GEOLOGICA ULTRAIECTINA

Mededelingen van het
Instituut voor Aardwetenschappen der
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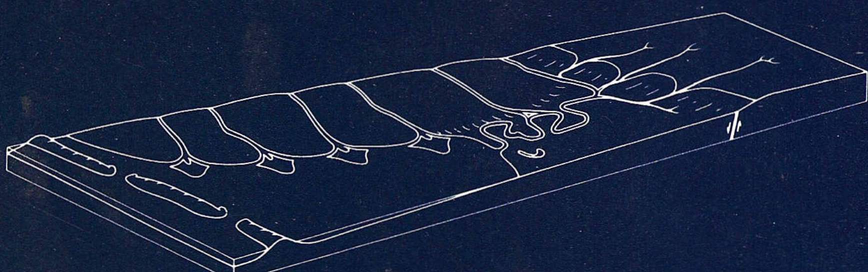
No.61

Sedimentary facies and sequential architecture of tide-influenced alluvial deposits.

An example from the middle Eocene Capella Formation, South-Central Pyrenees, Spain.



Margarita C. Cuevas Gozalo

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CIP-GEGEVENS KONINKLIJKE BIBLIOTHEEK, DEN HAAG

Cuevas Gozalo, Margarita Celia

Sedimentary facies and sequential architecture of
tide-influenced alluvial deposits : an example from the
middle Eocene Capella formation, South-Central Pyrenees,
Spain / Margarita Celia Cuevas Gozalo. - [Utrecht :
Instituut voor Aardwetenschappen der Rijksuniversiteit
Utrecht]. - (Geologica Ultraiectina, ISSN 0072-1026 ; no.
61)
Proefschrift Utrecht. - Met lit. opg. - Met samenvatting
in het Nederlands en Spaans.
ISBN 90-71577-14-7
SISO eu-span 565 UDC 551.3.051(234.12)(043.3)
Trefw.: sedimentologie ; Pyreneeën.

Sedimentary facies and sequential architecture of tide-influenced alluvial deposits

An example from the middle Eocene Capella Formation, South-Central Pyrenees, Spain.

Sedimentaire facies en sequentionele architectuur van getijde-beïnvloede alluviale afzettingen.

Een voorbeeld uit de midden-Eocene Capella Formatie, zuid-centrale Pyreneeën, Spanje

(met een samenvatting in het Nederlands)

Facies sedimentarias y secuencia arquitectónica de déposítos aluviales influidos por mareas.

Un ejemplo de la Formación Capella (Eoceno medio), Pirineos centro-septentrionales, España.

(con un resumen en español)

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 volgens besluit van het College van Dekanen in het openbaar te verdedigen op Woensdag 6 September 1989, des namiddags te 14.30 uur.

DOOR

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geboren op 21 Januari 1952 te Madrid, Spanje

PROMOTOR: PROF. DR. S.D. NIO

COPROMOTOR: DR. M. DIAZ MOLINA

A mis padres

Ecrire est un choix perpétuel entre mille expressions,
dont aucune ne me satisfait,
dont aucune surtout ne me satisfait sans les autres.

Marguerite Yourcenar
Alexis ou le Traité du Vain Combat (1929)

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SUMMARY

The sediments investigated consist of a thick sequence of clastic deposits of middle Eocene age, the Capella Formation.

At the time of deposition the sedimentary basin was tectonically active. Tectonic influence in the sedimentary sequence is recognized from angular unconformities, synsedimentary faults and vertical as well as lateral variations in the sedimentary facies and thickness distribution.

The most conspicuous structural feature is the synsedimentary Lascuarre fault system, which largely controlled the sedimentation of the Capella Formation.

Three unconformity-bounded basin infilling units or tectonosedimentary units are recognized. The boundaries of the tectonosedimentary units do not coincide with those of previously defined lithostratigraphic formations.

Sedimentation took place in and near a shallow bay, which was submitted to intertidal conditions. Sediment was supplied either by one or two alluvial depositional systems. Coeval depositional systems are distinguished by their facies association and paleocurrent direction.

Different process-related types of sediment bodies are recognized on the basis of the external geometry and internal organization: channel deposits, depositional lobes, sheet deposits and composite forms. A further differentiation of these groups is based on similar criteria.

These types are incorporated into the architectural analysis (i.e., mainly an evaluation in time and space). The interaction between sediment rate, tectonics and sea level fluctuation is discussed based on the architectural analysis.

As a result of the variations in the ratio between the rate of relative sea level rise and the rate of sedimentation, repeated oscillations in the depth of the basin occurred. Within the youngest tectonosedimentary unit stationary episodes, regressive episodes, tectonic-related deepening episodes and transgressive episodes are recognized. The overall trend is transgressive.

SAMENVATTING

De onderzochte gesteenten bestaan uit een dikke opeenvolging klastische afzettingen van midden-Eocene ouderdom, de Capella Formatie.

Ten tijde van afzetting was het sedimentaire bekken tektonisch actief. De invloed van de tektoniek op de sedimentatie uitte zich in het voorkomen van hoekdiscordanties, synsedimentaire breuken en verticale en laterale verschillen in sedimentaire facies en dikte.

Als stratigrafische onderverdeling worden drie tektono-sedimentaire eenheden onderscheiden, elk begrensd door discordanties. Deze begrenzingen komen niet overeen met oudere, op facies-criteria bepaalde formatiegrenzen.

Afzetting van de sedimenten van de Capella Formatie vond plaats in en aangrenzend aan een ondiepe baai, waarin intertidal condities heersten. Het sediment in de baai werd aangevoerd door twee alluviale systemen. De aanvoersystemen worden onderscheiden op grond van verschillen in facies-associatie en aanvoerrichting.

De verschillende sedimentaire processen produceerden verschillende soorten sedimentlichamen. Op grond van uitwendige geometrie en inwendige opbouw worden vier hoofdsoorten onderscheiden: geulafzettingen, lobvormige lichamen, dekenvormige afzettingen, en samengestelde vormen. Een verdere onderverdeling van deze vier soorten geschiedt op overeenkomstige criteria.

Analyse van de architectuur van de Capella Formatie, d.w.z. van de ontwikkeling in ruimte en tijd, geschiedt op basis van deze onderverdeling in soorten sedimentlichamen. De analyse van de architectonische opbouw wordt gebruikt om de interactie tussen afzettingssnelheid, tektoniek en zeespiegelfluctuaties te bediscussieren.

