a tendency for stepwise progradation followed by retreat within the systems. Two sheets extend beyond the generalized boundaries (system IV, level 3 and system V, level 2). Clusters of symmetrical channel fills constituting upstream sheet equivalents (especially in system IV, sections $B_2W$ and $B_3W$ in Fig. 3) have not been incorporated in Figure 5.

In Figure 5 a large area with northeastern supply can be recognized. A small area with southern supply lies at the southern margin of the basin along the north flank of the present Sierra de Monsech; in this area palaeocurrent directions indicate transport to the north and northwest (Fig. 4e). A subsidiary component to the west comes from downstream margins of coarse member layers. There is a zone with extra high fine sediment content (80% instead of 70%, van der Meulen, unpub. M.Sc. thesis, Univ. Leiden, 1978; Van der Meulen, 1982) in between the two areas. This zone can be traced towards the east through the valleys to the south of the road from Puente de Montañana to Monllobat (see Fig. 1 for locations), and this area has been defined as a mud zone by Nijman (1981).

### IV.4.3. Interpretation of the type of alluvial system

The general geometry and nature of Monllobat deposits point to an origin in coarse and fine sheetflood systems, as defined by Friend (1983) in a classification of alluvial systems. The transition of conglomerate sheets to mottled, flat sheets of silt- and mudstone in a down-palaeocurrent fashion in the systems is best illustrated in Figure 8b, discussed below. Transitions of conglomerates to large amounts of blue-grey and mottled fine sediments are also found at fan-deltaic sheet margins (van der Meulen, 1983).

Sheetflood systems are known from distal alluvial fans (Collinson, 1978*b*). The sheetfloods develop in that environment due to the widening of the channels incised in the proximal and midfan. The energy loss associated with the widening causes sedimentation and the consequent formation of a shallow braidplain. The transition of deeply-incised to shallow-braided channels forming a braidplain is a common feature of outwash fans (Boothroyd & Ashley, 1975). The presently exposed part of the Monllobat Formation comprises only the distal sheetflood deposits.

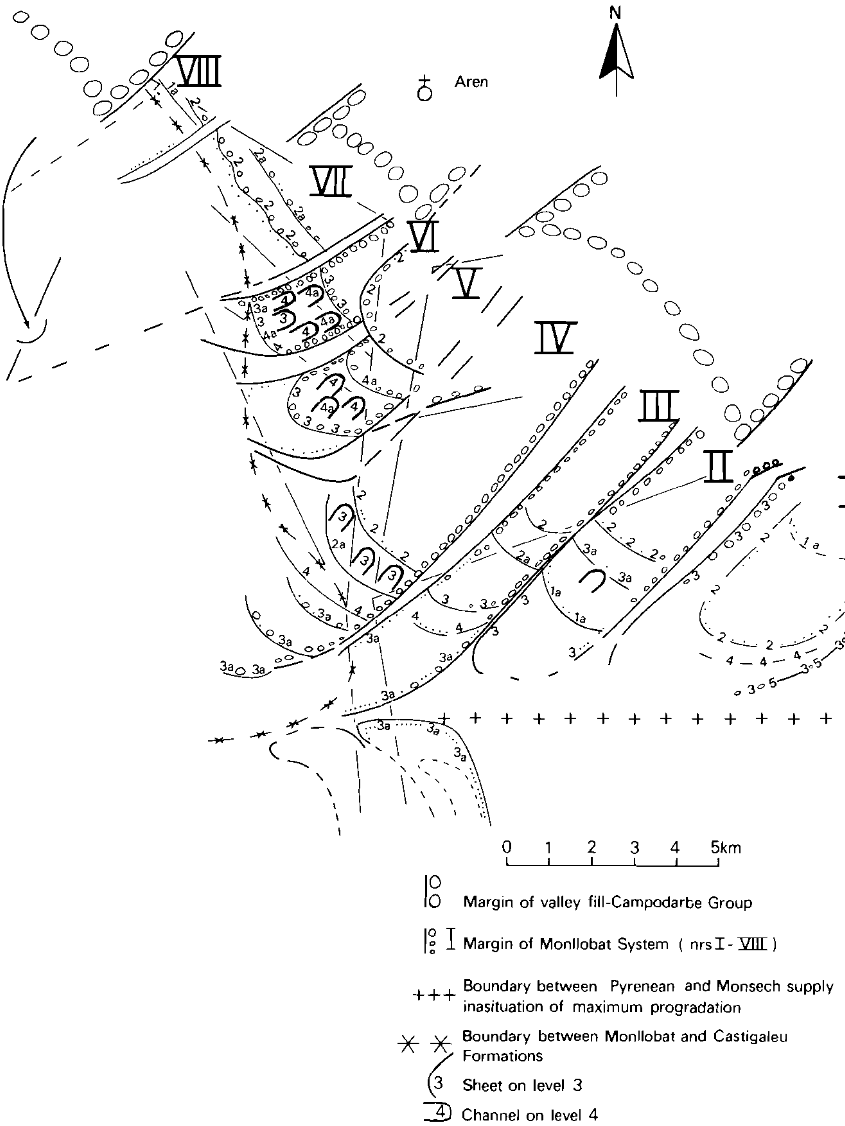


Figure 5. Horizontal projection of the system boundaries. The distribution of sheets at several levels is given as well as major clusters of channel fills. Conglomerate sheets of systems IV and V pass the marginal mudstones at levels 3 and 2 respectively.

The deposition of large amounts of fines is not a general characteristic of sheetflood systems, but is known from several recent and ancient environments with ephemeral but highly energetic streams. Parkash, Awashti & Gohain (1983) describe a complete downstream fining on a terminal fan, which comprises channelling and associated sheetflooding during sedimentation events. The sand-filled channels shallow downstream together with the fining. Well structured lobes of the finest sediment adjacent to graveliferous outwash are described by Gustavson, Ashley & Boothroyd (1975). Thick, massive fine members in association with ancient ephemeral stream deposits are described by Stear (1983) and Wells (1983). Allen (1981) developed sedimentation models resembling proglacial outwash plains for the Wealden of southern England. These sandy alluvial systems with distal fine member areas originated under a warm climate with periodic heavy rains. Graham (1983) and Tunbridge (1984) describe a combination of massive clay and well structured fine-grained sheetflood deposits in the distal areas of Devonian ephemeral streams. Heward

(1978*a*) and Hubert & Hyde (1982) describe ancient distal fan associations that only comprise the well structured sandy deposits, a type which is very scarce in the Monllobat fine member.

Point bar deposits of meandering rivers, associated with a fine member, are a minor feature of the distal areas of the Monllobat coarse member sheets. Point bar deposits are also incorporated in the well- and non-structured distal fan association described by Tunbridge (1984). Mixed- and suspended-load rivers deposited point bar sediments (Stewart, 1981, 1983) in the distal areas of the alluvial systems of the Wealden mentioned ealier. Finally distal meandering rivers are incorporated in the model for recent Alaskan outwash (Boothroyd & Nummedal, 1978).

The Monllobat fine member contains frequent clusters of symmetrical channel fills and fine member sheets down palaeocurrent from coarse member sheets. However in other cases clusters lie downstream from intervals of the fine member and also most of the fine member layers have no connection with the coarse member in the presently exposed parts of the Monllobat Formation. It is concluded that in some cases the sheetflood systems consisted of a wide graveliferous braidplain terminating in a muddy floodplain, possibly with symmetrical or meandering channels. However for much of Monllobat time this muddy floodplain covered the area of the systems completely, with only occasionally some selective gravel and sand deposition in the downstream areas and in the coastal plain. The sand–silt lobes, consisting of foresets, originated also in this area. A detailed study has shown that examples of this type were deposited on small-scale fan-deltas in a subaerial environment (van der Meulen, 1983).

### IV.4.4. Reconstruction of the depositional environment

Patterns of ancient multiple alluvial systems that have been analysed vary from divergent through parallel to convergent (respectively; Allen *et al.* 1983; Turner, Hirst & Friend, 1984; Tyler & Ethridge, 1983 – Campbell, 1976 – Behrensmeyer & Tauxe, 1982; Read & Dean, 1982). In the case of the Monllobat Formation the existence of eight linear systems of deposits points to the presence of eight sheetflood systems side by side and in fixed positions through time. Campbell (1976) has described parallel braided systems, incising each other's margins and thereby constituting a continuous sandstone sheet. Lateral deposition of fines prevented the connection of Monllobat sheets.

Read & Dean (1982) have shown that basement structures and differential subsidence can influence the courses of streams in sedimentary basins. In the case of the Monllobat systems a gentle anticline marks the boundary between systems VII and VIII. Further there is a major agreement between the fault/joint pattern and the distribution of palaeocurrent directions. However there are no faults which line the system boundaries and there are no indications of thickness variation perpendicular to the system axes. It is therefore most likely that the pattern of the systems was a primary factor of the sedimentation process. The presence of eight systems then points to the presence of eight feeder canyons in the Central Pyrenees.

In this respect it should be noted that the walls of three canyons from the later Campodarbe stage are present, incised in the northeastern margin of the basin fill. After uplift of the basin margin the discharge of the former eight Monllobat canyons became then concentrated in only three major valleys, as indicated in Figure 5. This will have enhanced the change from parallel to divergent systems (compare Turner, Hirst & Friend, 1984).

The Monllobat sheetflood systems lay fixed through time forming a bajada. Rachocki (1981) explains the flow distribution on alluvial fans and outwash plains using a model. A wide shallow stream forms with continuous flow division around diamond-shaped bars. Naturally there is also, but to a minor degree, flow rejoining at the bar terminations, in a given situation along the most favourable stream course. Consequently an axis with maximum discharge is formed. If there is no lateral restriction the axis will be able to shift over the sedimentation plain at appropriate times leading to fan formation. The lateral shifting of the Monllobat system axes was probably restricted by the well balanced distribution of sedimentation rates along the mountain front.

The bending of the terminal axes, together with the sediment fining, suggests changes of the gradient direction and magnitude in the central basin area. However the same bending together with fining is also found in vertical successions within sheets in this area. It is therefore likely that only weakened flows adapted to the altered gradient direction. Flow weakening will have been due to lowering of downstream gradients combined with the deceleration effects of infiltration and evaporation of discharge (Friend, 1978; Parkash, Awashti & Gohain, 1983) and to the widening process (Rachocki, 1981).

The eastern sheetflood systems were completely surrounded by laterally and distally deposited mud – the longitudinal and transverse mudzones respectively. The central sheetflood systems terminated in either continental environments or fan-deltas, depending on the eastwards extension of the marine environment. The western systems discharged perpendicular to the coast, deflection of the current direction and grading from gravel to sand taking place in the coastal zone. There are strongly aggrading transitional fan-delta deposits in the coastal zone (van der Meulen, 1983), similar to those known from shallow bodies of standing water (review by Church & Gilbert, 1975). The facies and the setting of the Castigaleu sandstones

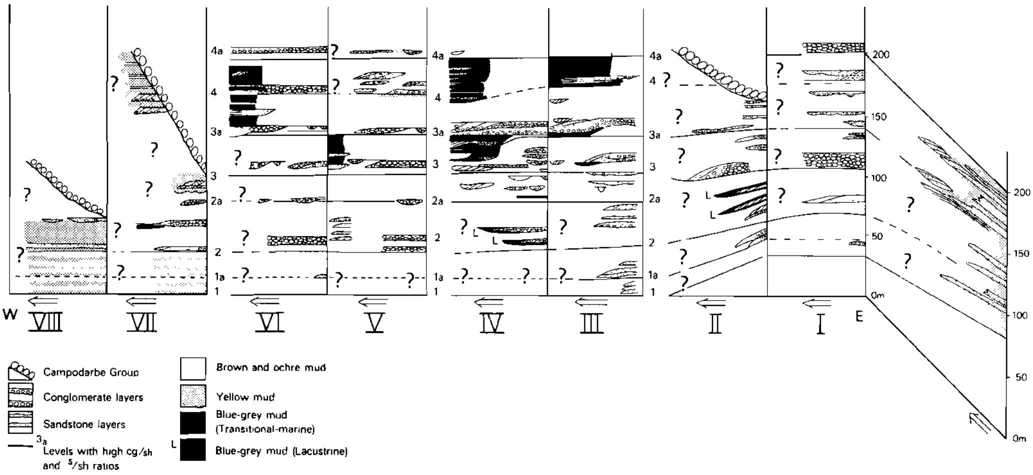


Figure 6. The distribution of the coarse members in sections along the axes of systems I–VIII and in the southern part of the Ribagorzana Valley. Double arrows indicate the palaeocurrent trend. The distributions of yellow and blue-grey mudstones are also given.

(lying in front of the northeastern Monllobat systems) point to minor redistribution in the former marine environment. The immature type of mineralogy (Fig. 7) agrees with such an interpretation.

To the south of the transverse mudzone there were northward-flowing sheetflood systems of a much smaller size. Distally their flow became deflected to the west. The southern and northeastern systems competed with each other, similar to the situation for the Oligocene Pyrenean and Iberian streams of the Ebro Basin, described by Allen & Mange-Rajetzky (1982). In the upper Monllobat Formation, the extent of the southern systems increased at the expense of the northeastern systems.

## IV.5. The vertical arrangement of Monllobat deposits

### IV.5.1. Vertical subdivision of the Monllobat Formation

The coarse member sheets occur at several levels within the Monllobat Formation. Twenty-six stratigraphic sections from the study area confirm this. Figure 6 summarizes the information in cross-sections along the axes of the systems I–VIII. Overall three main levels (2, 3 and 4) can be distinguished between the base of the Monllobat Formation (level 1) and the main sandstone body of the Castisent Formation, both of which are sandstone maxima.

The levels 2, 3 and 4 are accentuated by mudstone types, which deviate from the main fine member in between the levels. Underneath the sheets, blue-grey mudstone can be found to extend for kilometres eastwards and northeastwards from the Castigaleu boundary, as indicated in Figure 1. These blue-grey transgressive mudstones lie on top of an extra thick component of the caliche layers, which are regularly intercalated in the fine member. In a few cases yellow mudstone partly replaces the transgressive mudstone.

Four sublevels, approximately in the midst of the intervals bounded by the five mentioned levels, show layers of conglomerate and sandstone which are mostly of a much smaller size than those at main levels. Transgressive mudstone at the base of these sublevels is also less widely distributed than at the main levels in most cases. Nevertheless locally sublevels have the same facies as main levels (e.g. sublevel la, systems II and III and sublevel 3a, systems II–IV; Fig. 6).

### IV.5.2. Cycles

The Monllobat Formation comprises seven intervals, termed cycles, which consist of a thick fine member on top of a thin coarse member and often transgressive mudstone.

Sublevel 4a has been taken as the top of the Monllobat Formation, because of the sudden increase of southerly-derived material, marked by white sandstone layers. The maximum extent of this distribution is the main white sandstone body of the Castisent Formation, which covers a large part of the Monllobat Formation and extends as far as Colls in the northwestern area (Fig. 1). The vertical distance between the thin sheets and lenses at sublevel 4a and the main white sandstone body is 20 m, approximately the same as the vertical distance between levels 4 and 4a. For this reason level 5 is defined at the base of the main white sandstone body of the Castisent Formation.

Below sublevel 4a there is only a minor peak of the coarse member extent in the southern area at sublevel 3a (Fig. 3). At this sublevel, a sheet of symmetrical channel fills lies on top of more steeply dipping flat bedded silt- and mudstone layers in the southeastern wall of the Ribagorzana valley. The inclined fine member beds are not truncated by the upper layer, instead several possess flat upper ends underneath the sheet base. A small scale example of such a setting is illustrated in Figure 9a. Apart from the sublevel 3a the levels la-4 cannot be established in the southern area because of the very scattered distribution of the coarse members.

In the central and northern parts, the cycle thickness increases up to level 4 and then decreases. There is a parallel trend of marine regression up to level 4 succeeded by transgression. In the eastern area, at Monllobat, the interval thicknesses decrease suddenly above level 2 (Fig. 6).

### IV.5.3. Megasequences

The overall coarse member distribution varies laterally within the levels, both along and perpendicular to the southwestern direction of the system axes. In the northeastern parts, the simple two-tier cycle is commonest. In the central parts of the basin however, the coarse member sheets wedge and several lenses can be present. Downstream of the sheets several lenses can succeed one another. Coarse member successions in these parts show coarsening- or fining-upwards megasequences.

The megasequences are found especially at levels where the coarse member development is greatest. This distribution varies from system to system. Conglomerates of the eastern systems I–IV are concentrated in the central, thickest cycles. Conglomerates are rather evenly distributed over the cycle bases of systems V and VI. Further to the west coarse members devoid of conglomerate become of increasing importance in a combined upwards and eastwards trend. Such coarse members are accompanied by a yellow fine member. In systems VIII and VII such deposits are found at levels 2 and 3 respectively.

### IV.5.4. Development of cycles

The two-tier build-up of cycles can be explained by assuming periodic short accelerations of tectonic activity. The build-up of cycle bases can then be interpreted in terms of the model of Rust (1979).

(1) Acceleration of tectonic activity caused destablizing of the intensely weathered material in the feeder canyons. During rainstorms this material became stored in relatively thick siltstone sheets in the basin.

(2) Continuation of accelerated uplift/subsidence was accompanied by the accumulation of large quantities of material in canyons of the mountain range, which had temporarily a high relief. Extensive caliche formation in the basin was succeeded by transgressive mud sedimentation, when the minor sediment distribution could no longer compensate basin subsidence.

(3) The material in the feeder canyons reached maximum instability and distribution of the bulk of coarse material followed.

The descriptive section indicates a general applicability of this model which was first used for a few exposures from the transitional area (van der Meulen, 1983). Yellow mudstone at cycle bases is an additional feature. A link with hinterland composition will be established later.

The alternation of higher and lower rates of system progradation and marine transgression point to a similar rhythm in tectonic activity. This must have been due to tectonic processes on a regional scale too large for reconstruction on the basis of this study. The same goes for the basic tectonic processes leading to the increasing–decreasing trend of cycle thickness and the simultaneous overall trend of regression–transgression. In the latter case it is evident that both basin subsidence and Pyrenean uplift first increased simultaneously and then decreased simultaneously. There may have been a link with subsidence processes of the Ager Basin to the south, because southerly supplied material was ultimately distributed widely over the basin.

### IV.5.5. Cycle base variability and megasequences

The thickness of the basal cycle interval varies considerably because of the occurrence of multiple sheets and because the vertical position of sheets is somewhat variable laterally over the levels from system I to VIII (Fig. 6). Therefore other environmental factors must have been varying besides the broad regional tectonic conditions.

#### IV.5.5.1. *Timing and variability of sheetflood system sedimentation*

Sedimentation of the sheetflood systems must have been episodic as the necessary large quantities of material will only occasionally have been available in the feeder canyons. Furthermore, the large quantities of water needed to mobilize the material will also only occasionally have been precipitated, by rainstorms. The instability of weathered material in the feeder canyons of alluvial fans depends partly on the amount of material being supplied, but will also have been increased by major increases in precipitation (Heward, 1978*b*). Timing and extent of the sheetflood sedimentation along the Pyrenean mountain chain will therefore have varied with variable timing and precipitation rates of rainstorms as well as with the uplift rates of the mountains, which supplied the weathered material.