Ten gevolge van variaties in de verhouding tussen de snelheid van relatieve zeespiegelstijging en sedimentatiesnelheid ontstonden er bij herhaling oscillaties in de diepte van het bekken.

In de jongste tektono-sedimentaire eenheid worden zo stationaire episodes, regressieve episodes, episodes van verdieping als gevolg van tektoniek, en transgressieve episodes onderscheiden. In het algemeen genomen is de trend transgressief.

RESUMEN

Los sedimentos estudiados constituyen una potente sucesión de depósitos detríticos de edad Eoceno medio, la Formación Capella.

Durante su sedimentación la cuenca era tectónicamente activa. La influencia de la tectónica en la secuencia sedimentaria se reconoce por la presencia de discordancias angulares, fallas sinsedimentarias y por las variaciones verticales y laterales en la distribución de facies y espesor de sedimentos.

La estructura tectónica mas destacable es el sistema de fallas de Lascuarre, que controló en gran parte la sedimentación de la Formación Capella.

Se reconocen tres unidades tectonosedimentarias ó unidades de relleno limitadas por discordancias. Los límites de estas unidades no coinciden con los de las formaciones litoestratigráficas definidas por otros autores.

La sedimentación tuvo lugar en, y en las proximidades de, una bahía somera afectada por las mareas. El aporte de sedimentos a la cuenca se realizó, según el momento, por medio de uno ó dos sistemas deposicionales aluviales. La diferenciación entre sistemas deposicionales contemporáneos se basa en las asociaciones de facies y en las direcciones de paleocorrientes.

En base a la geometría y a la organización interna de los cuerpos sedimentarios se reconocen cuatro tipos principales: depósitos de canal, lóbulos deposicionales, cuerpos con geometría de capa y formas compuestas de canal y lóbulo. Estos tipos principales se han subdividido en base a los mismos criterios.

Los diferentes tipos de cuerpos sedimentarios se han incorporado en el análisis arquitéctonico, es decir en una evaluación temporal y espacial. En base al análisis arquitectónico se discute la interacción entre las tasas de sedimentación, de movimiento tectónico y de fluctuación del nivel del mar.

Como consecuencia de las variaciones en la relación tasa de subida relativa del nivel del mar / tasa de sedimentación, la profundidad de la cuenca experimentó repetidas oscilaciones. Dentro de la unidad tectonosedimentaria mas joven se reconocen episodios estacionarios, episodios regresivos, episodios de profundización relacionados con tectónica y episodios transgresivos. La tendencia general es transgresiva.

CHAPTER I: INTRODUCTION

1.- AIMS OF THE STUDY.

1.1.- Introduction

The Capella Formation (Middle Eocene) is a detritic sedimentary unit, which forms part of the Paleogene sediment fill of the Tremp-Graus Basin, in the Southern Pyrenees, Spain. The Capella Formation was defined by Garrido Megías (1968).

The Capella Formation has been studied in its type area, around the village of Capella in the Province of Huesca. The study has been limited to the Isábena and Esera valleys (fig. 1.1), where the best outcrop conditions are present. The lithofacies occurrences are most typical in the studied area and the sediment thickness is expected to be the largest.

The present study of the Capella Formation introduces a new approach to the sedimentary facies analysis in relation to the development of the Tremp-Graus Basin.

Current research in basin analysis focuses on four main aspects: basin formation, nature of the basin fill, diagenesis of the basin fill and timing of events (Klein, 1987).

The present study focuses on the analysis of the basin fill, which also contains information on the tectonic and sedimentary events which affected the basin. The tectonosedimentary analysis, the sedimentary facies analysis and the reconstruction of the large-scale architecture (i.e. mainly a geometrical evaluation in time and space) have been used to determine the internal organization and the nature of a part of the basin fill. Biostratigraphic data have been used to correlate events.

The identification and correlation of genetically related rock units or basin infilling units is considered in the analysis of the basin fill. The different scales at which basin infilling units can be recognized, as well as the different methods currently used for their determination and interpretation, are discussed in the following sections of chapter I.

Much has been written on the geodynamic origin and structural characteristics of the Tremp-Graus Basin in the last twenty years and it is summarized ably by Cámara and Klimowitz (1985) and Farrel et al. (1987). The most important structural aspects of the basin are mentioned in the Geological Setting (part 2 of chapter I). Chapter II details on the stratigraphy of the Capella Formation. The litho- and biostratigraphic position of the Capella Formation in relation to other formations as considered in previous works is discussed first. Then new stratigraphic units, as distinguished by the tectonosedimentary analysis (see definition in part 1.2 of this chapter), are proposed. A sedimentological approach is considered in chapter III: different process-related types of sediment bodies are recognized by the facies analysis. These types are incorporated into the architectural analysis. The interaction between sediment rate, tectonics and sea-level fluctuations is discussed based on the architectural analysis. In chapter IV a correlation is proposed between the tectonosedimentary units in the study area and other areas in the basin. Finally, in the same chapter, the evolution of the studied part of the Tremp-Graus Basin is discussed in terms of lateral subsidence variations in relation to large synsedimentary faults, as inferred from the facies, geometry and thickness of the different stratigraphic units of the Capella Formation.

1.2.- Basin sediment fill

The analysis of basin sediment fills concerns the identification and correlation of basin infilling units. These units can be recognized at various scales within a basin. In inverse hierarchical order they are:

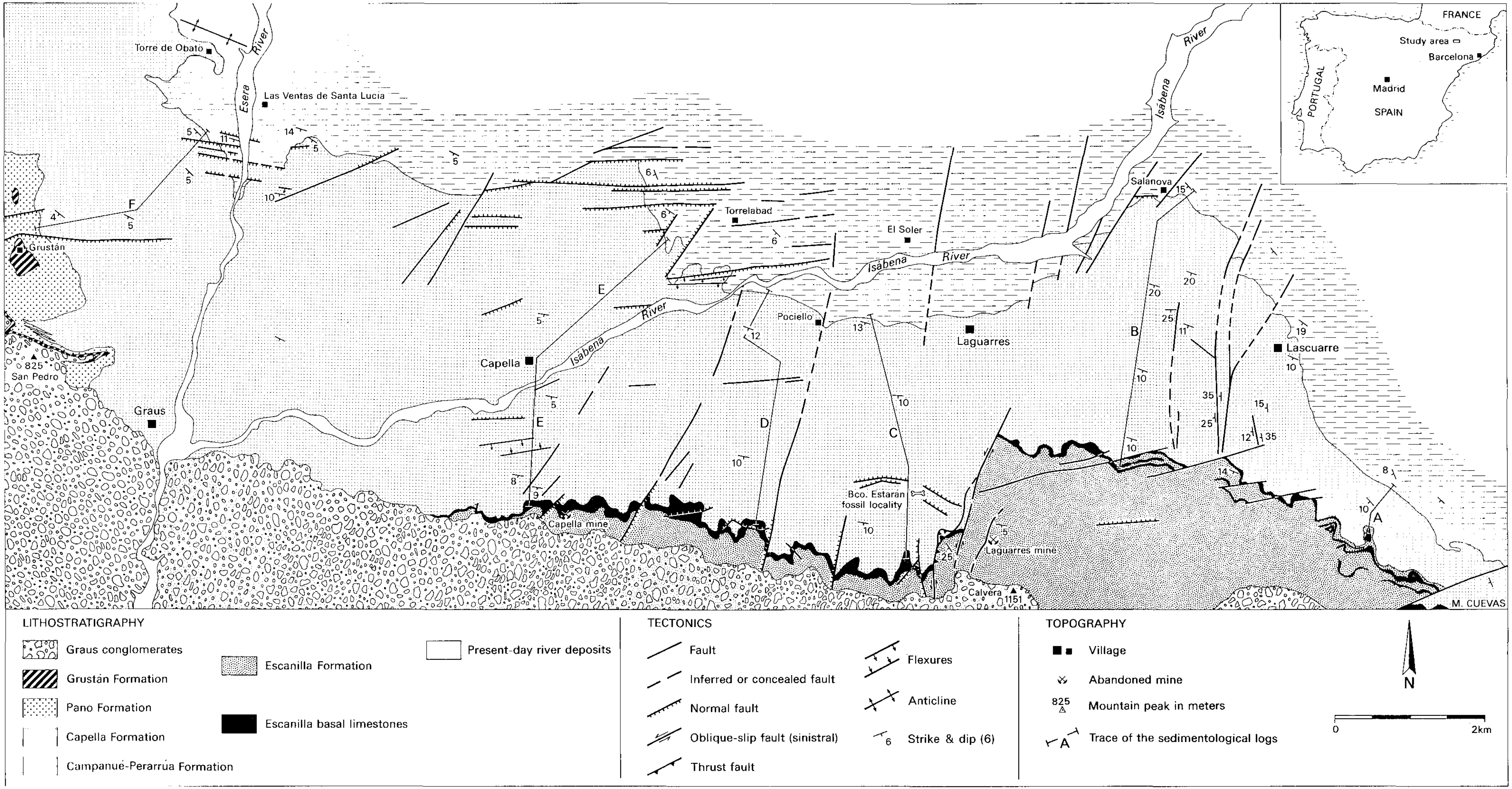
1- Depositional systems
The depositional systems are units of very variable scale. The units are defined by facies criteria and designated by genetic terms, i.e. deltaic system (Fischer and McGowen, 1969).

2- Depositional sequences
A depositional sequence is a stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities (Vail et al., 1977).

3- Tectonosedimentary units
A tectonosedimentary unit (TSU) is a stratigraphic unit consisting of strata (not necessarily conformable), which were deposited during a given geological time interval, under sedimentary- and tectonic- dynamic conditions characterized by a particular trend. The limits of any TSU are sedimentary discontinuities or ruptures of basin-wide extent or their correlative conformities (Garrido Megías, 1973; Megías, 1982). Tectonosedimentary units are recognized by means of tectonosedimentary analysis. This is a basin analysis method which allows the simultaneous development of a spatial and a time-related analysis of the sedimentary infilling (Megías, 1982).

Both depositional sequences and tectonosedimentary units are stratigraphic units with chronostratigraphic significance. The differences between depositional sequences and tectonosedimentary units are discussed in detail in the next section.

Fig. 1.1.- *Geological map of the study area. The position of the sedimentological logs and the vertebrate fossil localities are marked.*



Torre de Obato
Esera River
Las Ventas de Santa Lucia
Isábena River
FRANCE
Study area
Barcelona
PORTUGAL
Madrid
SPAIN
Salanova
Torrelabad
El Soler
Grustán
Pociello
Laguarres
Lascuarre
825
San Pedro
Capella
Graus
Bco. Estarán fossil locality
Capella mine
Laguarres mine
Calvera 1151
M. CUEVAS
A
B
C
D
E
F
LITHOSTRATIGRAPHY
Graus conglomerates
Grustán Formation
Pano Formation
Capella Formation
Campanué-Perarrúa Formation
Escanilla Formation
Escanilla basal limestones
Present-day river deposits
TECTONICS
Fault
Inferred or concealed fault
Normal fault
Oblique-slip fault (sinistral)
Thrust fault
Flexures
Anticline
Strike & dip (6)
TOPOGRAPHY
Village
Abandoned mine
Mountain peak in meters
Trace of the sedimentological logs
N
0
2km

4- Tectonosedimentary cycles

A tectonosedimentary cycle is a group of tectonosedimentary units. Sedimentary discontinuities separating these cycles are of higher rank than those bounding the tectonosedimentary units and they are related to large paleogeographic changes on a basin-wide scale (Megías, 1982).

5- Tectonosedimentary periods

A tectonosedimentary period is a group of tectonosedimentary cycles (Megías, 1982). The tectonosedimentary periods are comparable in scale to the stratigraphic sequences of Sloss (1963), to the system groups of Wang Hung (1966) and to the synthems of Chang (1975). These are rock units of the largest magnitude, bounded by unconformities and representing the deposits of a major geotectonic cycle. Stratigraphic sequences and synthems are defined in cratonic areas.

1.3.- Depositional sequences and tectonosedimentary units

There are differences in the basic concepts used to define the depositional sequences of Vail et al. (1977), Vail (1987) and Van Wagoner et al. (1987) and the tectonosedimentary units of Megías (1982), and in the interpretation of the causes which controlled their formation.

Main differences in the criteria for recognizing depositional sequences and tectonosedimentary units

Depositional sequences are bounded by purely geometrical discordances: erosional truncation surfaces, onlap surfaces and downlap surfaces (Vail et al., 1977). Each of these discordances forms a sequence boundary and resulted from a relative sea-level fall and its subsequent rise.