The variability of the latter has already been discussed. Episodic discharges are indicated by two characteristics of the deposits studied. First, point bars and marine transition fan-delta deposits show a regular alternation of sedimentation units, erosively based and often on top of mudstone layers, which occasionally are covered by mud pebbles (van der Meulen, 1982, 1983; Puigdefabregas & van Vliet, 1978). second the pseudogley-type of soil formation involves marked periods of dry and wet conditions under a warm climate (van der Meulen, 1982).

#### IV.5.5.2. *The effect of progradation processes – megasequences*

Progradation processes of the sheetflood systems were important factors in building megasequences, as is generally the case (Heward, 1978*b*). The variation of progradation rates depends on the precipitation/material supply rates (see also Heward, 1978*b*) as stated before. However the distribution areas of the Monllobat systems are not equally large because of the basin shape. Furthermore, some of the areas were partly subaqueous and there was differential subsidence along and perpendicular to the basin axis. These conditions must also greatly have influenced the progradation processes.

Megasequences are especially found in systems II–IV in the middle parts of the basin. These systems had the largest continental distribution areas. Distally the coarse sheetflood sediments covered the areas in more than one phase during maximum progradation events. This led to stacking of coarse member bodies in the transverse mud zone. The distal areas of systems III and IV intermittently possessed shallow offshoots of the marine environment and these were also completely filled with megasequential coarse member build-ups. Relatively rapid and slow progradation will have developed fining- and coarsening-upwards sequences in narrow and wide depressions respectively.

Within the smallest distribution areas, the completely continental system I, as well as the partly subaqueous systems V–VIII, gravel and sand were left in single sheets on the floodplain at all levels. In the subaqueous parts of systems V–VIII sand was distributed widely by high energy, heavily loaded flows and became deposited in thin sheets.

Differential subsidence can also strongly influence progradation processes (cf. Heward, 1978*b*). Hooke (1972) demonstrated that on a tilted basin floor alluvial fans are small at the relatively subsiding side and large at the opposite side. The cycle thickening laterally from system I to II and III might indicate tilting along the basin axis. However the palaeocurrent pattern does not show a corresponding deviation and therefore tilting of the basin surface in an east–west sense is not likely. Tilting perpendicular to the basin axis can explain upstream wedging and foresetting in the coarse and the fine member of the southern fringe. Major thinning towards the basin margin indicates the lowest relative subsidence rates. Finally it may be that thinning of the deposits now no longer present on the northeastern margin was a factor in the large-scale periodic progradation of the northeastern sheetflood systems.

## IV.6. Hinterland composition and variability of sedimentation

### IV.6.1. Petrography

The composition of the northeasterly-derived gravel is dominated by micritic limestone clasts. Small amounts of vein quartz pebbles are present. Granitic pebbles are very occasionally found. The mean size ranges from a few cm to 10 cm. The largest clasts are 20 cm. The gravel fraction of the southern fringe lacks extra formational gravel components; however varying amounts of rounded caliche pebbles are present.

The sandstones consist mainly of detrital limestone clasts with an extraformational origin. Furthermore mostly small percentages of quartz and feldspar have been found. Overall the quartz percentages are either below 10% or above 15% (Fig. 7). Sandstones with a high quartz percentage contain also considerable amounts of alkali feldspar, and have a white weathering appearance.

Within the Monllobat and laterally equivalent Castigaleu Formations quartz percentages are generally low. Only the lower 60 m shows some scattered sandstone bodies with high percentages and there is a white sandstone present in the upper part of system VIII. The palaeocurrent directions from the latter, a 10 m thick lobe with foresets, indicate an origin from a system further to the west. White sandstone bodies with high quartz percentages are also present in several exposures of the southern fringe. High quartz percentages dominate the Castigaleu Formation below the base of the Monllobat Formation and also the Castisent Formation on top.

### IV.6.2. Variability of sedimentation induced by hinterland composition

The configurations and dimensions of the eight feeder canyons in the source rock area remain largely unknown. The length cannot have exceeded some tens of kilometres because of the emergence of the Eocene Pyrenees as suggested by Mattauer & Henry (1974). The composition of the sandstones provides more detailed information on the rock types in the hinterland. Two main rock types delivered the sediments; massive micritic limestone cover and the granitic core of the Central Pyrenean build-up (Mattauer & Henry, 1974).

The ratio of the two constituents of the sediments varies. The lower Castigaleu Formation, parts of the lower cycles of the Monllobat Formation and the western San Esteban Fan are arkosic, incorporating also large amounts of micritic limestone clasts. The major parts of the Monllobat and laterally equivalent Castigaleu Formation are dominated by limestone clasts. This indicates that the accessibility of the core zone must have varied, presumably because the

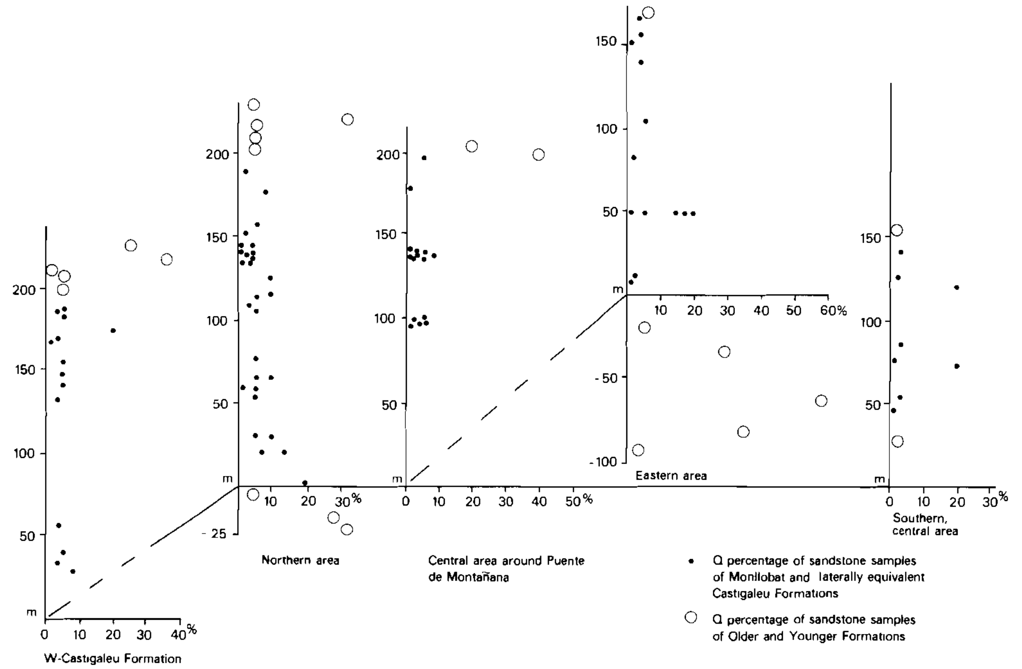


Figure 7. Quartz percentages of sandstone samples vary laterally as well as vertically over the Monllobat and Castigaleu Formations.

Mesozoic limestone zone blocked, to a variable degree, the supply of weathered granite. This blocking influence will have been promoted by the smaller susceptibility to weathering of the limestones relative to the granites.

Some of the variation of the systems of deposits can be linked with variable properties of the source rock area, laterally as well as in time. The arkosic sheet of system I at level 2 and all coarse member bodies of system VIII consist of sandstone with very minor amounts of gravel. A lower relief of the Mesozoic limestone zone, delivering only sand and only partly shielding the granitic zone, can have caused both features. There are a few arkosic incursions in system VIII with southeastern palaeocurrent directions. The balance between the Monllobat and the western Esteban systems must have been less fixed than between individual Monllobat systems.

Yellow mudstone is mainly present in the western system VIII, with an increasingly eastwards distribution upwards, and along the southern basin margin (Fig. 6). Scattered occurrences are present in between levels 1 and 2 of systems VII and VIII and at certain levels as already mentioned. These mudstones are accompanied by sandstones only (with relatively high as well as low quartz contents). Presumably in the lower Monllobat stages the Mesozoic limestone relief was not high enough to prevent fines from the weathered granitic area passing. During the upper stages the blockade was established, but with the tendency to be breached shifting eastwards from the hinterland of system VIII. Here the limestone zone is likely to have been lower as it bounded the San Esteban area with largely breached limestone terrain.

Yellow mudstones at the base of the upper levels of system V (in line with the eastwards shift of the distribution of this type) point to a temporary breaching of limestone relief at the onset of accelerations of tectonic activity. Yellow mudstones and the sandstones of the southern basin margins were derived from a southern hinterland with a composition comparable to the Pyrenees.

## IV.7. Tectonic features and the variability of palaeocurrent directions and coarse/fine member ratios

### IV.7.1. Distributions of fault and joint directions

The tectonic disturbance by faulting is of minor extent when the number of normal faults and their throws are taken into account: displacements are in the range of a few to 30 m. In nearly all cases these throws are achieved over several closely-spaced faults. Fault zones seem to be continuous (Fig. 2). However lateral tracing is hindered by the bad outcrop conditions on the flat, vegetated tops of ridges between the river valleys. Fault planes have generally high angles (70–85°).

Joint directions measured along the Ribagorzana valley and in the northwestern area are presented in Figure 4h, together with the directions of faults indicated in Figure 2. The fault and joint directions are near parallel except for the absence of southwesterly-directed faults. (Detailed measurements on the fault near Badias could not be obtained because of very bad outcrop conditions. The fault line has been traced on aerial photographs, see Fig. 2.)

Two features are conspicious, when Figures 4a–d and 4h are compared. Firstly the distributions of current directions and joint directions are largely overlapping and, secondly the fault surfaces of the upthrown blocks are mostly facing in an upstream direction. Reversals of the three components of the palaeocurrent pattern are found near faults. An overall scatter of transport directions (Fig. 4 f) appears at the southwesterly-directed fault at Badias, where dip directions vary strongly too.

### IV.7.2. Anomalous coarse/fine member ratios at fault zones

Coarse member lenses are very scarce in the fine member part of cycles except for the terminal parts of the systems. However at certain locations several lenses can be present in the succession in more proximal parts. Such locations, mostly accentuated by faults but occasionally also by synclines, lie kilometres apart (Fig. 2).

Overall five coarse member settings in tectonically disturbed regions can be distinguished:

(a) There are no variations of thickness, facies and palaeocurrent directions over the fault plane.

(b) Lenses fill longitudinal depressions along faults. Palaeocurrent directions are parallel to the fault. Meander lobes particularly are characteristic. An example occurs 30 m below a meander lobe on top of a sequence thickening towards a fault (described by van der Meulen, 1982). Both meander lobes have fault-parallel palaeocurrent trends. The margins of the lower lens are upturned (Fig. 9d).

(c) Wedges thin away from the fault. Palaeocurrent directions are perpendicular or parallel to the fault. Meandering and low-sinuosity channel fills and (lacustrine) fan-deltas occur at either side of the fault plane. Several examples are found at a fault zone consisting of two parallel faults (80–260° trend), located 500 m to the south of $A_2$ (Fig. 3). Several meander lobes are stacked just to the south of the fault at the base of the section. A wing emerging from one of the lobes can be traced over several kilometres to the south. A meander lobe with a northwards extending wing lies at the other side of the faults and at a higher stratigraphic position. Again somewhat higher in the sequence a thick lacustrine fan-delta has a westwards palaeocurrent trend along the faults.

The fault zone is also present on the other side of the Ribagorzana Valley (1 km to the south of $B_3$), where multicoloured mudstones and several sandstone layers are found. Compared with the surrounding Monllobat sections the following combination of additional characteristics is conspicious: the mudstone has a relatively dark brown colour and there is no regular intercalation of caliche layers. Only a minor amount of conglomerates is present. Nevertheless there are no blue-grey mudstones and oyster bearing sandstones of the Castigaleu Formation. The abundant tabular crossbedding in the sandstones is in several cases bi-directional. Six of the palaeocurrent readings are perpendicular to the fault trend; several of the other readings are parallel or slightly oblique (Fig. 4g).

Lobate build-ups of conglomerate are found at

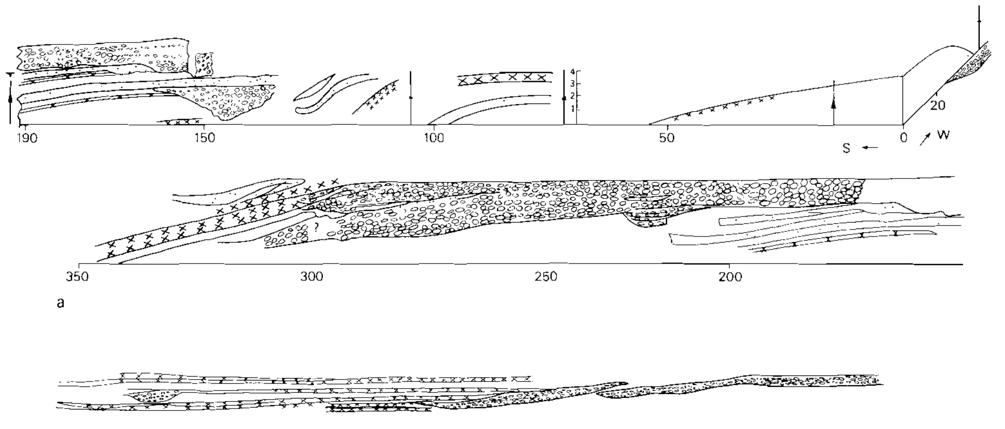


Figure 8. Two transitions of conglomerates to mottled siltstone and mudstone in the palaeocurrent directions. The location is close to $B_2$ in Figure 3.

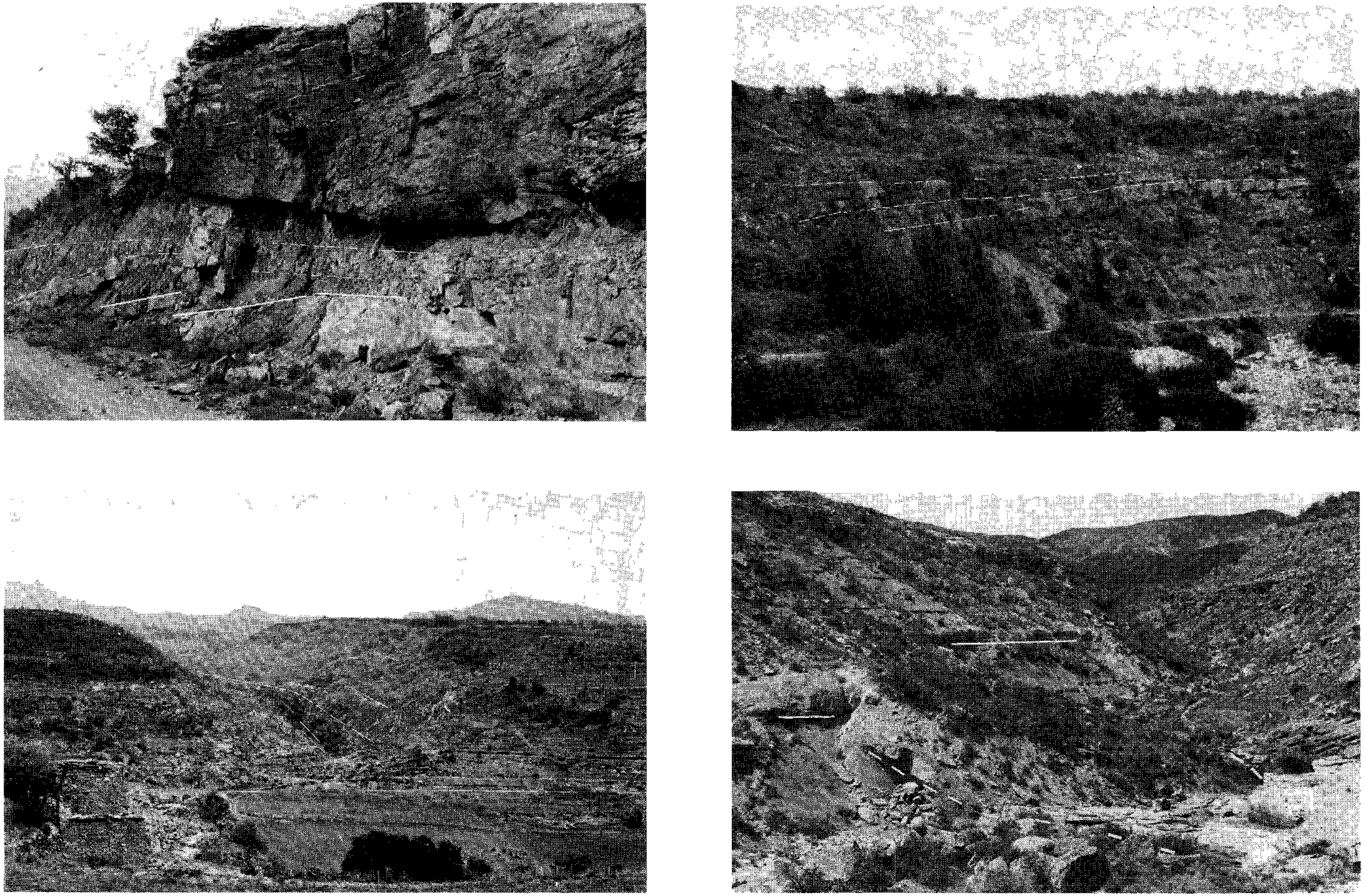


Figure 9. (*a*) Large-scale foresets on top of dipping, mottled mudstone layers with flat, upper ends.

(*b*) Margin of the syncline with high sand/shale ratios. The thickening consists of foresets. The dip of the flat, upper layers increases also in the syncline (compare Fig. 9d). Location at km 10 of the $A_1A_5$ section of Figure 3.

(*c*) Thickening at the syncline margin. Location at 0.5 km of the $B_4W$ section of Figure 3.

(*d*) Upturned margin of a meander lobe. A fault is situated near the other margin. The upturned margin is linked with a caliche layer, which is also present in the upthrown block at the background. Further in the background the same thickening as in Figure 9b is present.

either side of a 140–320° trending fault with a northeastern downthrow of a few metres. The exposure of Figure 8a is situated at the southwestern side (fault location: 300 m to the north of $B_2$ in Fig. 3). The wedging conglomerate sheet of Figure 8b is exposed at a higher stratigraphic level on the other, northeastern side of the fault zone. In both cases conglomerates interfinger with mottled mudstone and siltstone with caliche. The shape of the beds is highly irregular. The transition from conglomerates to the fine member takes place in the palaeocurrent direction, which is perpendicular to the fault trend.