The criteria indicated by Megías (1982) for recognizing tectonosedimentary unit boundaries are of several kinds:

* *geometrical:*
  - erosive unconformities,
  - onlap and downlap over the same surface,
* *lithological:*
  - in marine sediments, occurrence of facies indicating subaerial exposure (caliche) or no sedimentation (hardgrounds),
* *sequential:*
  - change in the vertical character of the megasequence from transgressive to regressive or vice versa,
  - change in the horizontal character of the megasequence from restrictive (offlapping, toplapping) to extensive (onlapping) or vice versa.

In some of the examples given by Megías (1982) the downlap-surface, which according to Vail et al. (1977) forms a sequence boundary, could not result from a relative fall of sea-level; on the contrary, it represents either a change in the ratio between the rate of sedimentation and the rate of relative sea-level rise or the superposition of two synchronous depositional systems with different progradation capacities (figs. 3 and 6 of Megías, 1982).

Van Wagoner et al. (1987) discarded the downlap surfaces as sequence boundaries. The depositional sequences are subdivided into system tracts. The system-tract boundaries are marked by downlap-surfaces; the downlap-surfaces indicate either the transition from restrictive conditions to extensive conditions (transgressive surface) or from transgressive conditions to regressive conditions (maximum flooding surface). Van Wagoner et al. (1987) included lithofacies criteria, i.e. the subaerial exposure, as the criteria for a sequence boundary and relative sea-level fall when there is no erosional truncation of continental deposits.

The criteria used for defining the system-tract boundaries coincide reasonably well with those used to define the tectonosedimentary unit boundaries; each system tract of Van Wagoner et al. (1987) coincides with a tectonosedimentary unit of Megías (1982), with one exception (fig. 6 of Megías, 1982).

However, the major difference between the sequence stratigraphy analysis of Vail et al. (1977), Vail (1987) and Van Wagoner et al. (1987) and the tectonosedimentary stratigraphy of Megías (1982) lies in the interpretation.

Differences in the interpretation of the sequence stratigraphy versus the tectonosedimentary stratigraphy

Sediment supply

Variations in the rate of sediment supply are accepted in theory by Vail et al. (1977) as a factor controlling the formation of either transgressive or regressive sequences and the facies distribution. However, Van Wagoner et al. (1987) interpreted the transition from the transgressive system tract to the highstand system tract to result from a change of eustatic origin in the rate of relative sea-level rise.

Megías (1982) considered that a change in the rate of sediment supply could produce such a change in the character of the sequence, even when the rate of relative sea-level rise remains stable.

Tectonics

Vail (1987) considered that tectonic subsidence is the main factor controlling the sediment thickness. However, the facies distribution as well as the unconformities and the rapid falls of relative-sea-level are of eustatic origin.

Megías (1982) considered differential tectonics as an important mechanism in the formation of unconformities and relative sea-level changes. Tectonics may play a double role in the basin: a) in the form of tectonic subsidence of the depositional trough, producing accommodation for the sediments and a relative rise of sea-level, and b) in the form of tectonic uplifting, resulting in the shallowing of the basin margins, their subaerial exposure and subsequent erosional truncation. Erosion favours an increase of sediment supply, which is more likely to produce an increase in the rate of sedimentation (fig. 1.2). Both subsidence

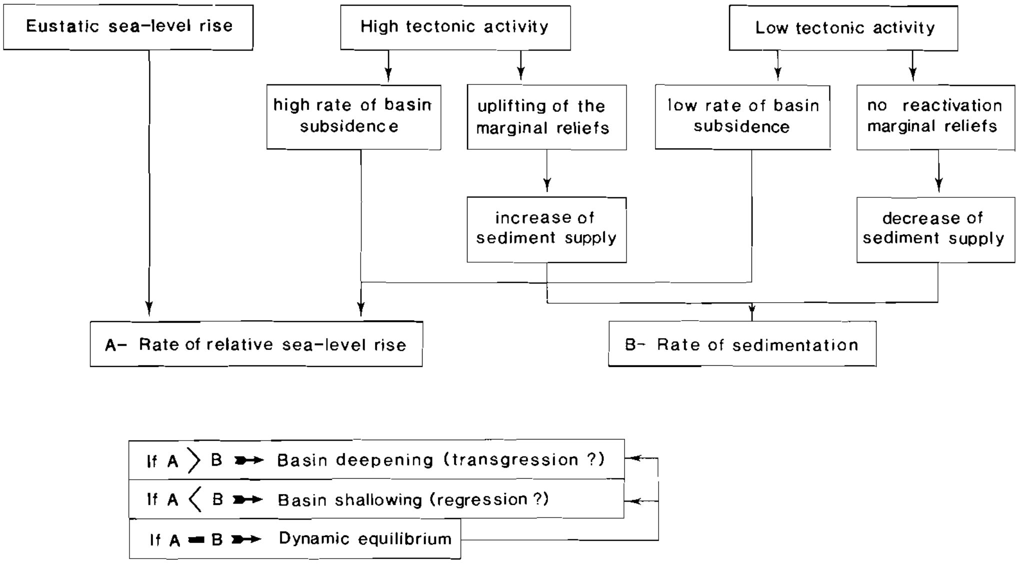


Fig. 1.2.- *Factors controlling the evolution of the basin during conditions of relative sea-level rise.*

and uplifting can occur simultaneously in different parts of the basin.

1.4.- Method of analysis and interpretation

In the present study several basin infilling units are distinguished based on the recognition of angular and erosive unconformities and their correlative conformities.

Erosive unconformities as a criterium for the boundaries between basin infilling units is considered in both the seismic stratigraphy analysis of Vail et al. (1977) and the tectonosedimentary analysis of Megías (1982). From this point of view, the basin infilling units recognized in the present study could be considered either as depositional units or tectonosedimentary units.

However, the conceptual model of causes used by the tectonosedimentary analysis for the interpretation has been preferred in the present study. Two main points have supported this choice:

1) The Tremp-Graus Basin was a tectonically active basin during the time of deposition of the sediments considered here. Changes in the intensity of the tectonic deformation - both in the form of uplifting of the basin marginal reliefs and of tectonic basin subsidence - should be considered as a factor controlling the nature of the basin fill. In the tectonosedimentary analysis

tectonics is contemplated not only as the main variable determining the accommodation of the sediments, but also as a factor indirectly controlling the facies distribution.

2) The deposits considered in the present study are terrigenous sediments which were deposited in transitional environments from alluvial to shallow marine. The accumulation and facies distribution of sediments in such environments is largely influenced both by the changes of relative sea-level and by the variations in the sediment supply (fig. 1.2).

   The sequence stratigraphy analysis of Vail et al. (1977), Vail (1987) and Van Wagoner et al. (1987) focus on the analysis of passive platform margins. Mainly shelf and deep-sea sedimentary successions are considered.

   The tectonosedimentary analysis of Megías (1982) can be applied to both continental or marine sequences. Moreover, changes in the rate of sediment supply are considered as one of the main factors controlling the evolution of the sedimentary succession.

2.- GEOLOGICAL AND GEOGRAPHICAL SETTING OF THE TREMP-GRAUS BASIN, SOUTHERN PYRENEES

The Pyrenees form an almost linear mountain chain, extending for approximately 1000 km from Cantabria in northern Spain eastwards to Provence in southern France (Williams, 1985). According to Grimaud et al. (1982) the Pyrenees continue to the west along the entire Iberian margin, to connect with the Azores-Biscay Rise and King's Trough lineament.

The evolution of the Pyrenean mountains as part of the Alpine orogeny is closely related to the movement of the African and Iberian plates relative to Eurasia. In terms of plate tectonics the Pyrenean mountains are situated between the Iberian plate and the European plate. The separation of the Iberian plate from North America started in the Early Cretaceous (Tapponier, 1977). The Iberian plate moved in an anti-clockwise sense around a shifting pole (Choukroune et al. 1973 a, b). Contact between the Iberian plate and the European plate occurred along the North Pyrenean Fault. During the Cretaceous, movement along this fault was sinistral (Le Pichon et al., 1970).

The change of movement of the African plate with respect to Eurasia, from a NE direction (Jurassic-Late Cretaceous) to a N direction (Eocene) forced the Iberian plate against the Eurasian plate (Tapponier, 1977). Between Latest Cretaceous and Late Eocene the Iberian plate moved northwestwards (Grimaud et al., 1982). The tectonic configuration of the Pyrenean orogenic belt was generated by a NNW-SSE compression, probably due to the relative dextral movement between the Iberian and European plates (Cámara and Klimowitz, 1985).

As a consequence of the collision significant crustal shortening took place (Choukroune et al., 1973 b). On the Iberian plate the frontal part (axial zone) was flanked to the south by the South Pyrenean zone (Seguret, 1972; fig. 1.3).

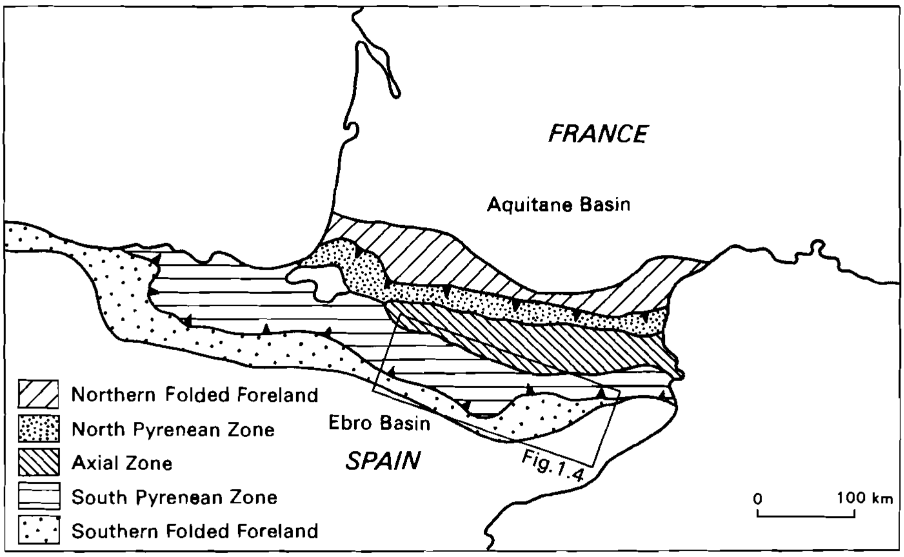


Fig. 1.3.- *Map showing the main structural divisions of the Pyrenees (after Choukroune and Seguret, 1973).*

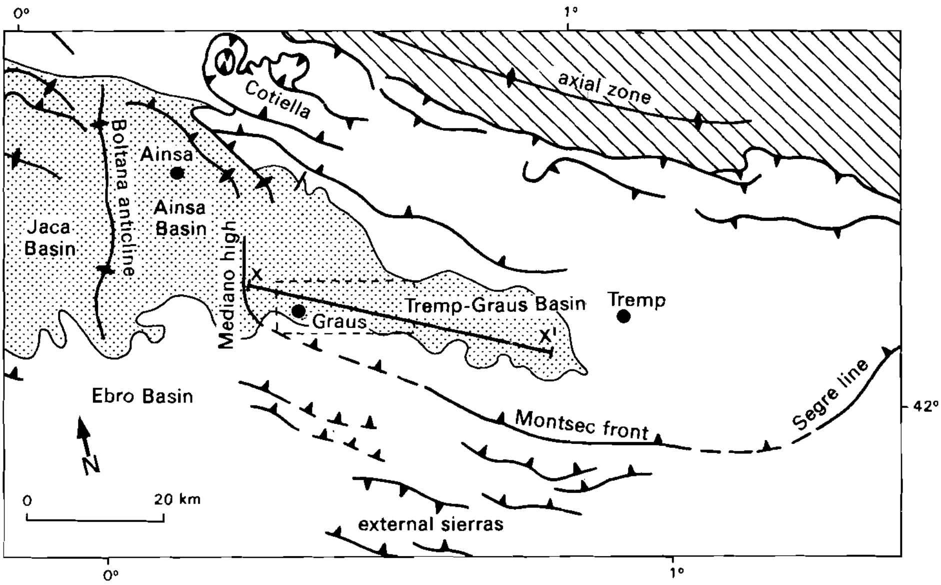


Fig. 1.4.- *Schematic geological map of the South Central Pyrenees to show the major structures and the Tremp-Graus and Ainsa compartments of the South Pyrenean foreland basin (modified from Farrel et al., 1987). The square indicates the study area in fig. 1.1. X-X': profile in fig. 4.1.*

The classical view of the South Pyrenean zone is that of several allochthonous units consisting of Mesozoic and Tertiary strata, which displaced southwards by gravity gliding on a décollement horizon of Triassic evaporites (Solé Sugrañes, 1978). Williams (1985) argued against "gravity gliding" as the main mechanism for allochthony and proposed a piggy-back thrust model for the South Central Pyrenees. A model of Eocene-Oligocene imbricated thrusts was also proposed by Cámara and Klimowitz (1985) for the central to western areas of the Southern Pyrenees.