(d) Facies variations occur at either side of the fault plane over relatively large areas. This type of setting is found just to the north of Monllobat on a westwards-directed fault (locations in Figs. 1 and 2). The two sheets of system I (levels 2 and 3) terminate in the southern uplifted block close to (level 2) or at the fault (level 3). Conspiciously the thickness of the upper one triples at the fault before wedging occurs. From the northeast, sandstone layers thin and wedge towards the fault. Blue-grey mudstone and fan-deltas only occur in the downthrown fault block.

(e) Shallow synclines have high sand/shale ratios. The most prominent example stretches in a northwestern direction from Castisent to Montañana (black zone in Fig. 2, briefly described by van der Meulen, 1982). Thickening and downwards bending of coarse member layers is given at three locations in Figure 3 (at 0.5 km in the $B_4W$ section, 11 km in the $B_1B_5$ section and 10 km in the $A_1A_4$ section). This feature is especially prominent at level 3a. Figure 9b shows the rapid thickening of a conglomeratic sandstone layer at the location in section $A_1A_4$ (km 10). Large foreset structures are found in the thick part. The dip of the layers higher in the section increases also. Figure 9c presents a detail of the thickening from the location in the $B_4W$ section (at 0.5 km). Meander lobes of the lower level 3 are exposed to the east. The palaeocurrent directions are along the synclinal axis. The meander lobes form a type (c) setting outside, but directly adjacent to the syncline.

Generally more than one type of coarse member setting is found at faults. Type (a) and (c) settings are commonest, but (b) and (d) types may also be present. Lack of variation may alternate with several types of variation in successions close to faults. It should be noted that wedges (type c) can occur alternately on the two sides of the fault plane.

### IV.7.3. Reconstruction of synsedimentary faulting

Faults dissect the layers without deformation of internal structures. Wedges and lenses do not lie against the fault planes but are truncated at the margins. The upturned margins of the type (b) example are sheared without fault contact. It is therefore concluded that fault propagation occurred after sediment consolidation and after sagging of the lenses.

Sanford (1959) simulated deep-seated fault propagation starting with elastic deformation and terminating with faulting. In the case of the Monllobat deposits initial elastic deformation will have affected the layers at the surface forming longitudinal depressions and flexures. These sites were then preferentially filled by gravel and sand, constituting type (b) and (c) settings respectively. After burial and consolidation, faults propagated upwards through the deposits.

In the case of the example of a type (d) setting there are clear indications of relative motions at the surface of large areas. Fan-deltas were developed in the downthrown block. However coarse member sheets were at times formed at the other block which then lay lowest. As in the other cases fault block motions seem to have alternated in sense of movement. The westwards-trending fault in this example constitutes the boundary between systems I and II. Discontinuity of the fault can explain why the southwestern system course became fault-parallel at this location only. On a much larger scale the transition from a fault to an anticline takes place in the western area, indicating a transition from extension to compression. The anticline constitutes the boundary between systems VII and VIII. Nijman (1981) interpreted a relation between the fault and a large-scale nappe rotation.

The wide, shallow synclines of type (e) probably developed because of secondary faulting at the buried margin of a fault block. Abundant fan-deltas originated along the margins of the wide synclines. Restriction of this type of faulting to part of the block margin would explain the anomolous features 1 km to the south of $B_3$. This synclinal depression seems to have been protected from the main coarse and fine sheetfloods leading to a lack of gravel in the coarse member and silt in the fine member. Silt-bound caliche is also lacking. Tabular crossbedded fan-deltas were formed by currents along, and perpendicular to, the synclinal axis. Possibly there was also a large-scale type (d) effect to the east of this location (at 500 m to the south of $A_2$, where type (a), (b) and (c) settings occur near the fault). The two sheets which cross the western boundaries of system IV and V lie in the downthrown block to the north of the fault. There are also several lacustrine intervals in this block to the north of this fault and therefore temporary westward deflection of sheetflood systems in this area seems likely.

In general the fault block motions have not influenced the sediment distribution on the scale of systems. The similarity of tectonic and transport directions is probably caused by some tilting of the fault blocks. Nevertheless most sheets cross fault zones without major variations, although later fault displacements suggest obstruction. Two reasons can be given for this. Firstly tilting must have been a gradual process with major gradient deviations only at faults.

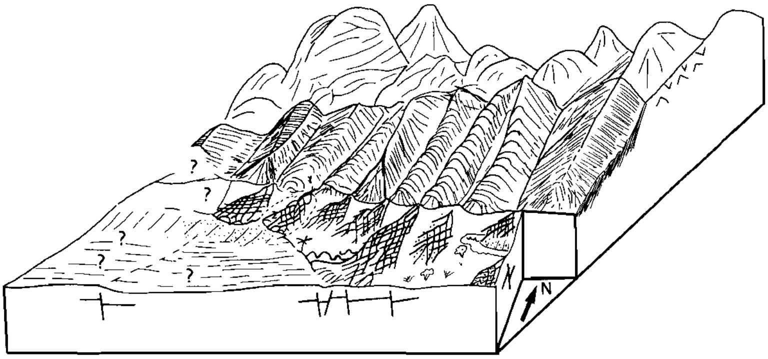


Figure 10. A reconstruction of the depositional environments of the Monllobat and Castigaleu Formations in front of the Pyrenees. Two main constituents of the Pyrenees delivered the sediments; the granitic core and the limestone cover. Deep-seated faulting influenced sedimentation of the sheetflood systems.

Secondly coarse member effects indicate non-uniform and reversible movements. In some cases a link between synsedimentary and later fault deformation is apparent; in other cases not.

### IV.7.4. Relation of synsedimentary faulting to regional tectonics

The three components of the fault/joint diagram in Figure 4h are also the main directions of faulting in a large-scale fault block model for the Pyrenees (Peybernès & Souquet, 1984). The faults in this model bound large Pyrenean blocks, developed and deformed in the process of oblique convergence of the European and Iberian plates in Upper Cretaceous and Palaeogene times. Fault block deformation was accompanied by cover detachment, however low-angle thrust planes were not found in the area studied. The fixed positions of the alluvial systems rule out major relative horizontal movements of fault blocks within the study area, except perhaps in a southwesterly direction.

## IV.8. Conclusions

Figure 10 illustrates the three major controlling features of the environment during the accumulation of the Monllobat and Castigaleu Formations.

(1) The Pyrenean mountain chain, which supplied most of the sediment, consisted of a granitic core with a limestone cover. The latter obstructed to a varying degree the discharge of weathered material from the granitic central area.

(2) Eight sheetflood systems have been distinguished. Their positions were fixed through Monllobat time. Deposits grade from gravel to sand downstream and there was a northwestwards, downstream deflection of current directions. Silt and mud were deposited around the margins of the coarse sediments. Some of the sheetfloods were depositing entirely on land, others discharged also into the sea.

(3) Synsedimentary movements of large fault blocks influenced sedimentation particularly by causing depressions at the surface. Additional features are:

(4) Seven cycles were formed.

(5) A transgressive mudstone layer marks each cycle base, which was then covered by extensive sheets of conglomerate and other coarse member layers. The much thicker upper part consists largely of a fine member. This sequence was due to an acceleration of tectonic processes during the formation of the basal cycle deposits.

(6) The cycles are relatively thickest in the middle part of the Monllobat Formation. A regressive trend changes to a transgressive one in the upper parts.

(7) The northeastern sheetflood systems competed with systems with a southern origin. These southern systems supplied relatively minor amounts of sediments, although this increased towards the top of the Monllobat Formation.

chapter V

# Eocene sheetflood systems and transitional fan-deltas, Southern Pyrenees, Spain

## V.1. Introduction

The Monllobat and Castigaleu Formations constitute the lower part of the Montañana Group, part of the Eocene fill of the elongate Tremp–Graus Basin in the Southern Pyrenees. The Castigaleu Formation lies partly below the base of the Monllobat Formation and is also partly its lateral equivalent (Figure 1). The continental Monllobat Formation is largely situated to the east and the laterally equivalent, marine Castigaleu Formation to the west of the present Ribagorzana Valley (Figure 1).

The Monllobat Formation contains northeasterly- and southerly-derived sediments (Nijman 1981). The former comprise the major part of the formation (Van der Meulen in press). Kilometres-wide conglomerate sheets, intercalated in and connected with large amounts of fines, constitute the bulk of the northeasterly-derived coarse members. Sheetfloods are thought to have deposited these sediments (Van der Meulen in press). Southerly-derived sediments are found in a narrow fringe along the southern basin margin. A mudstone-rich zone (to the east) and the marine deposits (to the west) separate the northern alluvial area from the southern area.

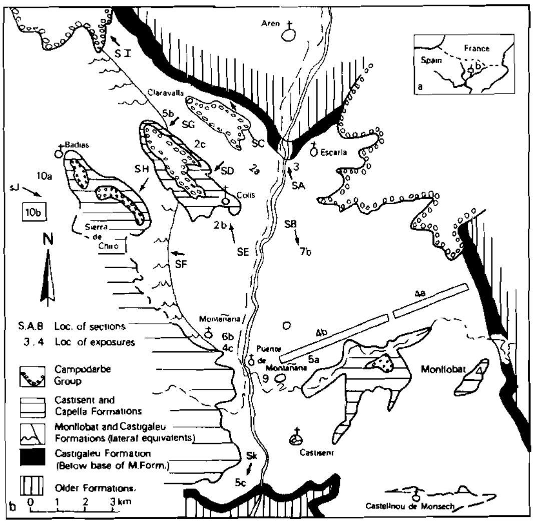


Figure 1. Geological map of the study area. Inset shows geographical location. The locations of the Sections A–K and exposures shown in Figures 2–10 are indicated on the geological map.

This study succeeds general sedimentological studies in the area (Van Eden 1970; Nijman and Nio 1975; Nijman 1981; Van der Meulen in press) and detailed investigations of some exposures (Puigdefabregas and Van Vliet 1978; Van der Meulen 1982, 1983; Leo and Allen 1984).

The purpose of this paper is to give the detailed sedimentology of the formations in the northwestern part of the study area (Figure 1). There, Monllobat sheetflood systems discharged perpendicular to the coastline into the

Castigaleu lagoon. The description is supplemented with illustrations of important, large exposures from the eastern area.

## V.2. The build-up of the Monllobat and Castigaleu Formations

Both formations have very thick fine members (mud and/or siltstone). The Monllobat fine member is multicoloured owing to ancient soil processes (Nijman and Nio 1975). These continental deposits contain 20–30 per cent conglomerate and sandstone. Minor amounts of oncoids occur. Castigaleu mudstones are blue–grey with 20–30 per cent sandstone distributed throughout. Oyster remains are common in the sandstones.

The northeasterly-derived conglomerate sheets overlie thin transgressive mudstones at the bases of seven Monllobat intervals (180 m thick in total). This points to rapid progradation of alluvial systems following transgressions, linked to accelerated basin subsidence/hinterland uplift (Van der Meulen 1983, in press).

The Monllobat deposits can be divided into eight juxtaposed compartments each of which can be identified by the position of its conglomerate sheets. These sheets at the bases of the intervals do not overlap with those of adjacent compartments but there is complete overlap of sheets in the same compartment.

Distally, conglomerate sheets wedge or split resulting in a number of thin, lenticular coarse members which are more evenly distributed throughout the sections. This trend continues into the Castigaleu Formation. The southern basin fringe consists of coarse member lenses and a few sheets distributed rather regularly throughout the sections.

Concentrations of coarse members throughout the vertical sections are found adjacent to some faults, which generally have a throw of up to 30 m. Lenses and wedges occur adjacent to faults and in some cases there are facies differences across the fault plane. These features were interpreted as having originated during deep-seated faulting (Van der Meulen in press).

The conglomerates and sandstones are dominantly composed of micritic limestone clasts. Minor amounts of vein quartz and rare granitic pebbles occur. Small percentages of quartz and alkali feldspar grains are incorporated in most sandstones. Very rarely the sandstones are arkosic.

## V.3. Description of the Monllobat Formation

### V.3.1. Fine member

Decimetre- to metre-scale flat layering is the most common structure in this member. Most of the Monllobat fine member is brown- or ochre-coloured with blue and/or red vertical mottle traces. There are no well-developed soil profiles within layers. Nevertheless non-mottled brown layers always separate blue-mottled brown and ochre layers from underlying blue-grey layers. Red mottles can replace blue ones higher up in the sequences. Generally the same basic colour (brown or ochre) is found throughout thick sections with occasional changes from one type to the other.

Over thick sequences, mottles increase in number upwards from non-mottled sediments. Mottling may gradually disappear in vertical transitions towards non-mottled fines. However, intense blue and red mottling is occasionally found underneath levels with transgressive mudstones and conglomerate sheets. Muddy channel fills are also intensely mottled.

Four types of caliche have been distinguished in the fine member on the basis of the size and the amount of carbonate concretions (*cf.* Reeves 1976).

1. A few small (diameter up to 0·5 cm), spherical or ellipsoidal concretions.
2. Numerous rather irregular concretions (diameter 0·5–1 cm).
3. Many nodular concretions (diameter 1–3 cm).
4. Large, closely packed nodules (diameter 3–10 cm) with clay between.

The type of concretion is generally the same within layers and only in a few layers is there a vertical sequence containing several types. Caliche-rich (types 3 and 4) and cemented siltstone layers stand out in regular alternations with mudstone often containing type 1 or 2 caliche (Figure 2a). The thickness distribution of caliche-mudstone couples shows gradual oscillatory patterns of increase and decrease. Coarse members can be present in the couples instead of caliche beds. Lateral transitions are also found where nodular caliche passes first into completely indurated sediment and then in coarser-grained sediment. Such a transition takes place from the left of Figure 2b towards the approximately 1 m thick channel fill. Winged, symmetrical channel fills of a larger size (2–5 m thick) may possess highly erosive bases incised into the mudstone.

The flat layers are almost always homogenized and mottled and therefore devoid of sedimentary structures. However, an exposure on top of the Castigaleu sandstones has an unaltered basal part with parallel lamination passing laterally into convolute (Figure 3b) and climbing ripple lamination. Mottling of the upper parts starts with blue-coloured horizontal burrows, agreeing with *Steinichnus* as described by Bromley and Asgaard (1979). The upper parts with vertical mottle traces show a typical weathering profile with protruding siltstone and excavated mudstone layers (Figure 3a).

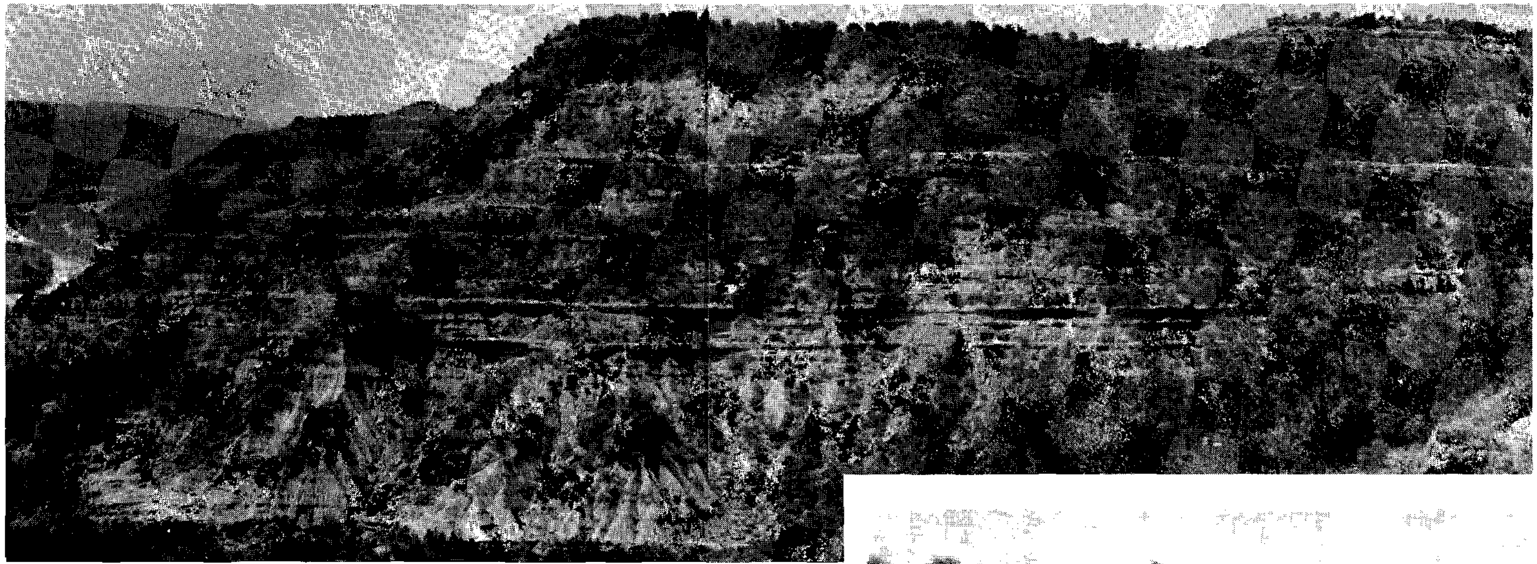
Yellow-coloured fine members occur in some intervals at the northwestern basin margin and along the southern margin. Mudstone with a metre-scale flat bedding dominates these fine members. There are blue and/or blue–grey vertical mottle traces and minor occurrences of gypsum caliche (types 1 and 2).

### V.3.2. Coarse member

*Facies 1—proximal conglomerate sheets.* These sheets are hundreds of metres to kilometres wide with thicknesses averaging 2–5 m and a maximum of 10 m for the central, thickest parts of some sheets. The bases are gently concave-upwards and the tops are flat. The conglomerates are closely packed, sometimes with coarse-grained sand in the matrix and as a decimetre thick upper layer. The pebbles are seldom well imbricated, except for a few strings of flat pebbles.

Individual channels can seldom be recognized within sheets (Figure 4a,b). Shallow erosion surfaces, combined with clast size differences (logs B–D, Figure 4a), which might indicate channels, are discontinuous. Only narrow, marginal channel fills with concordant conglomerate–sandstone interbedding are well established (Figure 4a, to the east, and log A).

Two aspects of logs A–E of Figure 4a are conspicuous. Logs C and D are first coarsening and then fining upwards. Palaeocurrent directions change from southwesterly in A, B and C to westerly in D. Log E shows an upwards trend from westerly to northwesterly.



(a)

(b)

Figure 2. (a) A thick sequence of mottled mudstone with caliche horizons constitute the valley wall at the foreground. Several small channel fills are present at caliche horizons. (b) Two small channel fills with wings of mottled material. The wings fine laterally away from the fill from sand to mud with caliche.

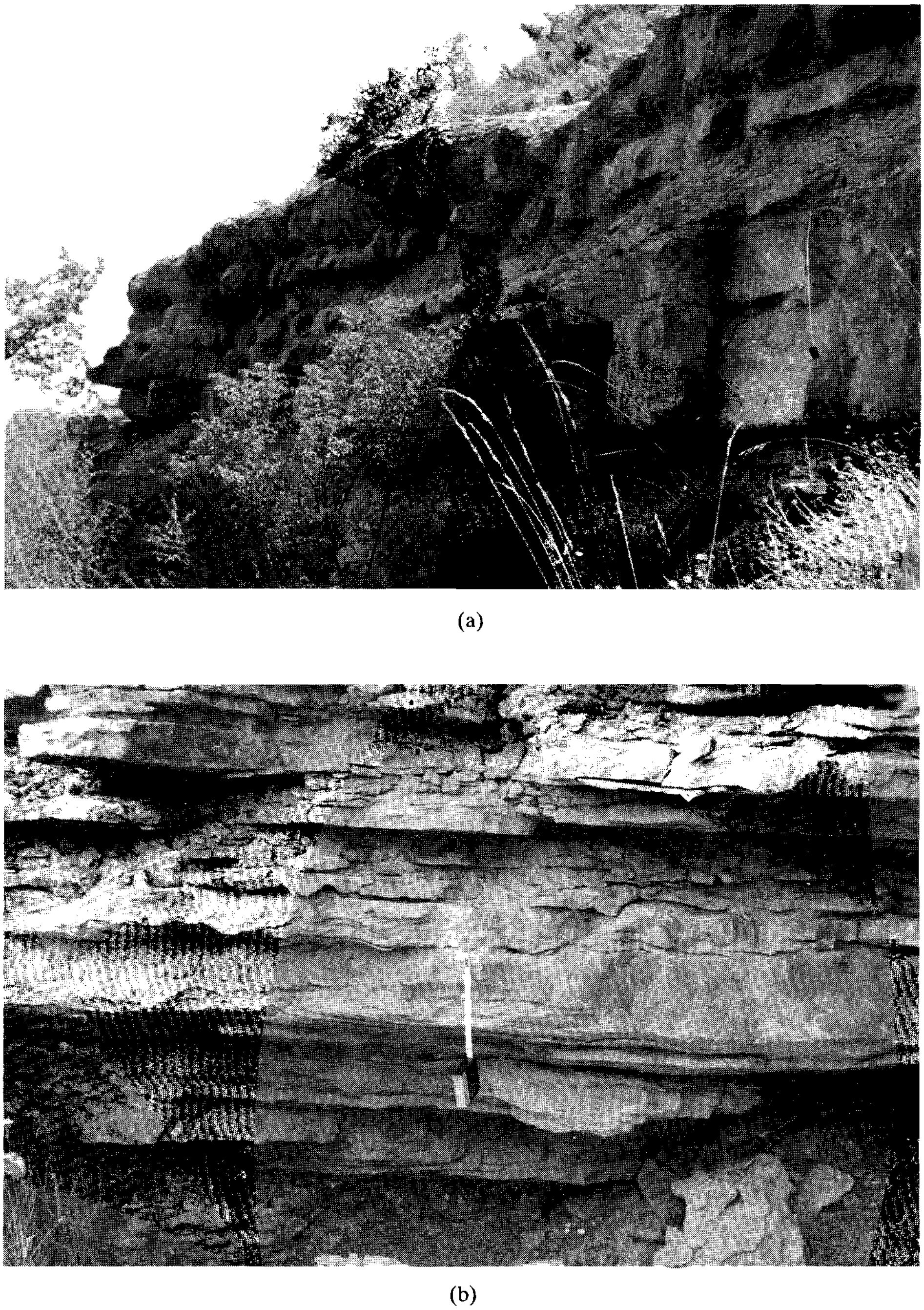


Figure 3. (a) The characteristic appearance of mottled layers in the field. The scale is 50 cm. (b) Non-mottled flat beds underneath the layers of Figure 3a. The sand beds have convolute laminations. The scale is 10 cm. Locations in Figure 1.

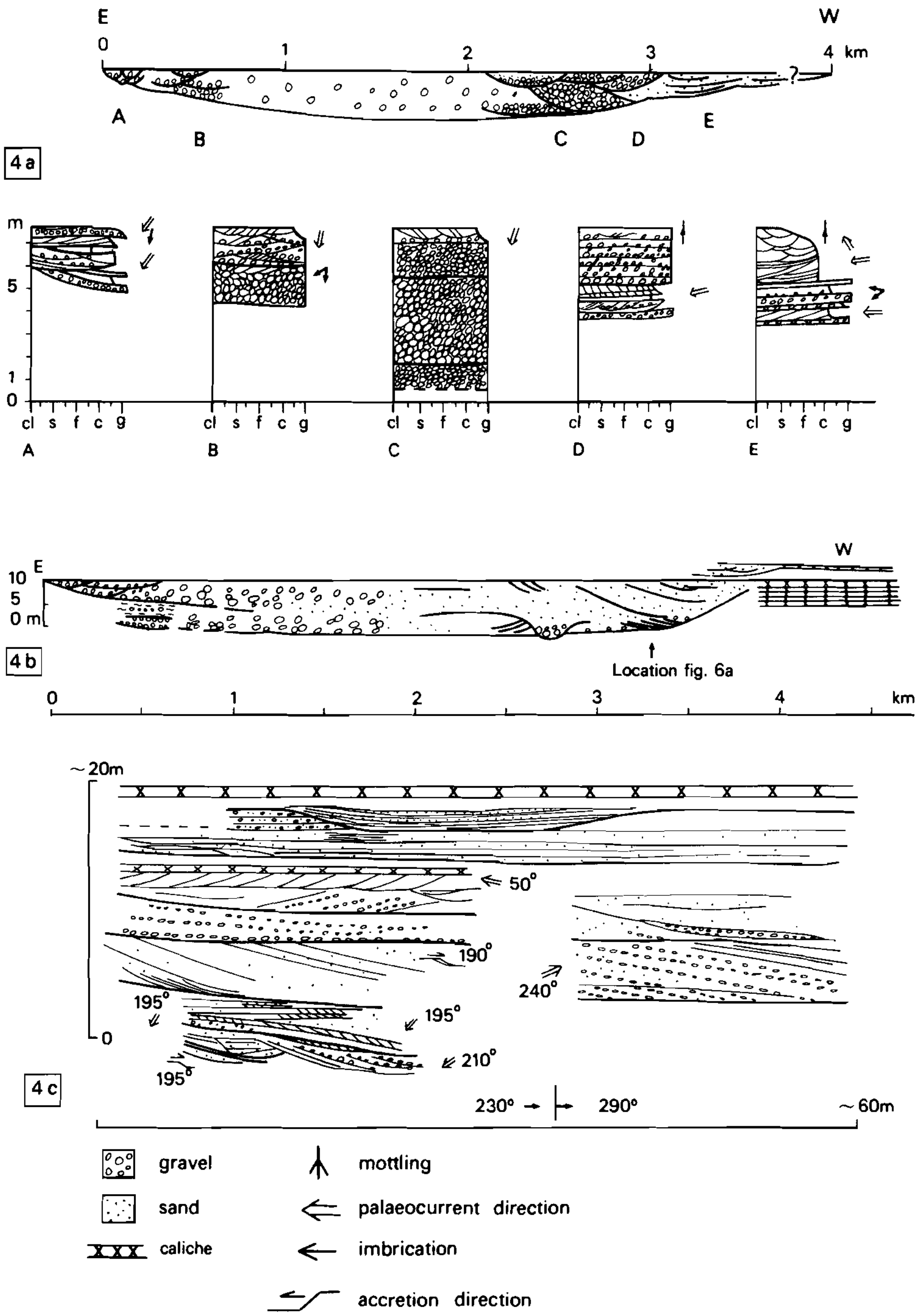


Figure 4. (a) A section of the distal parts of a kilometres wide gravel sheet. The section is oblique to the southwesterly palaeocurrent direction of the conglomeratic part. The trends of grain-sizes and palaeocurrent directions are variable. (b) Mottled mudstone interfingers with conglomerate beds of the eastern sheet part. Large foreset structures dominate the western part. The scale is the same as in Figure 4a. (c) A section showing thickening of a 3–4 m thick gravel layer. Locations in Figure 1.

*Facies 2—distal conglomerate and sandstone sheets.* Channel fills are well established in the terminal sheet parts where gravel grades to sand and palaeocurrents change from southerwesterly to westerly and northwesterly. Figure 4a shows this facies with shallow channelling in the western parts of sheets. However, several terminal sheet margins lie on top of transgressive blue–grey mudstone at the base of Monllobat intervals (western parts of Figure 4b, Figure 4c). Sheet thickness is maintained towards the distal margins. However, major thickenings can appear at sites where synsedimentary tectonic activity occurred (Figure 4c and Van der Meulen in press).

*Facies* 3a, b—*symmetrical channel fills.* These have widths of tens of metres and a maximum thickness of 5 m. The erosive bases of these channels (Figure 2c) are concave upwards and the tops are flat. The fills consist of massive conglomerates with possibly some cross-bedded (conglomeratic) sandstone. The flat tops of very coarse-grained sandstone pass laterally into siltstone wings with abundant caliche.

Small channel fills are generally intercalated in multiple siltstone beds. Thicker channel fills are usually associated with only one pair of wings, surrounded by mudstone. Therefore corresponding facies types 3a and 3b are distinguished respectively. The former may consist of sandstone as well as conglomerate. Channel fills lie mostly downstream from sheets, but in some cases a series of fills, linked by their wings lies in an upstream position.

*Facies* 4a, b, c—*foreset lobes.* The thicknesses of foreset lobes are equal to the ones of upstream sheets or channel fills. Single lobes have widths of tens of metres. Low-angled (4a) and high-angled (4b) conglomerate lobes consist of foresets dipping up to 15° and 15–30° respectively. Arcuate foreset plans can often be observed on top surfaces. In some cases the unit is overlain by a flat conglomerate layer or another facies type. Figure 5a shows how the base can be erosive and low-angled foresets can pass into high-angled ones. At the top, other facies types (for example 4c, a sand-siltstone lobe) can be present. Foresets consisting entirely of sandstone are not known from the northwestern Monllobat Formation but occur in the eastern parts and very rarely at the southern basin margin.

Lobes built of mottled sand–siltstone foresets (facies 4c) have a very well-established convex-upwards shape in transverse section (Figure 5b). The layering stands out both in cross-sections (Figure 5b) and in top surfaces (Figure 5a). Mudstone can be interbedded (Figure 5a). Individual beds grade from coarse- to fine-grained sand and silt (Figure 5b).

These foreset lobes are found underneath or, more commonly, on top of other layers (Figures 5a and 5b). Figure 5c illustrates the transition of a high-angled lobe of rounded caliche pebbles and sandstone to high-angled sand–siltstone layers at the western margin. The change of trend from proximal northern to distal western foreset directions illustrates the general current pattern at the southern basin margin.

Dispersed directions from Monllobat conglomerate lobes (Figure 11a) weaken the prevailing southwestern trend which is generally found in the sheet conglomerates and symmetrical channel fills. Sand–siltstone foreset lobes contain evidence of southwesterly directions (Figure 11b) adding a third component to the general western and northwestern current trends of the sandstone facies.

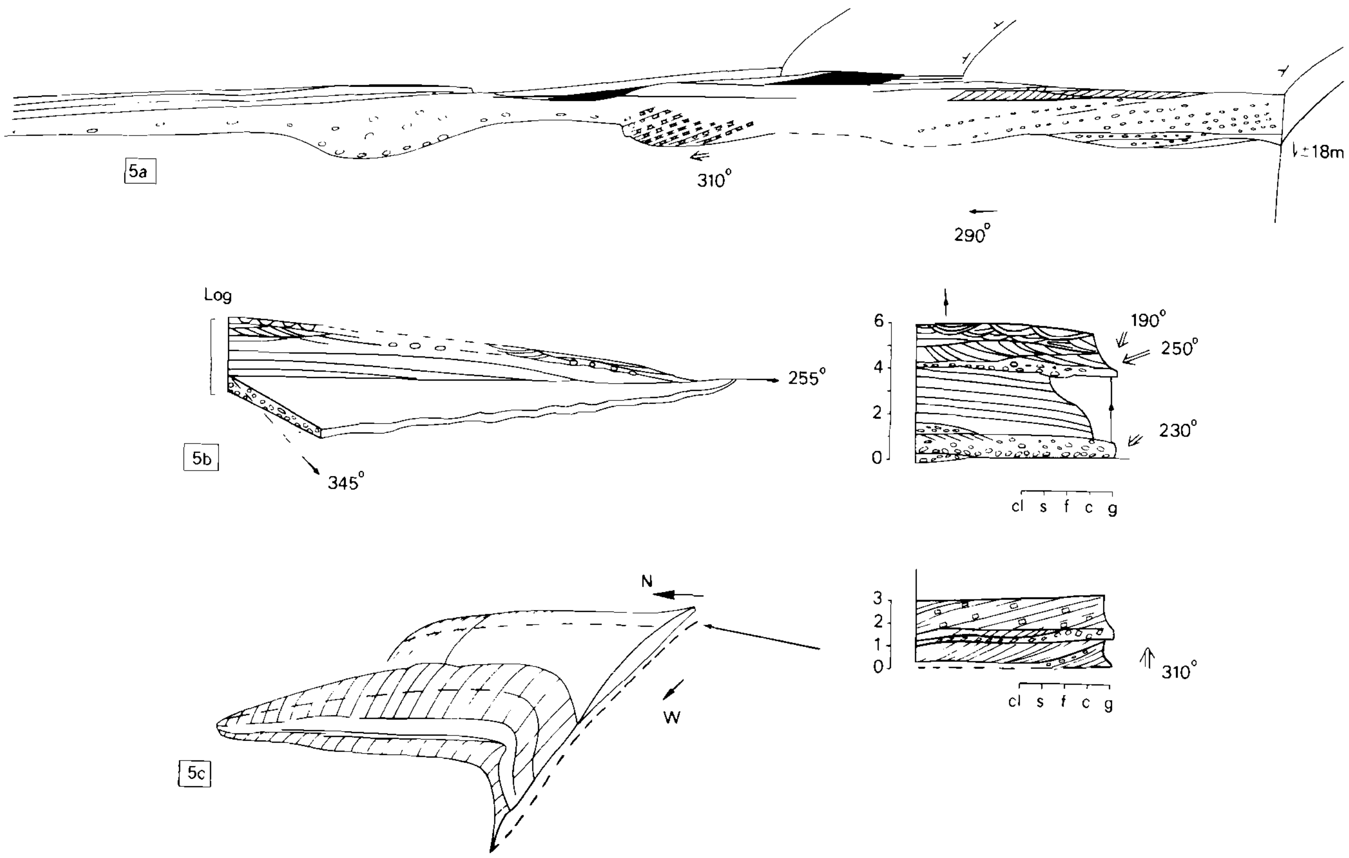


Figure 5. Three examples of deposits with a lobate build up, accreted in the palaeocurrent direction. (a) Low-angled foresets pass into high-angled ones in the lower parts. Lobate, low-angled foresets at the top consist of sand fining upwards to silt. (b) Sand–silt lobe with prominent convex-upwards geometry shows a good fining-upwards sequence. (c) An irregularly-built lobe from the southern area. The proximal southern part (log) with high-angled foresets passes in a low-angled, lobate foreset and some high-angled beds at the western termination. Locations in Figure 1.

*Facies 5—lateral accretion lobes.* The widths of these lobes range from a hundred to several hundreds of metres with thicknesses of 2–10 m. The dominant lithofacies consists of a relatively thick upper part of inclined mottled sand–siltstone layers on top of a thin, trough cross-bedded sandstone (Puigdefabregas and Van Vliet 1978; Van der Meulen 1982).

The upper lithofacies of inclined beds with arcuate geometry resembles facies type 4c. The beds are also completely homogenized and mottled. However during three-dimensional analysis of facies type 4c deposits (Van der Meulen 1983, Figures 3 and 14) three major differences with the earlier developed model for laterally accreted deposits (Van der Meulen 1982) were found. These differences are the lack of a marginal channel fill, the fact that the geometry is related to substratum relief and the setting of the type 4c deposists on top of other facies types instead of on the basal lithofacies of type 5.

The spatial model for facies type 5 was developed on the basis of dimensions, lithofacies and palaeocurrent directions (Van der Meulen 1982). Two sections of lateral accretion deposits illustrate most of the features of this model. Figure 6b shows the relative positions, some lithofacies details of the upper one having been previously described by Van der Meulen (1982). The sections are equally thick, but the upper one (Figure 7a) is much wider than the lower one (Figure 6c). In both cases there is a flat-topped lower point bar platform underneath a heavily mottled mudstone channel fill (to the west). However the wider section (Figure 7a) consists largely of the dominant lateral accretion facies, connected with relatively coarse initial deposits at the eastern margin, while the other (Figure 6c) incorporates massive, non-bedded mudstone in the upper lateral accretion interval. Palaeocurrent directions from cross beds are largely oblique to the

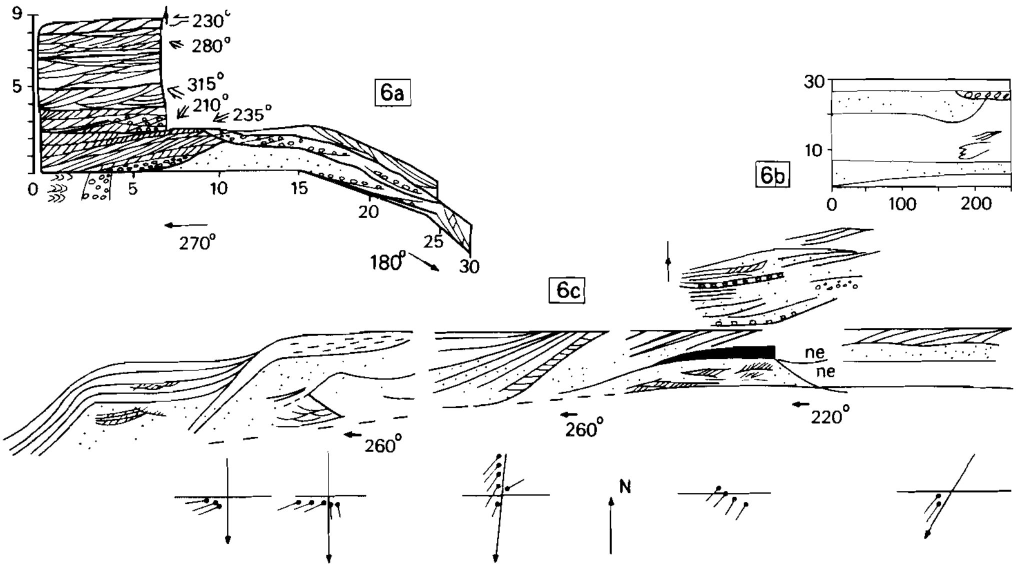


Figure 6. (a) Detail of Figure 4b. The geometry, upwards fining grain-size sequence and the type of sedimentary structures occur in lateral accretion as well as foreset deposits. (b) The setting and measures of lateral accretion deposits at the entrance of the tributary valley to Montañana. Location in Figure 1. (c) A detailed representation of the lower half of Figure 6b. Cross-bedding directions (figure base) are most obliquely to the strike (arrow) of the upper point bar layers.