Several sedimentary basins developed upon and in front of the thrust sheets in the Southern Pyrenees. The Eocene infilling of the basins was coeval with the southwards displacement of the thrust sheets (Williams, 1985; Cámara and Klimowitz, 1985).

The Tremp-Graus Basin is an ESE-WNW trending synclinal depression which extends across the provinces of Lérida and Huesca in northern Spain. As limited by Nijman and Nio (1975) the Tremp-Graus Basin extends over a distance of 95 km, with a width of 20 km from the village of Isona, 20 km east of Tremp, northwestwards to the Boltaña High. Actually these boundaries include two basins within the South Pyrenean Zone: a) the Tremp-Graus Basin sensu stricto, from Isona to the Mediano High, west of Graus, and b) the Ainsa Basin, between the Mediano High and the Boltaña High (fig. 1.4). In the present study the name Tremp-Graus Basin is used only for the Tremp-Graus Basin sensu stricto.

The Tremp-Graus Basin developed on the Montsech-Cotiella nappe of the "Central Southern Pyrenean unit" of Seguret (1972), also named the Montsech unit by Garrido Megías (1973) or Montsech thrust-sheet by Williams (1985). The Ainsa Basin developed on the more western allochthonous "Gavarnie unit" (Seguret, 1972). Farrel et al. (1987) considered both the Tremp-Graus Basin and the Ainsa Basin as thrust-associated foreland basins, bounded by the north-south trending Mediano and Boltaña ramp anticlines. Besides the above mentioned ramp anticlines, other thrust-related features are indicated by Cámara and Klimowitz (1985) within the Tremp-Ainsa Basin based on their seismic interpretations.

During the Middle Eocene the Tremp-Graus Basin was occupied by a fluvio-deltaic complex (Van Eden, 1970) which fed a submarine fan complex in the Ainsa Basin (Mutti et al., 1972). Nijman and Nio (1975) recognized two major groups or fluvio-deltaic systems within the Eocene clastic infilling of the Tremp-Ainsa Basin, the Montañana Group and the Campodarbe Group. The Montañana Group was restricted to the eastern part of the basin, i.e. the Tremp-Graus Basin s.s.. At the top of the Montañana Group, distal facies of the eastern Montañana delta interfered with the northern conglomeratic Campanúe fan delta to form the shallow marine Perarrúa Formation. According to Nijman and Nio (1975), the subsequent westward progradation of the Montañana delta produced the gradational superposition of the alluvial Capella Formation over the Perarrúa Formation. The Campodarbe Group extended further to the west, unconformably overlapping the Mediano and Boltaña Highs. The Escanilla Formation (Garrido Megías, 1968) comprises the continental facies of this group.

CHAPTER II: STRATIGRAPHY OF THE CAPELLA FORMATION IN THE ISABENA AND ESERA VALLEYS

1.- INTRODUCTION

Stratigraphical studies on the Eocene deposits of the Tremp-Graus Basin in general and on the deposits of the Capella Formation in particular date back to the mid-1950s (fig. 2.1). The main emphasis of research during the following fifteen years was to establish the biostratigraphic and lithostratigraphic framework of the basin (see fig. 2.1 and later figs. 2.3 and 2.7). Most of the lithostratigraphic formations in present use were defined during this period (Garrido Megías, 1968; fig. 2.3).

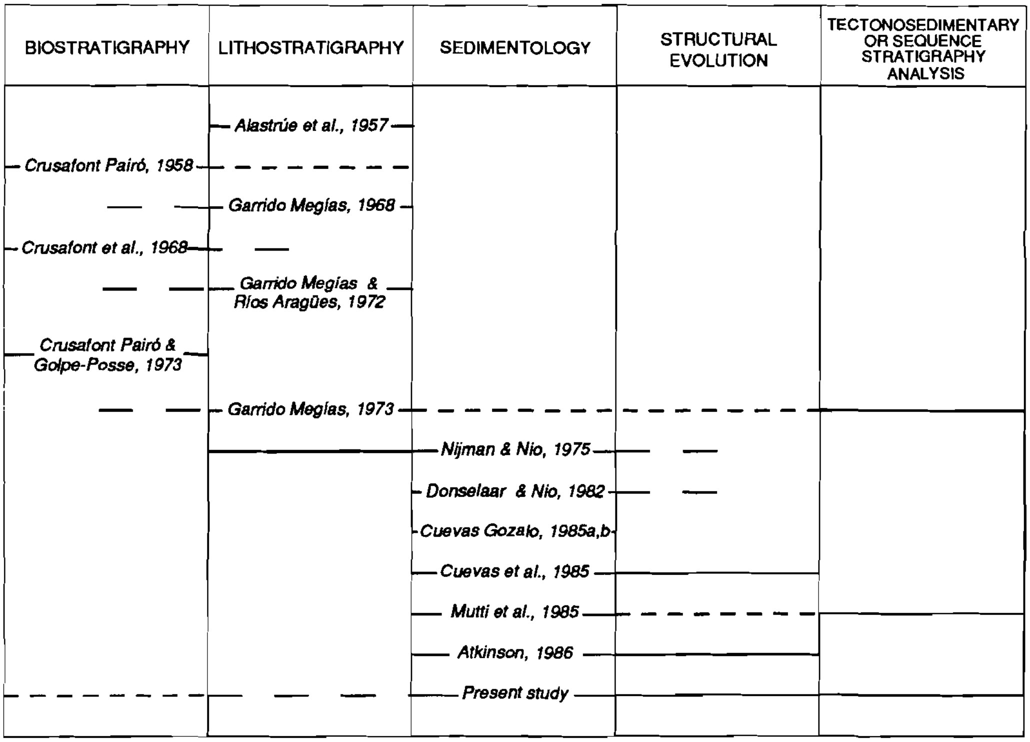


Fig. 2.1.- *Historical review of studies concerning the deposits of the Capella Formation.*

Garrido Megías (1973) used the structural evolution of the basin as the most important factor controlling sedimentation in the Tremp-Graus Basin. He defined specific sedimentary successions, "rhythms", which were bounded either by unconformities or by changes in the general sedimentary sequence. Consequently, the boundaries of the lithostratigraphic formations do not coincide with those of the tectonosedimentary "rhythms" (see later fig. 2.3). The "rhythms" were later renamed as tectonosedimentary units (Megías, 1982).