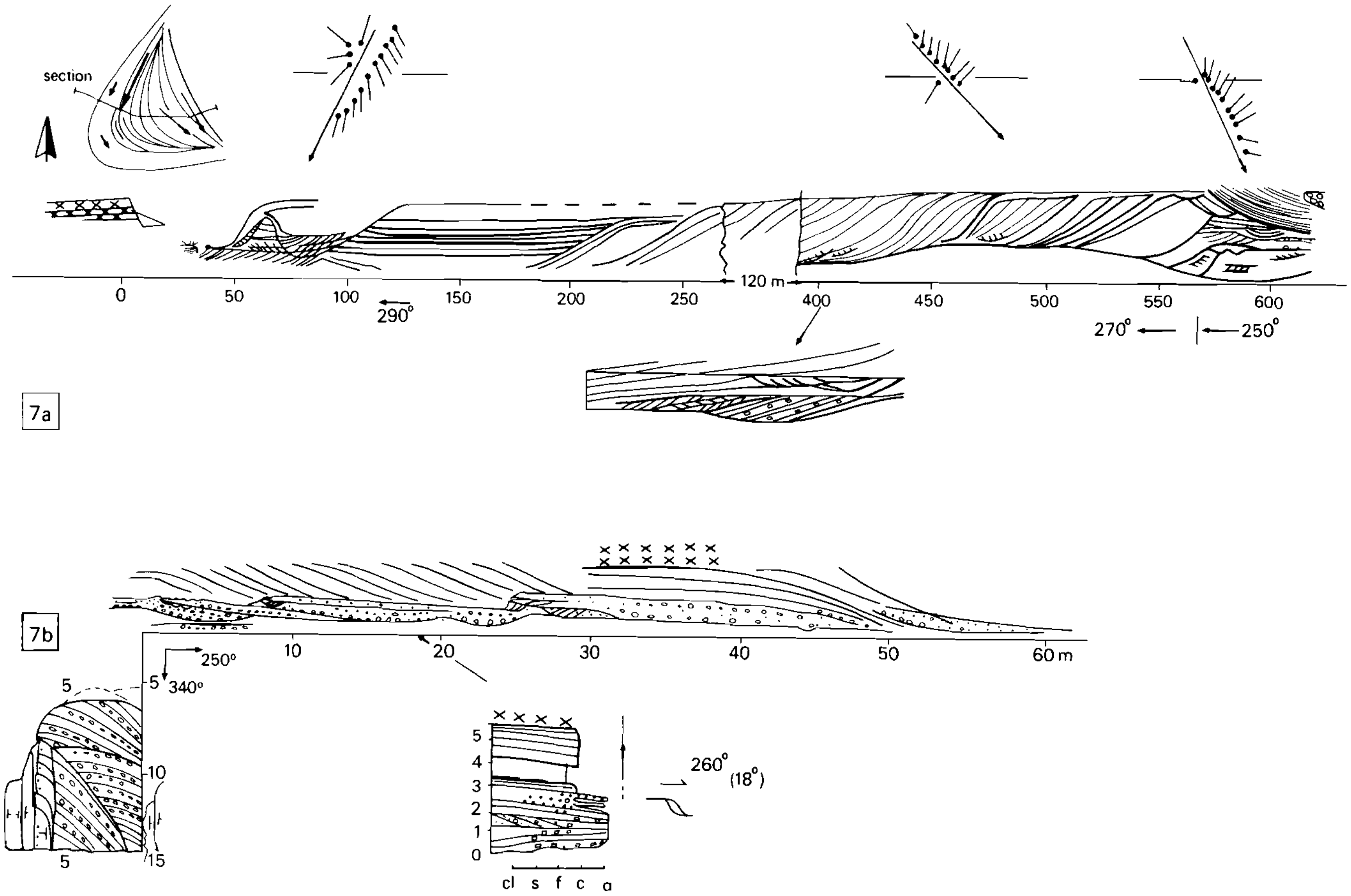


Figure 7. (a) An overall view of the upper layer of Figure 6b. The thickening at 400 m comprises lateral accretion beds traceable all through the layer. At the other side of the valley, the lower part of the thickening is made up of cross-bedded foresets in the accretion direction (see inset at base). Non-cross-bedded foresets are the internal structures of a platform underneath the marginal channel fill to the left. Cross-bedding directions and the strike (arrows) of upper point bar deposits are given at the top—together with the position of the section line in the reconstructed meander lobe plan. (b) These point bar layers are mostly relatively coarse as compared with point bar deposits of Figure 7a (see log). The plan of the inset to the left shows large trough-shaped structures and a recently weathered tabular set (to the east). Location in Figure 1

accretionary strike in a sense towards the cutbank in Figure 6c and towards the point bar in Figure 7a. Foresets at the flat-topped platform margin and incipient scroll bars strengthen these two trends.

The model does not incorporate the incision of the terminal point bars as in Figure 7a, the cross bedding of flat-topped platforms in Figure 6c and thickening of the central parts of the lobe as in Figure 7a (at 400 m). Downcurrent, foresets make up the thickening (see inset). The section of Figure 7b resembles the deposits in Figure 7a, but the grain-size is much coarser. The inclined upper beds consist of poorly-sorted, coarse-grained sand; mudstone interbeds are lacking. The relatively large trough cross beds at the base are directed along the accretionary strike with a gradual change towards the upper point bar (see left hand plan) with ultimately a straight-crested scroll bar in a scour (Figure 8).

The exposure shown in Figure 7b, which has a northwesterly palaeocurrent direction trend, lies directly adjacent to a northwesterly trending fault. The exposures shown in Figure 6b also lie along a northwesterly trending structure, but have a southeasterly palaeocurrent trend.

Sandy lobes with arcuate top structures from sheet margins resemble lateral accretion deposits. Details from Figure 4b show four other features known from lateral accretion deposits. There are good fining-upwards sequences, well-developed accretion planes with inclined, mottled beds at the top, and the accretion direction of the latter beds is perpendicular to the trend of the cross bedding (Figure 6a). However, close to this exposure the build-up is irregular, with several extensive erosion surfaces and with first coarsening- and then fining-upwards sequences. Furthermore, a marginal channel fill is lacking, there is

Figure 8. Detail of Figure 7b. The tabular set is an incipient scroll bar, situated in a scour close to the upper point bar base.

a basal, blue–grey mudstone interval and the palaeocurrent directions express the overall pattern of sheets as previously mentioned (see also Figure 4a).

*Facies 6—structureless flat beds.* These beds consist of sandstone, occasionally with small amounts of pebbles. Beds are extensive with thicknesses up to a few decimetres. These sandstones are mottled but otherwise completely homogeneous.

V.3.2.1.*Oncoid occurrences.* The Monllobat Formation contains a few exposures with oncoids, occasionally in large numbers. There is some relation between facies types and oncoid morphology. Coated pebbles have been found in symmetrical channel fills and in a channel lag from a meandering river deposit. Small, densely laminated oncoids with a single nucleus are known from the top of symmetrical channel fills and from the abandoned marginal palaeochannel at a meander lobe. This channel also contains a large cushion-shaped oncoid with fenestrate structures at the lower point bar. Discoid oncoids with several nuclei are abundant in a blue–grey mudstone interval enveloped by mottled layers. Large attached oncoids with fenestrate fabrics lie in a slightly mottled mudstone matrix, adjacent to the mottled top of a Castigaleu sandstone lobe and of a type 4c lobe.

## V.4. Interpretation of the Monllobat Formation

### V.4.1. Fine member

The colour pattern and caliche intercalations resemble the characteristics of pseudogley-type soils (*cf.* Buurman 1980). The brown, ochre and yellow colours are a consequence of oxidation during subaerial exposure. The distinction between the brown and ochre and the yellow suite was ascribed to differing source rock compositions (Van der Meulen in press).

The mottles consist of multicoloured iron compounds, developed along vertical rootlets during periods of water saturation of the soils. The mottle traces are generally, of metre-length, indicating rootlets reaching down to the groundwater table (*cf.* Cohen 1982), which consequently must have been at low levels for considerable periods of time.

The ochre-coloured and blue-mottled layers succeed blue-grey mudstone only after intercalation of brown mudstone, implying relatively lower groundwater tables during formation of ochre colours and mottles. Red mottling appearing again higher in such successions then indicates the relatively driest conditions. Thick ochre and brown successions with moderate mottling alternate, which can be linked to rather constant average positions of the groundwater table over long periods of time. Very occasional jumps to a different groundwater level may then have caused brown to ochre transitions and vice versa. Intense red and blue mottling laterally from transgressive mudstones at the base of coarse member sheets points to relatively long periods of soil formation. This can be the consequence of relatively low sedimentation and high subsidence rates, preceding high sedimentation rates. The intense mottling of muddy channel fills is a consequence of the great sensitivity to water saturation of this subenvironment.

The minor occurrence of caliche in the brown, ochre and yellow mudstone is explained by the low permeability. In contrast, the permeable silt beds received large amounts of water and dissolved carbonate from upstream gravel and sand layers because of groundwater discharge during maximum run off (*cf.* the process in the proglacial environment, Boothroyd and Nummedal 1978) and after precipitation rates too small to cause run off at the surface (*cf.* McGee 1897). The porosity determined whether nodules (relatively low porosities) or cements (relatively high porosities) were formed (*cf.* Reeves 1976). The mineralogy of the sediments determined whther carbonate or gypsum caliche was formed. Other described characteristics of the fine member will be linked with the sedimentation processes.

The large amounts of caliche indicate an evaporation surplus. This implies a hot climate with major dry periods for the Monllobat Formation, even though investigations of pollen pointed to a humid tropical climate (Haseldonkx 1973; see Van der Meulen 1982 for discussion).

## V.4.2 Coarse member

*Facies 1—proximal conglomerate sheets.* A few channel fills, hundreds of metres wide, can be recognized (Figure 4a, at 2–3 km) and, together with the flatness of discontinuous erosion surfaces, this suggests the action of hundreds of metres wide, 1–5 m deep, gravelly streams. The massive gravel with some coarse-grained sand was probably deposited as thin sheets and longitudinal bars (*cf.* Rust 1975, 1978; Miall 1977, 1978). Rapid accumulation of gravel from a heavily-loaded stream during sheetflooding may have resulted in a minor degree of clast orientation. Minor arrays of well-imbricated flat pebbles could be due to reworking at bar tops.

The gravelly braidplains would have resembled recent proglacial outwash (e.g. Boothroyd and Ashley 1975). Despite the shallow depth there would have been variation of stream activity perpendicular to the axis of the stream above the gravel sheets (*Cf.* Clague 1975), with sandy deposits (facies type 2) outside of the main stream area. Narrow, marginal channels could have had a pulsating discharge (*cf.* Clague 1975), which would explain the regular alternation of gravel and sand beds. Lateral migration of the active stream over these deposits would have resulted in the development of extensive erosion surfaces combined with grain-size alternations. Prior to final abandonment thin sands were deposited on top of the gravels.

*Facies 2—distal conglomerate and sandstone sheets.* Outside the main gravelly stream, sandy deposits developed under currents deflected from southwesterly to westerly and northwesterly directions. The sediment fining and the current deflection could be due to simultaneous lowering and directional change of the gradients. The sand and some gravel were deposited in shallow channels and at transverse bars, forming a sandtail of relatively small dimensions to the gravel sheet. Coarsening- and fining-upwards sequences, together with rotation of the current directions, could be explained by lateral migration of the stream over its deposits, accompanied by aggradation, as well as by progradation in a downstream direction.

*Facies 3—symmetrical channel fills.* Away from the gravel sheets there was considerable aggradation with extensive fine member sheets with winged channel fills. Channel sedimentation took place on thin gravel sheets and longitudinal and transverse bars. The presence of siltstone wings implies the restriction of bed load transport to scoured channels, while the fines travelled in suspension-loaded thin sheets of water. Such a situation is described from a recent terminal fan by Parkash, Awashti and Gohain (1983). In some areas channel scouring and filling would have prevailed, leading to type 3b deposits, whereas in others sheetflooding dominated with only minor channelling, depositing type 3a facies. Channels can be absent from the fine member over large distances and then there is a complete analogy with the thin blankets of fine sediment resulting from a sheetflood process, described by McGee (1897). McGee (1897) thought that channelling was prohibited because of the overloading with fine sediment. McGowen (1970) has also pointed to the positive effect of high suspension loads on sheetflooding.

Sedimentary structures pointing to sheetflood conditions are known from fine-grained recent (Gustavson, Ashley and Boothroyd 1975; Parkash, Awashti and Gohain 1983; Sneh 1983) and ancient environments (Heward 1978; Hubert and Hyde 1982; Tunbridge 1981a, b, 1984), but were probably destroyed by soil processes after sedimentation of Monllobat deposits. Only anomously high groundwater tables could preserve the structures as above Castigaleu deposits (in Figure 3 and top of Figure 4c). Here also *Steinichnus* burrows were produced by an arthropod in wet mud under subaerial conditions.

The sedimentary characteristics nevertheless imply that sheetflood conditions dominated the sedimentation, both on the floodplain and the braidplain. The position of the gravel sheets indicates the presence of eight sheetflood systems (*sensu* Friend 1983) side by side in a fixed position at the Pyrenean mountain front (Van der Meulen in press). A bajada was in existence with a gradual distribution of sedimentation rates leading to a flat-bedded fine member instead of lobes (e.g. Gustavson, Ashley and Boothroyd 1975; Parkash, Awashti and Gohain 1983). A similar situation with networks of low-sinuosity, sandy streams succeeded by planar sheetfloods with sand and mud was interpreted by Tunbridge (1981a, b, 1984), as analogous to recent environments with downstream lengths of tens to hundreds of kilometres. However, sheetfloods passing the proximal areas without sedimentation (*cf.* McGee 1897) or minor sedimentation did also occur in the case of the Monllobat environment. Upstream winged channel fills connected with downstream gravel sheets could originate in such situations.

*Facies 4—foreset lobes.* The mottling of the fine member surrounding type 4a and 4b conglomeratic lobes points to an origin in a subaerial environment. The lobate geometry and the flatness of associated fine member beds indicate an origin at the transition from braidplains and channels to floodplain. Cross bedding in the low-angled (type 4a) lobes points to sedimentation from partly maintained alluvial flow. High-angled (type 4b) lobes developed when flow separation cells occurred at the foreset. The southwest–northwest spread of foreset directions is consistent with an origin as crevasse lobes prograding onto the floodplain.

The spatial setting of sand–siltstone (type 4c) lobes at the oxidized tops of fan-deltas (as illustrated by Van der Meulen 1983) suggests formation of bars at stream bed rises rather than as crevasse lobes. Nevertheless, the terminal position indicates deposition from decelerating currents. The dominance of southwesterly-dipping foresets, otherwise occurring only in the sandy cross beds in conglomerates, suggests deposition from initially high energy currents.

In flume experiments, Saunderson and Lockett (1981, 1983) described humpback dunes at the transition from dunes to plane lamination. Allen (1983) described humpback bars as large isolated features resembling humpback dunes. The flat topset and convex-upwards foreset geometry of the sand-silt lobe of type 4c are analogous with Allen's humpback bars. However, in the Monllobat Formation internal parallel lamination cannot be recognized because of homogenization of the beds by soil processes. Although the gravel-sized particles of Allen's examples are lacking, there is nevertheless a well-established fining-upwards of individual beds. Therefore an origin as humpback bar is suggested for facies type 4c with decelerating sheetfloods, loaded with sand and silt, depositing the lobes.

The angle of foresets in Figure 5 is directly related to the amount of flow separation. The interpreted deceleration of currents is expressed by the transition from low to high angles at the lobe termination in Figure 5c, where the sediment load also becomes finer. The transition from a high-angle bar-form to the humpback bar only took place after aggradation and with a considerable decrease in grain-size of the sediment.

*Facies 5—lateral accretion lobes.* The development of single meander lobes, grouped as strings distal to a sheetflood environment, can explain the variability of point bar deposits described. Complete cycles of meander lobe evolution were preserved, in contrast to meander belt settings, where one specific curvature preferentially occurs (Hickin 1983; see Bridge and Diemer 1983). With increasing curvature subenvironments of the point bar were thought to have a differentiated evolution in the meander lobe model (Van der Meulen 1982), although also in this case there is a dominant lithofacies.

In the model the middle part of the point bar produced the dominant facies since it covered the largest area and had maximum growth rates. Consequently transverse sections of this part of the lobe are widest (as in Figure 7a, compared to the section in Figure 6c). In the central parts flow separation at the upper steeply inclined point bar caused sand deposition during maximum discharge. During falling discharge mud was deposited. The bed load was deposited in trough cross beds at the point bar base, partly as incipient scroll bars in scours, during maximum discharge. As a supplementary lithofacies, foresets accreted at the margin of flat-topped platforms at the point bar apex, but apparently the platform could also grow by aggradation by cross bed as in Figure 6c.

A transition from lateral to downstream migration in the last point bar growth stage (described from recent meander lobes by Hickin 1974) caused a major downstream enlargement of the apical platform and consequently it is found underneath most of the abandoned channel fill. However, an earlier enlargement (as in Figure 7a) could be due to an important break of the regular growth pattern of the point bar. Such breaks had been observed before in the top surface of a lobe (Van der Meulen 1982). Incision of the terminal point bars (Figure 7a) points to chute incision. Although not incorporated in the model this is yet a process likely to occur during downstream migration when the potential of point bar erosion was largest.

The trend of palaeocurrent directions was linked to the presence of a large helicoidal circulation cell as generally is present in such environments (e.g. Jackson 1976). The apical platform must have formed owing to interference with the upstream, reversed circulation cell, but the refined point bar growth model of Jackson (1976) could not be applied further. Incipient scroll bars resulting from the helicoidal flow originated preferentially in scours (Figure 8). This scouring phenomenon, together with chute incision, may have been promoted by the presence of the river downstream from a sheetflood environment. Such a situation has been described by Tunbridge (1984). The relatively coarse sediment in Figure 7b and the relatively coarse initial part of the section in Figure 7a could result from the proximity of the sheetflood environment. The flow direction along linear depressions above deep-seated faults (see also Van der Meulen in press) would have promoted meandering river formation.

It is concluded that dimensions, lithofacies and palaeocurrent directions in sections of type 5 deposits can be directly related to the positions in a meander lobe, although the model needed some extension. Application of the model shows that Figures 7a and 7b concern a middle part section and Figure 6c a section in between middle part and apex.

Leo and Allen (1984) presented a quite different environmental interpretation for two Monllobat examples not studied by this author. In their view, inclined mud/sandstone interbedding, an occasional mud-draped cross bed and the juxtaposition of conglomeratic and sand/mudstone point bar series, are comparable with features of recent tidally-influenced meandering rivers in France. However the three features used by Leo and Allen (1984) are not exclusive to tidal action. Sand–mud interbedding has also been described from suspension-loaded streams with no tidal influence (Jackson 1981). Mustafa Alam, Crook and Taylor (1985) have described mud-draped and herringbone cross beds from purely fluviatile settings where there is the possibility of reversible water surface gradients. The setting in tectonic depressions may have caused this in Monllobat environments (e.g. the deposits in Figure 6b and 7b have opposed palaeocurrent trends). Other causes for mud drapes in tabular cross beds are the intercalation of mud beds in tabular bedforms on a point bar with undulating relief (Pasierbiewicz 1982) and intercalation of muddy detritus, eroded from falling discharge mud drapes on the point bar and cut bank, deposited in rising stage tabular cross beds. The oxidation and mottling of the major part of the deposits (*cf.* Van der Meulen 1982) point to seasonal discharges, with the minor possibility of tidal flooding.

The initial conglomeratic point bars are interpreted by Leo and Allen (1984) as being succeeded by sand/mudstone ones, because transgressions moved the flow decelerating tidal floods far upstream. Relatively coarse deposits built during initial channel incision (Van der Meulen 1982) seem a reasonable alternative, especially as the distal part of a sheetflood environment is concerned (Van der Meulen in press). The completely tidal segment of the recent environment, coarsening-upwards tidal bars (Leo and Allen 1984), has not been found in the marine equivalent of the Monllobat Formation.

*Facies 6—structureless sandstone beds.* These beds replace siltstone beds in silt/mudstone alternations. The regular occurrence (near the contact zone with the Castigaleu Formation) suggests that the coarsest sand fractions could be transported as suspension loads by sheetfloods, even when some gravel is intercalated.

The position of the flat sandstone beds and other coarse member types in the fine member couplets indicates that one couplet represents one progradation event of a sheetflood system in the floodplain. Gradual thickness variation of the couplets must therefore reflect a gradual variation of progradation and retreat rates.

V.4.2.1. *Oncoid formation.* Transitions of unattached to attached oncoids and of densely-laminated to fenestrate fabrics have been related to decreasing energy conditions (Nickel 1983). Pebbles and small objects were coated in highly active and marginal channel areas respectively. In lakes, flat oncoids with several growth nuclei were not often transported. Fenestrate-structured, flat cushions originated in the deepest part of an abandoned channel, attached to the lower point bar. Fenestrate-structured pillows were developed in shallow depressions in front of transversely accreted lobes.

## V.5. Description of the Castigaleu Formation

### V.5.1. Fine member

The fine member is characterized by massive, blue-grey mudstones devoid of intercalations of siltstone. Fossils are scarce.

### V.5.2. Coarse member

The coarse member comprises several types of lobes. These are described separately, since one type usually dominates in individual exposures. The thickness of the deposits is mostly less than 10 m. The thickness can be equal to the one of the upstream sheet or channel fill, but is generally larger.

*Facies 1—low-angled conglomeratic foreset lobes.* The foresets of these lobes, which dip up to 15°, consist of gravel with variable amounts of sand and therefore variable packing. Intercalated (pebbly) sandstone beds are cross bedded. The flat upper part (topset) can show large-scale cross beds in conglomerate, but finer beds may also occur. Especially thick layers (5–10 m thickness) may have flat bottomsets of tabular cross-bedded and cross-laminated sandstone. Basal and top surfaces are undulating both in and perpendicular to the palaeocurrent direction.

*Facies 2—high-angled conglomeratic foreset lobes.* These lobes have foresets dipping at angles of 15–30°. They consist of massive conglomerates and conglomeratic sandstone, often with a fining-upwards sequence towards the top of a layer.

*Facies 3—low-angled sandstone foreset lobes.* These lobes comprise tabular and trough cross beds in foresets dipping at less than 15°. The bases of the foresets can be strongly erosive. The top can be largely flat, however with arcuate features on a scale of tens of metres. Isolated bodies with convex-upwards geometries have also been found. The sequence is often fining upwards.

*Facies 4—high-angled sandstone foreset lobes.* These comprise centimetre to decimetre thick foresets dipping at angles of 15–25° with relatively flat upper parts. The top part may be fining upwards.

*Facies 5—flat layers of conglomerate.* These comprise closely-packed pebbles in extensive layers with flat base and top. They occur in the contact zone of the Castigaleu with the Monllobat Formation. The thickness ranges up to 5 m.

*Facies 6—tabular cross-bedded sandstone layers.* Layers of tabular cross-bedded sandstone can be very extensive with flat bases and tops. However, there are also some flat-bedded lobes. Layer thicknesses range from 1–5 m. Set thicknesses and grain-sizes are mostly constant between base and top. Occasionally there is more than one unit, each with a constant set-thickness and grain-size, which generally is in the range of coarse- to fine-grained sand. The set thickness is in the range of 0·1–1·5 m. The thickest sets occur solitarily in most cases, although a 0·1 m thick set may be present at the base.

*Facies 7—parallel-laminated sandstone layers.* Layers with parallel lamination are flat and extensive with thicknesses of 1–5 m. Layers always comprise several beds, 0·1–0·5 m thick. There is hardly any grain-size variation in vertical successions, except that a relatively coarse or fine bed can be present at the base or on top. The grain-size is in the range from fine to coarse sand.

These layers may contain considerable amounts of plant debris, especially in Section H near to the Monllobat Formation. The occurrence of the debris is linked with dissolution holes and with concretions and calcified burrows. Bioturbation—occasionally present in Section H—is much more abundant in layers of this type in Section J and K (more distant from the Monllobat Formation, Figure 1).

*Facies* 8—Flat beds of massive sandstone. These are medium- to very coarse-grained sandstone beds, 0·1–0·3 m thick. A lag of pebbles may be present and some beds show fining upwards. The beds are mostly single, successions of a few beds have mudstone interbeds.

V.5.2.1.*Lateral variation of Castigaleu deposits.* Foreset lobes often occur in areas adjacent to the Monllobat Formation. Conglomeratic lobes, in particular, are concentrated in the transition zone. Lobate sheet margins on top of blue–grey mudstone contribute a significant number of lobes. Foresets with mottled tops are mostly found as in the western parts of the exposure shown in Figure 4b. Very low-angled foresets may contain large foresets of a lower order as at the base of Figure 4c. Nearly flat cross beds with opposed directions form the middle part of the section. Some sandstones contain fossil oysters and blue–grey mudstones are intercalated. The layer as a whole thickens at the margin of a shallow syncline from 4–20 m laterally over 200 m in the palaeocurrent direction.

Flat-bedded sheet margins are rare, but do occur. Conglomerate sheets passing into unidirectional tabular cross beds have been observed and a flat-layered sandstone sheet lies in an eastwards extension of the Castigaleu Formation near Puente de Montañana (Figure 1). Coarse-grained tabular cross beds (1–1·5 m thick) lie at the flanks of very large trough cross beds on top of tabular cross-bedding of a smaller size and grain-size (Figure 9). A channel fill of sandstone and mudstone at the base of the central parts erodes the underlying blue–grey mudstone and caliche layer. An oyster reef lies at the southeastern margin and oyster debris occurs in the sandstones.

Away from the contact zone with the Monllobat Formation, the Castigaleu Formation is exposed in a valley to the south of Badias (Figure 1). To the east of this valley, parallel-laminated sandstone beds dominate exposures near to the contact zone (Section H). In the Badias Valley itself thick (up to 1·5 m), tabular sets decrease in thickness southwards. Further to the south, finer, parallel-laminated and massive sandstone beds are found (Section J), the former with an occasional, convoluted tabular cross bed. At the base of the valley there are some thin mottled mudstone intervals and low-angled sandstone lobes–occasionally

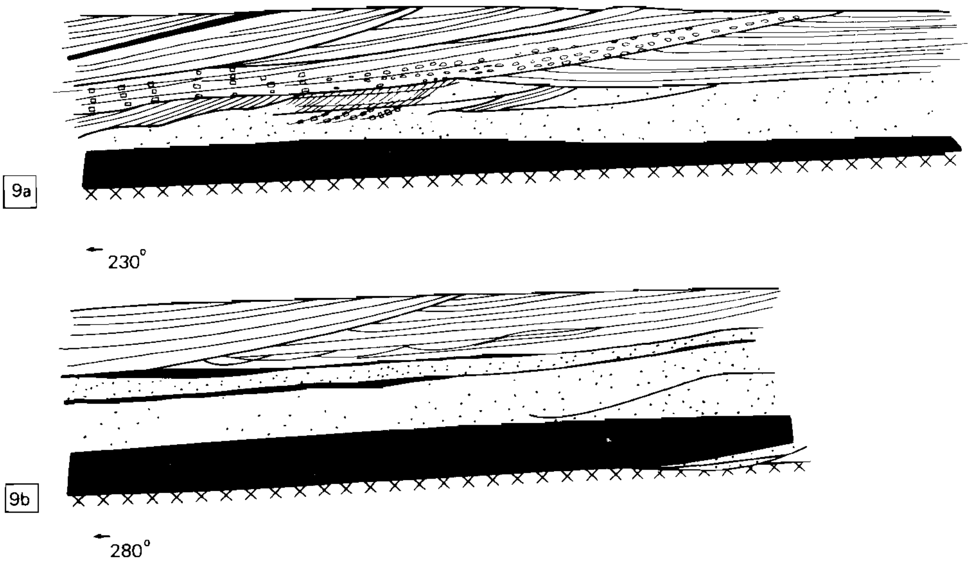


Figure 9. Very large trough-shaped structures on top of a tabular cross-bedded interval. (a) Section in the palaeocurrent direction. (b) Section oblique to the palaeocurrent direction. Location in Figure 1.

with mottled tops. There are single and multiple lobes, the latter with strongly radiating axes. The upwards facies change is linked with a large-scale transgression in the upper parts of the Formations (Van der Meulen in press). The eastwards extension of the Castigaleu layer in Figure 9a, b is also due to this transgression.

V.5.2.2. *Tectonic features and the variability of geometries and palaeocurrent directions.* Sheet margins may thicken in tectonically-deformed zones, as shown in Figure 4c. Away from the contact zone with the Monllobat Formation, geometries may also have a relation with tectonic anomalies, reflected by changes in dips and dip directions. In the Badias Valley the exposure of Figure 10b shows a large foreset in the direction of the suddenly increasing, southeastwards dip. At the top the foreset is connected with a flat, planar-bedded layer, which passes without thickness change into the inclined layer with descending cross-bedded sets. Cross-bedded sets of the upper parts have some mudstone interbeds, containing gravel, large mud clasts and flattened tree trunks (Figure 10b). Figure 10a shows a layer from the Badias Valley with a northeastwards dip, opposed to the regional dip direction and also to the palaeocurrent direction. The tabular cross beds pass distally in concave-upwards, wedging beds. A caliche bed with some mottles lies 0·5 m underneath the base.

The dip directions of the layers in the northern parts of the Badias Valley are variable and in several cases opposed to the regional southwesterly dip directions. The palaeocurrent directions of the tabular sets in this area are highly variable. Further to the south only the dip of the layers varies and the palaeocurrent directions are unidirectional. Faults traversing the valley are often associated with changes in amount of dip and dip direction.

An example of tabular sets with highly variable directions in a lacustrine setting is present in Section B, which lies in the vicinity of a fault (the meander lobe of

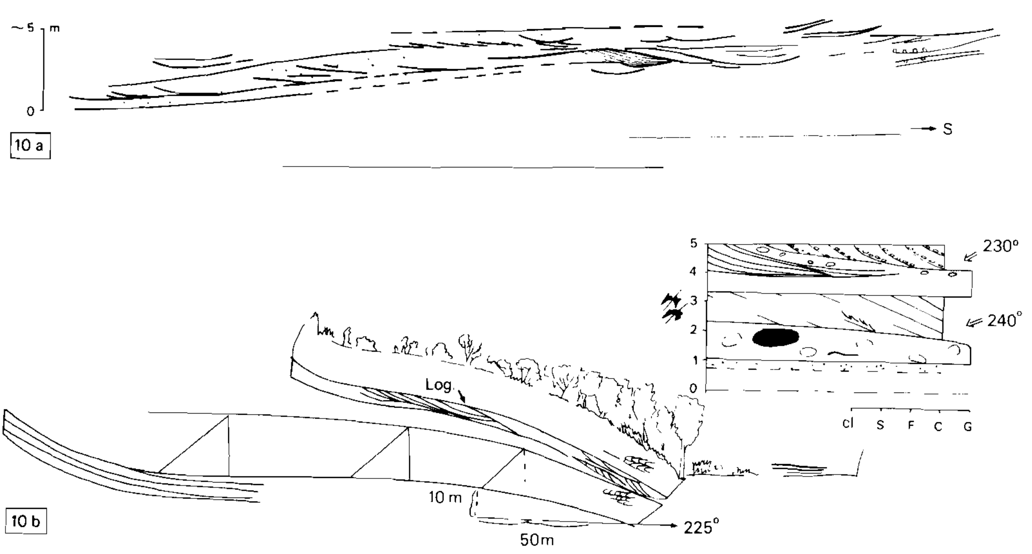


Figure 10. (a) The tabular cross-bedded sandstone layer passes into wedging layers with upturned ends. (b) The large foreset with flat top is constituted by a layer with constant thickness. Mudstone layers with pebbles, large mud clasts and flattened tree trunks are intercalated. Locations in Figure 1.

Figure 7b lies directly adjacent to this fault). A conglomerate sheet with southwesterly imbrication directions wedges at the fault. The equivalent layer on the other side consists of tabular cross bedded sandstone. Palaeocurrent directions of 275° and 325° are succeeded upwards by 55° directions. Laterally, westerly directions are found.

V.5.2.3. *Distribution of palaeocurrent directions.* Foreset lobes of conglomerate have the same directional distribution as Monllobat lobes of this type (Figure 11a and 11c). Castigaleu sandstone lobes have a highly variable distribution, in contrast to the sand–siltstone foreset lobes of the Monllobat Formation (Figure 11b and 11d). The major directional variation is shown by the tabular cross-bedded sandstone facies (Figure 11e), which also goes for the lacustrine tabular cross beds.

## V.6. Distribution of facies types

The distribution of the Monllobat and Castigaleu facies types throughout 11 sections is given in Table 1. A subdivision of the Monllobat deposits has been added as 'the lacustrine suite', comprising layers enveloped by blue–grey mudstone but without contact with the Castigaleu Formation. These consist of lobes (type 1) or tabular cross-bedded sandstone (type 2). Conglomerate layers with mottled tops, but with erosive bases on blue–grey mudstone are put in class 1 or 3 of the alluvial Monllobat suite on the basis of the mottling and geometry.

There is a link between the position of the sections and facies types present. Sections A, B and E comprise only continental types, part of which consist of conglomerates. Section D, closer to the Castigaleu Formation, shows only small conglomeratic channel fills, silty foreset lobes and, beside lacustrine deposits, two

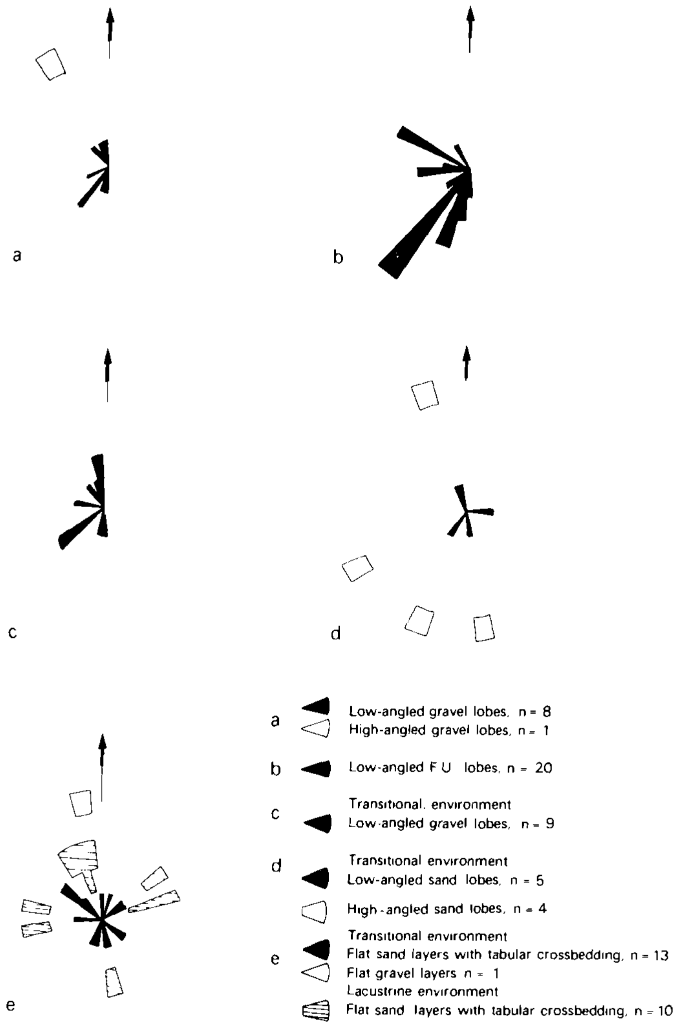


Figure 11. The distributions of foreset directions of various lobetypes and of palaeocurrent directions. (a) and (b) comprise readings from the Monllobat Formation. (c), (d), (e) are from the Castigaleu Formation. Readings from Monllobat layers with a lacustrine origin are added in (e).

Castigaleu layers. Sections F, G and I at the Formation boundary have about equal proportions of Monllobat and Castigaleu facies types. Over the boundary, Section H and J consist of Castigaleu layers only. Section K at the southern basin margin lies close to the formation boundary and comprises mainly continental types intercalated with mottled mudstones.

Monllobat coarse member sheets are almost absent from the sections. This is partly explained by the wedging and partly by the choice of section locations which overall are best exposed.