A new aspect of research was introduced in the 1970s by the Dutch school, which carried out detailed sedimentological studies (Nijman and Nio, 1975). They postulated a basin model with diachronous lithostratigraphic formations (fig. 2.3). Later sedimentological studies have consisted of thematic subjects (e.g. Donselaar and Nio, 1982; Cuevas Gozalo, 1985 a,b). More recent studies try to integrate the spatial facies development with the structural evolution of the basin (Cuevas et al., 1985; Atkinson, 1986). Sequence-stratigraphy studies were introduced by Mutti et al. (1985).

2.- LITHOLOGICAL AND STRATIGRAPHIC FRAMEWORK

2.1.- The Capella Formation: Lithological characteristics

The Capella Formation was defined by Garrido Megías (1968) on the western bank of the Esera River, between Torre del Obato and Grustán (fig. 1.1). The formation consists here of a 460 m thick succession of grey sandstones interbedded in ochre and occasionally grey mudstones; in the Isábena valley the formation is up to 1070 m thick and consists of sandstones, conglomerates and some gypsum levels interbedded with multicoloured mudstones. East of the Isábena valley, a sharp decrease of the total sediment thickness is observed. The Capella Formation consists here of a 210 m thick interval of conglomerates and sandstones interbedded with ochre mudstones; a thin limestone level occurs locally.

Lithological characteristics and formation thickness in the Esera and Isábena valleys as inferred from the present study do not differ much from those indicated by Garrido Megías (1973).

Additional characteristics of the Capella Formation are the generally low sand/mud ratio and the consequent low interconnectedness rate of the sediment bodies, the scarcity of body-fossils, the vertical patterns of bioturbation and the mottling.

Mud deposits

Mud deposits form the bulk of the Capella Formation; they represent about 75 % of the total volume of sediments. The individual mudstone intervals reach thicknesses of more than 20 m. The mudstones are yellow-brownish and commonly mottled; violet to green mudstones occur in the transition with the underlying shallow-marine Perarrúa Formation; bluish-grey mudstones occur in the more seaward area in the western part of the study area and in the transition to the overlying Escanilla Formation. The body-fossil content is poor. The deposits are, however, intensely bioturbated; vertical cylindrical burrows are predominant and can be observed at the boundaries with coarser grained deposits.

Silt deposits occur interbedded in the mudstones. Siltstones can be distinguished from mudstones because of their better cementation. Cementation is related to a slightly coarser grain size. A better efficient-porosity and permeability enabled intrastratal fluid movements and enhanced cementation. The siltstones are highly bioturbated and have a mottled ochre-colour. Other sedimentary structures are rare. Siltstones are commonly laterally or vertically associated to coarser deposits.

Conglomerates

The conglomerates of the Capella Formation, with a few exceptions, contain 70 % or more of carbonate-rock fragments in the gravel fraction (fig. 2.2). Most of the carbonate fragments originate from the Mesozoic cover of the Central Pyrenees and also from the Paleogene deposits of the basin itself. The latter were uplifted and exposed to erosion by synsedimentary thrust structures. Fragments of grey calcareous mudstone (Upper Cenomanian-Turonian; Werver, pers. com.), laminated sandy limestone (Santonian-Maestrichtian) and alveoline limestone (late Paleocene-early Ypresian) are recognized. Subordinate components are black chert (Upper Cretaceous), quartzite, white micritic limestone, acid and basic igneous rocks and quartz. The white micritic limestone probably has its origin in Lutetian lacustrine limestones which infilled more eastern parts of the basin. The fragments of acid igneous rocks probably come from Carboniferous granodiorites in the central and eastern part of the Pyrenean Axial Zone. The fragments of basic igneous rocks probably come from Late Triassic dolerites in the central Pyrenees. The non-carbonate subordinate components occur usually as pebbles or small cobbles, while the dominant carbonate-rock fragments reach diameters up to 40 cm. Other subordinate components are of intraformational origin: terrigenous-mudstone pebbles and cobbles, pebble-sized calcareous nodules and oncolites. The intraformational components occur usually as lag deposits or in the foresets of sandstone bodies.

Sandstones

Most of the sandstones in the Capella Formation are medium to very fine grained. They are lithic arenites (c.p. Pettijohn et al., 1972); the main component is quartz; carbonate-rock fragments are abundant; feldspar can be present. The sandstones are generally grey, although fine grained sandstones present ochre-coloured mottling associated with bioturbation.

Among the coarse to pebbly sandstones two petrological types can be differentiated by field observations: a) detritic-carbonate rich sandstones (carbonatic litharenite) and b) feldspar-rich sandstones (subarkoses) (fig. 2.2).

	CONGLOMERATIC FRACTION (-2 ≥ Ø ≥ -9)	LITHIC ARENITES		CARBONATIC LITHARENITE (1 > Ø > -2)		SUBARKOSE (1 > Ø > -2)	
		PETTIJOHN ET AL. 1972	CAPELLA FM	PETTIJOHN ET AL. 1972	CAPELLA FM	PETTIJOHN ET AL. 1972	CAPELLA FM
QUARTZ		< 75 %	50 - 75 %	< 75 %	25 - 75 %	75 - 95 %	75 - 95 %
FD + LC		> 25 %	25 - 50 %	> 25 %	25 - 75 %	5 - 25 %	5 - 25 %
FD / LC RATIO		< 1	< 1	< 1	< 1	> 1	> 1
CARBONATIC LC	> 70 %			> 12,5 %	> 20 %		X
OTHERS (Ø < - 2):							
CHERT							X
QUARTZITE							X
WHITE QUARTZ	< 30 %						X
ACID IGNEOUS ROCKS							X
BASIC IGNEOUS ROCKS							(X)

Fig. 2.2.- *Field estimation of the petrological composition of the conglomerates and coarsest sandstones of the Capella Formation. FD: feldspar; LC: rock fragments; X: present; (X): rare.*

Carbonatic litharenites

The term carbonatic litharenites is used here for the lithic-arenites in which rock fragments are mainly limestone (fig. 2.2).

Carbonatic litharenites often contain small pebbles of grey limestone; transported calcareous nodules occur locally. Coarse and very coarse carbonatic litharenites are brown, poor in matrix and sometimes grain-supported.