Table 1. The number of occurrences of facies types in the sections A–K. Locations of the sections are in Figure 1

| | Sections | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | A | B | C | D | E | F | G | H | I | J | K |
| Monllobat facies (alluvial) | | | | | | | | | | | |
| 1 Conglomerate sheets | 1 | | | | | | | | | | |
| 2 Sandstone sheets (pebbly) | | | | | | | | | | | |
| 3a small | | | | 4 | 3 | | | | | | |
| 3 Symmetrical channel fills | | | | | | | | | | | |
| 3b large | 1 | 2 | | | 2 | 1 | | | | | |
| 4a L.-a. foresets | 2 | | 1 | | 3 | | | | | | |
| 4 Conglomeratic lobes | | | | | | | | | | | |
| 4b H.-a, foresets | | | | | 1 | | | | | | |
| 4c Silty foreset lobes | | 1 | 3 | 3 | 1 | 4 | 2 | | 1 | | 5 |
| 5 Lateral accretion lobes | | 1 | | | | | | | | | |
| 6 Structureless sandstone | 1 | 8 | 8 | | 1 | 1 | 1 | | 1 | | 2 |
| Monllobat facies (lacustrine) | | | | | | | | | | | |
| 1 Lobes | 1 | | | 2 | | | | | | | 2 |
| 2 Tabular cross beds | | 1 | | | 1 | | | | | | 6 |
| Castigaleu facies | | | | | | | | | | | |
| 1 Conglomeratic foresets (l.a.) | | | | 1 | | | | 1 | | | |
| 2 Conglomeratic foresets (h.a.) | | | | | | | | | | | |
| 3 Sandstone foresets (l.a.) | | | | | | | 1 | 1 | 2 | 2 | |
| 4 Sandstone foresets (h.a.) | | | | | | | | | | 1 | |
| 5 Flat conglomerate layers | | | | | | | | 3 | | | |
| 6 Tabular cross bedded sandstone | | | | | | 4 | | | 3 | | 1 |
| 7 Parallel-laminated sandstone | | | | | | | 1 | 2 | | 2 | |
| 8 Massive sandstone | | | | 1 | | 1 | 4 | 5 | | 3 | 2 |

## V.7. Interpretation of the Castigaleu Formation

### V.7.1. Interpretation of inflow processes at the coastline

The distribution of facies types shows that near to the coastline sheetflooding (especially of fines) was the most important depositional process. Only a small part of the gravelly sheetfloods and channels reached the coastline.

The variability of alluvial inflow is primarily due to the dynamics of the stream, the sediment type and the depth ratio (Jopling 1963). The deposits resulting from inflow (facies type 1–8) can therefore be related to depth ratios and grain-sizes (Jopling 1965), since the grain-size indicates the ratio of flow velocity/fall velocity of the grains. The relative positions of the Castigaleu types in a grain-size/depth ratio diagram are presented in Figure 12.

In a proglacial outwash environment comparable to the situation with gravelly sheetfloods and channels, Church and Gilbert (1975) distinguished distally a back water area at the coastline and various types of inflow in a standing water environment. The proglacial inflow is characterized by high energy floods and by the supply of large amounts of sediment, two features which can also be inferred

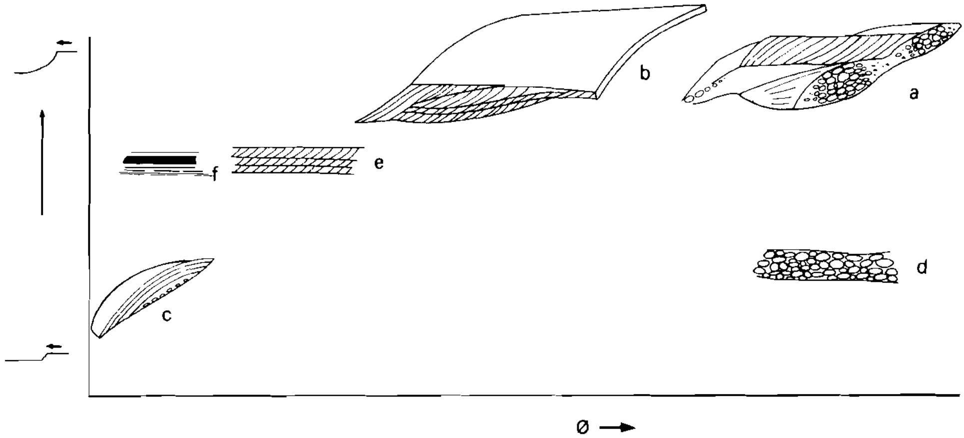


Figure 12. The variation of deposits together with depth ratio and grain-size. (a) and (b) represent gravel and sand lobes. (c) is a transversally accreted sand–silt lobe. (d) is a flat layer of gravel. (e) consists of flat beds of tabular cross bedding and (f) of parallel-laminated or massive flat beds.

for the Monllobat alluvial systems. Part of the transport capacity and competence is lost in the backwater area which causes a sediment wedge to grow in both up- and downstream directions in order to attain equilibrium. Upstream the stream bed will aggrade, downstream an inclined front will develop during progradation. The depth ratio (stream/basin depth) determines the size of the consequently formed sediment wedge. Topset–foreset–bottomset build-ups have been described from a similar setting in Pleistocene lakes adjacent to outwash (Thomas 1984). Generally the depth ratio at the Castigaleu coastline will not have been less than 0·5. Only differential subsidence has led to lower ratios as in the case of the wedge in Figure 4c.

Outflow processes at a fan-deltaic front have been interpreted previously (Van der Meulen 1983), on the basis of a three-dimensionally exposed conglomerate wedge. Low-angled gravel foresets (type 1) were distally succeeded by a few high-angled ones (type 2) with a decrease of energy (see Figure 12-(a)). The flow became decelerated by a combination of inertial and frictional forces (type 1) and by inertial forces alone (type 2). Type 1 foresets are often connected with bottomsets as transport beyond the foresets was possible. Aggradation of the upstream area constituted topsets. The outflow relief at the top of the wedge was determined by longitudinal thickenings with convex-upwards geometry. This could involve the formation of arcuate foresets.

Low- and high-angled sand lobes (types 3 and 4; see Figure 12-(b)) situated at sheet margins must have originated in a similar way to gravel lobes. However, the outflow relief was less pronounced. The wide variation of foreset directions in the exposures points to distributive progradation from the orifices. This, together with the smaller grain-size, points to a lower flow velocity than in the case of types 1 and 2. The rather constant foreset dips of most Castigaleu lobes indicate more regular outflow processes at the coastline than during lobe formation in the continental environment.

A sand–silt foreset lobe of the Monllobat suite is shown in Figure 12-(c) as this facies type was often formed at the top of deposition types a and b. This combination of Castigaleu and Monllobat types is due to a major aggradation during fan-delta development. Major aggradation is also known from Hjullström delta types (Church and Gilbert 1975), resulting from high competence streams entering a shallow marine environment.

### V.7.2. Interpretation of the inflow in the shallow marine environment

Facies types 5–8 are mostly found in extensive flat layers. Massive flat gravel layers (type 5; see Figure 12-(d)) were formed when there was only minor energy loss at the coastline, due to a minor break of slope. This type of deposit is very scarce in the Castigaleu Formation, but most lacustrine environments were filled by flows not losing their alluvial properties during inflow or only after a considerable journey in the new environment. Transitions from massive conglomerates to tabular cross beds in a flat layer must have taken place in such a situation with very gradual flow deceleration (see Figure 12-(d) and (e) respectively).

Tabular cross beds accompanied by some parallel lamination and massive beds are also found in flat lenses in front of the conglomerate wedge studied in detail by Van der Meulen (1983). These deposits were interpreted as having formed by hyperpycnal inflow beyond the inclined fan-delta face. This type of inflow, defined by Bates (1953), is known to descend beneath the standing marine water during proglacial sheetflooding (e.g. Church and Gilbert 1975). The heavy sediment load causes a relatively higher density of the descending flow which forms a turbidity current. Kuenen (1964) has stated that, in general, sheetfloods passing the coastline will become turbidity currents. He inferred sheetflooding in the coastal environment from parallel-laminated sand beds with basal current marks as described by Cummins (1958). On the basis of these considerations, hyperpycnal inflow by originally gravel-carrying floods, as well as fine sheetfloods, is interpreted to have formed facies types 6 (see Figure 12-(e)) and 7 and 8 (see Figure 12-(f)). However, coastal wedges will have been a less common feature of the fine sheetfloods as the influence of competence loss was less important than in the case of gravelly floods.

### V.7.3. Variability of sedimentation by hyperpycnal flows

Inflow extending beyond the coastline as hyperpycnal flows soon became self-propelled due to the action of gravity on the sediment. The lack of vertical grain-size and structure variation of the sediments indicates deposition during bypassing of a hyperpycnal flow which was strongly mixed. A true turbidity current with a differentiated sediment load will, in most cases, not have developed.

The episodic nature of the inflow is indicated by bioturbation taking place after the sedimentation of each parallel-laminated sand bed. The top of successions is often massive because of thorough bioturbation after cessation of sand deposition. The massive character of isolated flat beds may have a similar origin, but could also have been due to sediment dumping. The sandy sediments were also preferred sites for oyster growth. Episodic flood events would occasionally have removed them leading to debris in the sandstones. The restriction of oysters to the coarse members may be linked to the flooding process, since the open circulation with a more fully marine environment may have only existed along the pathways of the hyperpycnal flows.

Deposition from the self-propelled flows was no longer dependent on the depth ratio at the coastline, but on the geometry and gradients of the standing water basin. To the west, the largest distribution area existed and the hyperpycnal flows became most gradually decelerated (by bottom friction). With a downstream fining, fields of tabular cross beds were succeeded by thin parallel-laminated sheets (Section J, compare Figure 1). In the area of Section H, relatively coarse parallel-laminated sheets lie close to the ancient coastline, pointing to a more rapid decleration in a smaller distribution area.

To the east, the shallow marine environment narrowed down to offshoots in which seawards transport met major obstructions. The sand sheet of Figure 9 originated after incision of the floor of the offshoot. After subsequent channel filling the depth was further decreased by aggradation of a cross-bedded sheet during the next flood. Floods of the final filling phase became largely blocked, building large erosively bound cross-beds in the central area and regularly prograding forms to the sides. In a small offshoot in the eastern area, transport of hyperpycnal flows seawards from a gravelly fan-delta was only possible after a major deflection of the primary southwesterly current direction. Together with continuing aggradation, first lenses and ultimately backsets and flat sheetflood deposits developed when the transport became totally blocked (Van der Meulen 1983).

Gradient anomalies due to synsedimentary tectonic activity were reflected in the megaripple directions in the marine as well as the lacustrine environment. The major susceptibility of this structure to gradient variation indicates the relatively lowest energy levels during formation, as compared to other types with directional properties. In some cases, differential subsidence caused gradient variations of such a scale that geometries were strongly influenced. The effects of an opposed gradient and of a break in slope are illustrated in Figures 10a and 10b respectively. The geometry in Figure 10a is also found underneath the backsets of the example with blocked hyperpycnal flow in the previous section.

### V.7.4. *Reworking processes*

In the preceding interpretation fan-delta processes were constructed assuming high energy inflow, extending considerable distances into the marine environment. High energy alluvial processes are deduced from the interpretation of well-exposed Monllobat deposits and there are transitions between alluvial and fan-deltaic deposits. On the other hand the influence of marine reworking should be considered, for it is known that episodic stream deposits tend to have a low preservation potential since there are long periods available to marine reworking (Glennie 1970). Large proglacial fan–deltas show major redistribution by waves and tides, with wave action as a strong control on geometry (Hayes and Michel 1982). In the case of the Castigaleu environment reworking of the gravel and sand lobes will have been limited because of the major aggradation and the burial underneath mud. There are no indications of continuous sand acccumulation at the coastline (possible wave-built barriers) and beach deposits are lacking. Major accumulations of the Castigaleu coarse member all lie in line with Monllobat alluvial systems.

Criteria for the recognition of tidal deposits, as listed by DeVries Klein (1977) and Reineck and Singh (1980) are generally not met by the shallow marine deposits: a bimodal size division of the sedimentary structures in cross-bedding and cross-lamination is not found and neither are reactivation surfaces and abundant mud drapes. Desiccation cracks are absent. There is no evidence of reworking of mud in tidal channels, unlike many recent tidal deposits. The directional variability of tabular cross beds in the shallow marine environment might be an indication for tidal processes. However, the setting implies an origin by tectonically influenced sedimentation. The same directional variability is found in lacustrine environments. In a few instances deviating cross-bed directions are inherent to the sedimentary environment. Because of high energy inflow major return flows could develop over the fan-delta face building cross-bedding (see also Van der Meulen 1983). An example is shown in Figure 4c.

The overall palaeogeographical setting and the mineralogical composition of the sandstones both suggest that marine reworking was not a major factor in the Castigaleu environment. There was salt diapirism at the entrance to the inland Tremp–Graus Basin, with large alluvial fans further seawards from the described area (Nijman and Nio 1975). Extraformational limestone clasts dominate sandstones at either side of the coastline. Quartz arenites dominate many environments with marine reworking (e.g. DeVries Klein 1977; Cotter 1983; Teyssen 1984), even in fan-deltaic environments. Button and Vos (1977), Tankard and Hobday (1977) and Vos and Erikson (1977) describe fan-deltaic deposits with a distinct change from lithic arenites and arkoses to quartz arenites over the coastline. These considerations, together with the sedimentary facies and fossil content, suggest that the Castigaleu Formation was deposited in a lagoonal environment with coarse incursions from sheetfloods.

## V.8. Discussion and conclusions

Sheetflood systems are the most likely origin of the Monllobat conglomerate sheets on the basis of the very large W/D ratios for the streams and the massive pebble packing (environment M1 in Figure 13). The small size of distal sand tails (M2 in Figure 13) deposited by decelerated and deflected currents and rapid transitions to wide fine member sheets support this interpretation. The connection of conglomeratic channel fills with fine wings (M3 in Figure 13), conglomeratic crevasse lobes (M4a, b in Figure 13) and humpback bars (M4c in Figure 13) point also to the high energy level of the Monllobat environment. The vertical variation in caliche development and mottling of the fine member (M3 in

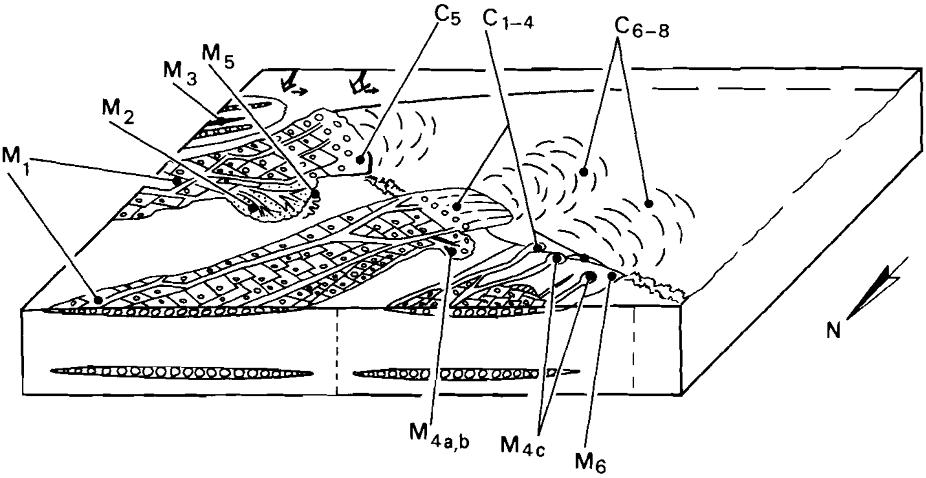


Figure 13. An impression of Monllobat sheetflood systems discharging on land (to the east) and in the Castigaleu lagoon. Monllobat facies comprise gravel sheets (M1), pebbly sand sheets (M2), winged symmetrical channel fills in silt–mud sheets (M3), gravel and sand–silt lobes (M4a, b and c), meander lobes (M5) and flat sand beds (M6). The Castigaleu facies association consists of gravel (C1, 2) and sand lobes (C3, 4), gravel sheets (C5) and tabular cross-bedded (C6), parallel-laminated (C7) and massive (C8) sand beds.

Figure 13) was directly related to the episodic/thick deposit event type of sedimentation. Occasionally the grain-size could be larger than silt (M6 in Figure 13). Oncoid structures reflect the positions relative to the highest energy sheetfloods. Point bar deposits (M5 in Figure 13) are well preserved, but originated in marginal environments with minor distributions.

The sheetflood systems formed the subaerial agents of a large-scale fan-deltaic environment as defined by Wescott and Ethridge (1980). The coastline–lagoonal environment of the Castigaleu Formation represents the transition zone of this classification. The submarine fan part of the fan-delta environment was not present in the study area. Sheetflood deposition in lakes was of minor importance in the Monllobat environment. Otherwise the distinction between sheetflood and transition zone environments may prove to be difficult, as in the case of coarse member wedges enveloped by fines of a wet and lacustrine flood basin, which were termed fan-deltas as a whole by Steel and Aasheim (1978). The term 'fan-delta' has also been used in this paper as a general term for deposits at high energy stream mouths.

Sedimentation processes entirely due to alluvial stream outflow in standing water environments were not considered by Wescott and Ethridge (1980). Their definition of transitional deposits is concerned with reworked sediments only. Daily, Moore and Rust (1980) and Gjelberg and Steel (1983) described ancient transitional fan-deltaic deposits, which partly show reworked features (at the top) and further possess characteristics of alluvial sediments. Hayward (1983) defines a coastal alluvial fan, because transitional deposits could only be distinguished on the basis of additional evidence as fossils and reefs. Conglomerates have completely alluvial properties.

In this paper the conclusion was reached that the depth ratio at the Castigaleu coastline determined the extent to which alluvial inflow in the lagoon became altered. Deposits of non-altered flow (C5 in Figure 13) are rare. In most cases flow deceleration caused the separation of the coarsest sediment from the floods, leading to deposition of gravel and sand lobes (C1–4 in Figure 13) at the coastline. Flat sand beds (C6-8 in Figure 13) were deposited in the lagoon by hyperpycnal flows, which could be deflected by tectonically formed relief elements. The depth ratio was not sufficient to cause slumping of the delta front sediments and consequent mass flows as in the cases described by Wescott and Ethridge (1980, 1983) and Postma (1983).

## chapter VI

# SUMMARY

Sedimentological investigations of the Eocene Monllobat and adjacent Castigaleu Formations in the Southern Pyrenees followed two main lines. In the first place the interrelationships of the coarse and the fine members were studied and detailed sedimentation models were developed. In the second place the distributions of the coarse members and the sedimentary facies were recorded, in order to obtain detailed models of the palaeogeography and the ancient sedimentation processes. The approach to the study comprised three elements:

1 Objective observation of rock characteristics, as much as possible represented in a manner which enables repetition. The field study was based on:
a Centimetre-decimetre scale facies observations in three-dimensional exposures of several $100^2$ $m^2$ with a thickness of up to 20 m, and in 40 - 180 m thick vertical sections.
b Metre-scale observations of the coarse member distributions in valley wall sections in the Ribagorzana drainage area, which are up to tens-of-kilometres wide. Microscopic study of sandstone thin sections supported field definitions of certain types, and was a major tool in hinterland reconstructions.

2 Interpretation of the rock characteristics in the most logic and simple way - summarized in models of the palaeogeography at a large-scale, and also, in some select cases, in great detail, as inferred from ancient sedimentary reliefs.

3 Continuation of and comparison with other research.

In general three types of facies models can be recognized (Reading, 1986), depending on the stage of research. A type model consisting of initial working hypotheses is succeeded by an actual model or realistic interpretation with a general, but not detailed environmental zoning, and, in the ideal case, the study is perfected to a palaeogeographical or local model with an exact representation of the facies distribution. The latter stage can never be reached in large-scale research; the detailed representation of two relatively complex, but very well (3 D type) exposed deposits of only several $100^2$ $m^2$ each, took one complete fieldwork season. Nevertheless the specific Monllobat facies made it possible to trace major coarse members over the cross-sections of the study area, but the distribution of detail facies had to be based on its development in regularly-spaced (kilometres in between) logs. The preceding M Sc thesis (Van der Meulen, 1978) with an inventory of the northeastern parts of the Monllobat Formation was based on regularly-spaced logs, but also on lateral facies organization on the scale of exposures, and therefore just surpassed research stage 2. The initial research in the basin, as summarized by Nijman & Nio (1975) is scaled in between stages 1 and 2, the main effect of which was a major structuring of subsequent work by adoption of the elaborate delta model.

From this study it emerged, that the Monllobat Formation is characterized by kilometre-wide conglomerate and silt- and mudstone sheets derived from the rising Pyrenees to the northeast. A modern environment with similar deposits has been described by Harvey (1984) from alluvial fans in the southeast of Spain. There, extensive gravel and silt sheets make up the lower fan parts. The fine material has been derived from a soil cover of the adjacent mountains. A similar situation in the Monllobat source area is in line with the attribution of coarse member concentrations at distinct Monllobat levels to events of accelerated tectonic activity. During such periods gravel and sand erosion from the relatively high relief would be favored over soil formation.

Kilometre-wide massive conglomerate sheets originate during deposition by heavily-laden sheetfloods (Harvey, 1984, Nemec & Steel, 1984, Hein, 1984, Flint & Turner, 1988). Sediment concentrations lie in the range of 40 - 80 % (Nemec & Steel, 1984). Deposits are thought to range from only slightly channelized and poorly graded to strongly channelized with normal grading and in this case even complete stratification Nemec & Steel, 1984). Gradation in the massive Monllobat conglomerate sheets is poorly developed, however the tails with deflected palaeocurrent trends do show well-developed downstream fining, often combined with decreasing size of sedimentary structures. Adaptation towards a lower gradient with another direction will have involved this transition, which is due to relatively lower sediment contents of the flow. In the sand tail environment sheetflood conditions still prevailed, as is shown by the large width and minor thickness of the deposits (aggradation remained dominant) and the often perpendicular shift of palaeocurrent directions within only metre-thick vertical sections (which requires very wide stream beds). The finest fractions of the sediment were carried by sheetfloods beyond the sandtail, with minor amounts of symmetrical channels to transport the sand and gravel residue. In the fine member, caliche in siltstone beds regularly alternates with massive mudstone. However, towards intercalated symmetrical channel fills the caliche horizon can split up and become interbedded with mudstone. This, together with nodular to cemented caliche transitions towards the fills, points to the major imprint of sedimentary characteristics on caliche formation and also to lateral supply of water with dissolved carbonate.

Additional coarse member types with small volumes comprise foreset lobes, either developed as crevasses of the main sheetfloods in the flood basin, or as drapes over sedimentary reliefs. Meandering river deposits (meander lobes especially) are minor features of the distal area. The setting implies deposition along zones over sites with deep-seated synsedimentary fault activity. The five coarse member types and the fine member sheets together make up the deposits of a 'coarse and fine' sheetflood system as defined by Friend (1983).

Only during the interpreted events of accelerated tectonic activity the coarse Monllobat sheetfloods prograded in the study area, where sheetflooding with fines dominated. The major progradation rates were probably favored by the fixed setting of several 'coarse and fine' sheetflood systems alongside each other on a bajada, because of lateral confinement (Nemec & Steel,1987). Bajadas constituting an orthogonal pattern of depositional systems are most likely to be found in small basins in orogenic situations (Miall, 1981). At present the northeasterly-supplied one lacks the upper to midfan regions, where the incision probably took place and debris flows have been active (cf. Harvey, 1984). The southerly-supplied material only represents the marginal area of the coarse sheetflood environment (with foreset lobes and a single sheet), dominated by fine sheetfloods. This system truncation is linked with a major stratigraphic thinning in an upstream direction. It is stressed that in both cases a similar overall environment of deposition is involved, as is best illustrated by Figures 8a and 9a in chapter 4, which show a combination of fine and coarse member progradation with a coarsening-upwards sequence.

Even at times of maximum progradation the coarse deposits of individual sheetflood systems did not become connected, neither in a transverse, nor in a longitudinal direction, because of the major aggradation with fines along the systems margins. The essential separation of the two bajadas with different source areas is also manifest from the palaeocurrent directions (to the southwest with northwestern deflections and to the north with western deflections respectively). This points to a different course of gradients at the surface of the two bajadas.

The specific type of sedimentation by 'coarse and fine' sheetfloods produces thin, but very extensive coarse layers enveloped by fine sheets. There is a rather uniform facies development with only a few, but rapid lateral facies transitions. This situation is due to major progradation rates. In this case vertical facies analysis in order to establish the relationships of the several coarse member types can prove to meet serious obstacles (Walther's Facies Law - the basis of this approach - states, thatfacies zones have to shift gradually in order to find a parallel in vertical sequences (Middleton, 1973, Reading, 1986)). Therefore, during the investigations the emphasis had to be put on large-scale mapping, detailed three-dimensional study of facies contact zones, and lateral tracing of coarse members. The latter is facilitated by the coarse member setting in a matrix of fines and also to a major degree by the very good outcrop conditions. The additional element of lateral facies analysis also is likely to overcome some imperfections of the vertical facies analysis method (Miall, 1983, 1985 , Reading, 1986), and, on the other hand, the pursued methods automatically resulted in the reconstruction of the larger depositional forms. These forms (as in chs. 2-4) are rather insensitive to hydrodynamic changes during an individual event, but do respond to the overall geomorphological regimen (Jackson, 1975, Friend, 1983).

Because of the palaeogeographical situation, part of the Monllobat sheetflood systems met the coastline of the shallow, restricted marine Castigaleu environment; at that time reaching halfway in the study area from the west-northwest. In this situation, shelf-type fan-deltas developed (cf. Ethridge & Wescott, 1984 and Massari & Colella, 1988). The amount of relief determined the extent of flow deceleration of the sheetflood types. Minor reliefs promoted a relatively gradual deceleration of the flow, which caused deposition of flat beds. Larger relief breaks were linked with lobe development with internal foresets (small Gilbertian fan-deltas). Synsedimentary tectonics are likely to have played a role in this, since fault and flexure zones in this environment are often lined with lobes with internal foresets (ch. 4, fig. 9b,c).

The main fan-delta front facies of the Castigaleu environment, tabular cross-bedded and parallel-laminated sands in a fine matrix, are known from a setting closely related to the sedimentation by high-density sediment gravity flows. The sand beds originate as deposits of residual turbidity currents in the latest stages of proximal, high-density turbidity currents (Lowe, 1982). Eyles et al. (1987) describe trough-crossbedded and parallel-laminated sand from a lacustrine (Pleistocene, supraglacial) setting - also in relation with proximal, high-density turbidity current deposits. As described in chapter 3 the coarsest load of the Monllobat sheetfloods became stored near the coastline - subsidence of the flood in the shallow marine environment is likely to have developed such a residual turbidity current-like agent during thorough mixing of the sand and mud sediment in the decelerated flow. Mud deposition can have been linked directly (deposition of ungraded mud by strong deceleration of fully turbulent turbidity currents, McCave & Jones, 1988) and indirectly (sedimentation from suspension after flood events) with turbidity current development at the fan-delta front.

As in ordinary deltas, fan-deltas of the shelf-type are subject to the interaction of alluvial and marine (especially wave and tidal) processes (Orton, 1988 and Massari & Colella, 1988). Wave action is the most likely agent to be encountered, because of the rather straight fronts of fan-deltas and dominance of aggradational, alluvial processes over incisive ones. The minor number of (meandering) channels in the Monllobat Formation is mostly mottled and oxidized nearly down to the base (arguing against regular tidal incursions over the fan-delta top), and the lack of winnowing of the muddy sediments from the fan-delta front (pointing to a lack of wave action) are both in line with a major lack of marine reworking in the Castigaleu environment (see further discussion in ch. 5).

The recognition in fan-deltaic deposits of minor amounts of sediments with features pointing to tidal action in particular can pose a problem which is specific of the fan-deltaic environment, as parts of the relatively fine range of the alluvially-derived deposits can show similar characteristics. For instance, Tunbridge (1984) describes silt drapes on bar surfaces, and Cherven (1984) shows mud-draped ripples and flasers. These structures are inherent to the major contents of suspended fines of diverse types of the flows. This may be an important factor, when mud-draped crossbedding is considered; De Mowbray & Visser (1984) indicate that mud drapes in their examples consist especially of biologically pelletized and flocculated sediment, since slack water periods are far too short to produce settled-suspension drapes. Furthermore, Breyer & McCabe (1986) state that their examples of tidal, mud-draped cross beds do not show rhythmic patterns associated with tidal oscillations and that such patterns are apparently not abundant in ancient sequences. On the other hand rhythmic patterns can be produced in diverse environments, due to interference of small with large bedforms (Rubin, 1987) and therefore the exact definition of environment on the basis of mud drapes and rhythmic patterns will be even harder.

It should be noted that the Castigaleu crossbedding is not mud-draped. Another tidal feature, strongly variable palaeocurrent directions with even reversals is present at a few locations with direct indications of synsedimentary fault activity underneath the sites, or, as a minor feature, at lobes. These specific occurrences with an overall very minor distribution (chs. 3, 4 and 5) led to the most simple interpretation in terms of the highly-energetic sedimentation processes. In general, it is known that contained turbidity currents can be deflected and even reflected, which produces a wide scatter of palaeocurrent directions (Pickering & Hiscott, 1985). Seeing the shallowness and minor size of the basin (with the possibility of major synsedimentary tectonic relief breaks) involved with shelf-type fan-deltas, the influence of containment on fan-deltaic outflow should also be considered. Lateral return flows had already been identified as producers of countercurrent crossbedding at fan-delta lobes (ch. 3). In deltas lateral and vertical return flows occur during relatively low and high outflow levels respectively (Wright & Coleman, 1974). In this specific type of fan-delta, lateral return flows are likely to dominate during high flood stages, because of the flood size relative to the small depth of the standing water body.

It can be concluded that the model for shelf-type fan-deltas still needs considerable expansion. This can be illustrated by the contrasting interpretations made for the Roda Formation in the Tremp-Graus Basin - a fan-deltaic origin with some tidal influence (Puigdefàbregas & Souquet, 1986 and Puigdefàbregas et al., 1987), as well as a completely tidal origin (eg. Nio, 1976, Molenaar et al., 1988) have been proposed.

## VI.1. Short summaries of chapters 2 - 5

VI.1.1. Chapter 2 A meander lobe (defined as a succession of inner bend deposits - point bars - of a meandering river) has been described and reconstructed in three dimensions, together with the palaeochannel and associated floodplain fines. A detailed facies description from a comparable vertical lobe section is added. It is conspicious, that although the general characteristics of the meandering river sedimentation model are met, the palaeochannel has its own identity. Some specific features are different from the general model, but do occur solely in quite different types of the wide range of recent meandering rivers. Such features are; widening and shallowing at meander bends, sediment differentiation in rather low platforms at the point bar base (bedload) and steeply-inclined upper point bar layers (suspension load), a discontinuous growth pattern with final asymmetric growth and point bar modification and point bar erosion, abundant epsilon cross stratification (ECS) and lateral sediment variation along the point bar.These features are distributed over recent rivers of a relatively high (graveliferous) as well as of a low energy level. In this respect it is conspicuous that a relatively coarse initial part with deviating features has been found in both exposures. This was explained to be the result of the distal position in the sheetflood system.

VI.1.2. Chapter 3 A three-dimensional study has been made of a conglomeratic wedge, with isolated sandstone lenses distally. The exposure is situated at the boundary of the Monllobat and Castigaleu Formations. The proximal wedge consists of lobes with a topset-foreset-bottomset architecture. The sediments grade from gravel to sand to mud over the fan-delta face. Some of the foresets are erosive to bottomset rises. A 4.5 m high foreset face constitutes the downstream boundary of the wedge.

The lower sandstone lenses distally from the proximal wedge have concave -upwards bases and convex-upwards tops in transverse as well as in longitudinal sections. Downstream wedging can also be due to thickening of intercalated mudstone layers. Tabular crossbedding and parallel lamination are the dominant sedimentary structures. Frozen climbing megaripples cover the top of the lenses.

Backsets facing the proximal foreset bound the upper two lenses. The backsets end at a flat top connected with a downstream foreset. Palaeocurrent directions of the main structures of lobes and lenses are to the SW. Small structures have W and NW directions.

The outflow processes at the described fan-delta must have been significantly different from ordinary delta processes. During outflow, the heavily-laden streams dumped its coarsest load (gravel) in lobes. Thin sandy gravel layers were deposited in between lobes - this involves undulating bases and tops in transverse section with several orifices.

Low-angled foresets developed during initial highest energy outflow. With progradation and aggradation low energy avalanching built high-angled foresets.

Finally gravel transport decreased and major sand lenses could be built and preserved in the bottomset area. Sandy backsets were formed during overwash of the strongly aggraded bottomset area.

Subsequently the subaerial top was raised by sandy and muddy sheetfloods. The lobes were covered with sand and silt foresets with an arcuate geometry (these structures resemble meander lobes). The distal area became covered by a sand sheet grading to silt and mud.

VI.1.3. Chapter 4 The sedimentary stratigraphy of the Monllobat Formation is described and interpreted in this chapter.

Description: The studied part of the Monllobat Formation can be subdivided in eight compartments lying side by side with vertical boundaries. The compartments protrude in the Castigaleu Formation. Within the compartments, conglomeratic sheets lie at continuous levels, with thick fine members in between and along the vertical compartment boundaries. Levels are also indicated by basal transgressive mudstone and extra thick caliche layers. The base and top of the Formations show sandstone concentrations.

The coarse member sheets wedge or split up into several lenses distally. The palaeocurrent directions shift simultaneously from the SW to the NW. The number of coarse members lenses increases in the distal area and even more over the boundary with the Castigaleu Formation. Southerly-supplied sediment occurs in lenses and in a single sheet at the southern basin margin. A mudstone zone separates this area from the northeasterly-supplied sediments. The southern supply increased in a major way at the top of the Formation.
Concentrations of the coarse member occur at several fault sites; lenses, wedges and sheet thickenings are found and the facies can differ largely at either side of the fault. In general couples of coarse member lenses can show megasequences.

The normal faults are steeply inclined with throws of a few metres to a few tens of metres. The directional pattern of faults and joints has the same three major components as the palaeocurrent distribution.

Interpretation; The Pyrenean detritus was distributed over the area by eight parallel sheetfloods systems emerging from eight feeder canyons. The systems remained fixed at the same locations. Relatively highest energy flows deposited flat, massive gravel layers. Decelerated flows turned from SW through W to NW directions, depositing sand and gravel and flat layers of silt and mud.

Major distribution of gravel and sand occurred only during acceleration events of the large-scale tectonic processes. A relatively large amount of fines was distributed at the onset of the events. Transgression in the subsided basin, leaving mud, subsequently preceded major deposition of gravel and sand, eroded from the risen hinterland.

Gradual increases and decreases of interval thicknesses in between levels, simultaneously with regression and transgression respectively, reflect regional variation of the balance between basin and hinterland positions. The same goes for the major regressions at the base and the top of the Formations. These regional processes fell outside the scope of this study.

Development of coarse member megasequences in between levels in the centre of the basin can point to relatively highest subsidence rates there. However during acceleration of tectonic events also a wide zone along the northern basin margin received a thick gravel cover. The latter points to relatively constant subsidence rates. On a smaller scale megasequences developed in depressions over deep-seated faults. At these sites isolated channels and small fan-deltas deposited sediments. The sedimentation pattern could be changed on a somewhat larger scale by differential fault block motions. Tilting of fault blocks determined the gradients of the basin floor and thereby the current directions. The fault pattern corresponds with the directions of major faults in the Pyrenees.

The immaturity of the sediment grains (largely carbonatic rock fragments) points to a minor distance of transport (which is in line with deposition by a sheetflood system).

VI.1.4. Chapter 5 This chapter gives an inventory of the sections taken at regular distances in the northwestern and western area, supplemented by one southern section. The facies type is generally constant within the individual, isolated layers.

Monllobat Formation: Facies types of the coarse member; 1 Kilometre-wide sheets (massive conglomerate). 2 Relatively small sheet terminations (conglomeratic sandstone). 3 Winged, symmetrical channel fills (conglomerate or sandstone). 4 Foreset lobes (conglomerate or sand-/siltstone). 5 Meander lobes (sand-/siltstone). 6 Decimetre-thick sheets (coarse-grained sandstone). Facies types of the fine member: Decimetre- to metre-scale flat layering shows a regular alternation of mudstone and caliche beds. The latter are somewhat coarser (silt) and can contain type 3 coarse members. Vertical mottling is constant within the brown- or ochre-coloured layers. Blue and red mottles can be metre-long. A clayey yellow fine member is locally present, but contains only minor amounts of (gypsum) caliche.

The Monllobat sheetflood systems dumped gravel in flat layers. Distally decelerated flows became deflected and deposited a sandy off-shoot. Silt and mud layers were deposited by sheetfloods outside the bed. Erosional, symmetrical channels could be incorporated in the sheetfloods. Diverse types of foreset lobes could be produced during progradation of the sheetflood system in the flood basin. Meandering rivers were very scarce features (along tectonic depressions) in the terminal sheetflood system parts. Sheetflooding with coarse sand occurred occasionally in the distal part of the systems.

Because of episodic sedimentation, relatively short periods of sedimentation alternated with long periods of soil formation. Caliche in 50 % of the fine member layers points to a marked evaporation surplus. The metre-long vertical mottles indicate low ground-water tables with oxidation of the soils. The type of ornamentation of scarce (horizontal) burrows is proof of digging in dried mud (in the vicinity of lacustrine and (brackish-) marine environments. Oncoids were locally abundant in depressions above the deep-seated faults.

A climatic reconstruction comprises warm, rather dry conditions with episodic rainstorms triggering vast sheetfloods.

Castigaleu Formation: Facies types of the coarse member: 1 - 4 Low- to high-angled foreset lobes of conglomerate and of sandstone. 5 Flat conglomerate layers. 6 Flat, tabular crossbedded sandstone layers. 7 Flat, parallel-laminated sandstone layers. 8 Flat layers of massive sandstone.

Single oyster shells are often common in the coarse members. Locally plant debris is abundant.

Facies type of the fine member: Bluegrey-coloured mudstone comprises 70 - 80 % of the total volume of this Formation.

The Castigaleu facies types originated during inflow of sheetfloods in shallow standing water. The same facies distribution as in a small fan-delta (ch. 3) has also been found on a kilometre-scale and outflow processes are presumed to have been the same. Gravel was preferentially dumped at the coast, building a proximal sheet or (at significant breaks of slope) interconnected lobes. High suspension flows travelled through the shallow coastal zone, depositing sand and leaving environments for oyster growth.

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