Nummulites sp. of 2-3 mm diameter are also found in carbonatic litharenites. The nummulites are very dispersed in medium to coarse sandstones which also contain small pebbles and granules. The nummulites are not corroded very much, suggesting a short transport distance. The occurrence of nummulites is associated with stratigraphic intervals in which bipolar paleocurrent directions are frequently seen. On the one hand it may be inferred that the nummulites were transported landwards by tidal currents, not only from the west, but eventually also from the southwest and south. On the other hand, it has been observed that well-preserved nummulites occur as transported fragments from older rocks in Eocene conglomeratic deposits at the basin's northern margin (Cajigar Formation). The association in the Capella Formation of these fossils with the coarser carbonate detritic facies and their relative scarcity in the most seaward section, suggest that the nummulites were reworked from older intrabasinal Paleogene deposits. Under the influence of tectonic movements the basin experienced process of "cannibalism", i.e., older deposits of the same basin were eroded at the marginal areas and transported to the depositional part of the basin. Whether these nummulites are reworked from older formations in the basin or are transported from contemporary shallow marine areas is still under discussion.

Subarkosic sandstones

The term subarkose is used here in the sense of Pettijohn et al. (1972) (fig. 2.2). The subarkosic sandstones are characterized by a white or light grey colour in the field. The white colour is produced by weathering of the feldspar.

The subarkoses are commonly rich in small cobbles and pebbles of grey limestone, quartzite and occasionally of acid and basic igneous rocks. The rock fragment content indicates a mixed provenance.

Evaporites

Gypsum concentrations in the deposits of the Capella Formation are reported by Garrido Megías (1973) and Nijman and Nio (1975). Several gypsum habits are recognized, all occurring in red mudstones:

1) Massive gypsum. Massive gypsum crops out at a single locality, south of Salanova (section B in fig. 1.1). It occurs in the middle part of the Capella Formation and not in the transition with the underlying marine deposits as indicated by Nijman and Nio (1975) (fig. 2.6). The massive gypsum is vertical and laterally associated with conglomerates and sandstones.
2) Concretional or nodular gypsum. This consists of small crystal aggregates, which are dispersed in mudstones.
3) Fibrous gypsum veins. They are seams up to 6 cm thick, which arbitrarily cut the stratification planes. Although this gypsum habit is of secondary origin, its occurrence in particular stratigraphic intervals suggests that it was formed by diagenetic mobilization of primary gypsum concentrations in the host sediment.

Limestones

Thin micritic limestone beds (a few decimetres thick) occur locally in the western part of the study area. They are generally highly bioturbated and occur in the transition from the marine deposits of the Perarrúa Formation to the continental deposits of the Capella Formation. The limestones were probably formed by precipitation from lagoonal brackish-waters.

Higher in the sequence and south of Lascuarre (fig. 1.1), some patches of white micritic limestones are interbedded in the detritic deposits of the Capella Formation. The limestones, which are several decimetres thick, have an irregular, locally agglomeratic internal pattern. Both the red mottling of the underlying sandstones and the apparent "pedoturbation" of the limestones suggest the association of these limestones with pedogenetic processes. Probably, they were formed as palustrine limestones (as defined by Freytet (1964)) or desiccation breccias (as defined by Freytet and Plaziat (1982)).

Paleosols

Calcitic-carbonate nodules up to 2 cm of diameter are characteristic of the ochre mudstones at the base of the Capella Formation. The nodules are disorthic in that they possess sharp boundaries and can easily be removed from the enclosed matrix (c.p. Brever, 1964; Wieder and Yaalan, 1974 and Atkinson, 1986). In some cases the nodules occur in ochre-grey mudstones with violet vertical mottling, suggesting a calcitic gley soil (c.p. Freytet and Plaziat, 1982). In the most eastern part of the studied area the calcitic-carbonate nodules are agglomerated and form ruiniform horizons; these represent an extreme stage of nodule coalescence (c.p. Freytet, 1971 and Freytet and Plaziat, 1982). Calcitic carbonate nodules are also found as transported material in the erosively overlying sandstone deposits. The disorthic nodules and the ruiniform horizons which appear at the base of the Capella Formation in the study area, are similar to those described by Atkinson (1986) as occurring throughout the Formation in more eastern areas (Atkinson's locations C2 and C3).

Hydromorphic soils of gley and pseudo-gley type are described in the Isábena valley by Atkinson (1986, his location C1): these paleosols consist of cyclic arrangements of grey-yellow-orange horizons within the mud-siltstones. Their most diagnostic feature is the presence at many levels of well-developed colour mottling. According to Atkinson (1986) these soils developed in a flood plain setting where the water-table was either high or constantly fluctuating. Similar paleosols are recognized at several stratigraphic levels within the Capella Formation in the study area.

With the exception of the ruiniform calcitic horizon, the calcitic as well as the hydromorfic paleosols in the studied area occur within unconsolidated mudstones. Their morphological appearance is similar to that of the ubiquitous mud deposits. They do not form reference levels and can only be correlated when the general sequence is considered.

2.2.- The Capella Formation as a stratigraphic unit

In the Tremp-Graus Basin Garrido Megías (1968) and Nijman and Nio (1975) have frequently defined Eocene stratigraphic formations by lithofacies criteria instead of purely lithological ones. This makes the determination and mapping of the formation boundaries rather difficult. Facies may change repeatedly, both vertically and laterally, within a chronostratigraphic interval and diachronous depositional intervals may present similar facies. The diachronous character of the facies-related formation boundaries is obvious. Sedimentological analysis and reconstructions based on the comparison or correlation of parts of a formation should consider the diachronous character of facies.

The Capella Formation is an example of a facies-related stratigraphic formation. Both the lower and locally the upper boundary are defined by a change of facies from marine to continental facies and vice versa. This is the reason why in the literature the boundaries of the Capella Formation have been moved (fig. 2.3).

In the present study the formations are not used as basic units for the stratigraphic analysis. Instead, three tectonosedimentary units are recognized on the basis of unconformities and their correlative conformities (figs. 2.3 and 2.4). The unconformity-bounded sediment successions are units of chronostratigraphic significance for the interpretation of the geology (Vail et al., 1977, 1984; Megías, 1982; Vail, 1987). The unconformity-related boundaries of the tectono-sedimentary units do not coincide with the facies-related boundaries of the stratigraphic formations (figs. 2.3 and 2.5).

Capella Formation lower boundary

In the area studied the Capella Formation is underlain by the Perarrúa Formation (figs. 1.1 and 2.3). The latter consists of sandstones, mudstones and some conglomerates in shallow-marine facies (Garrido Megías, 1968; Nijman and Nio, 1975).

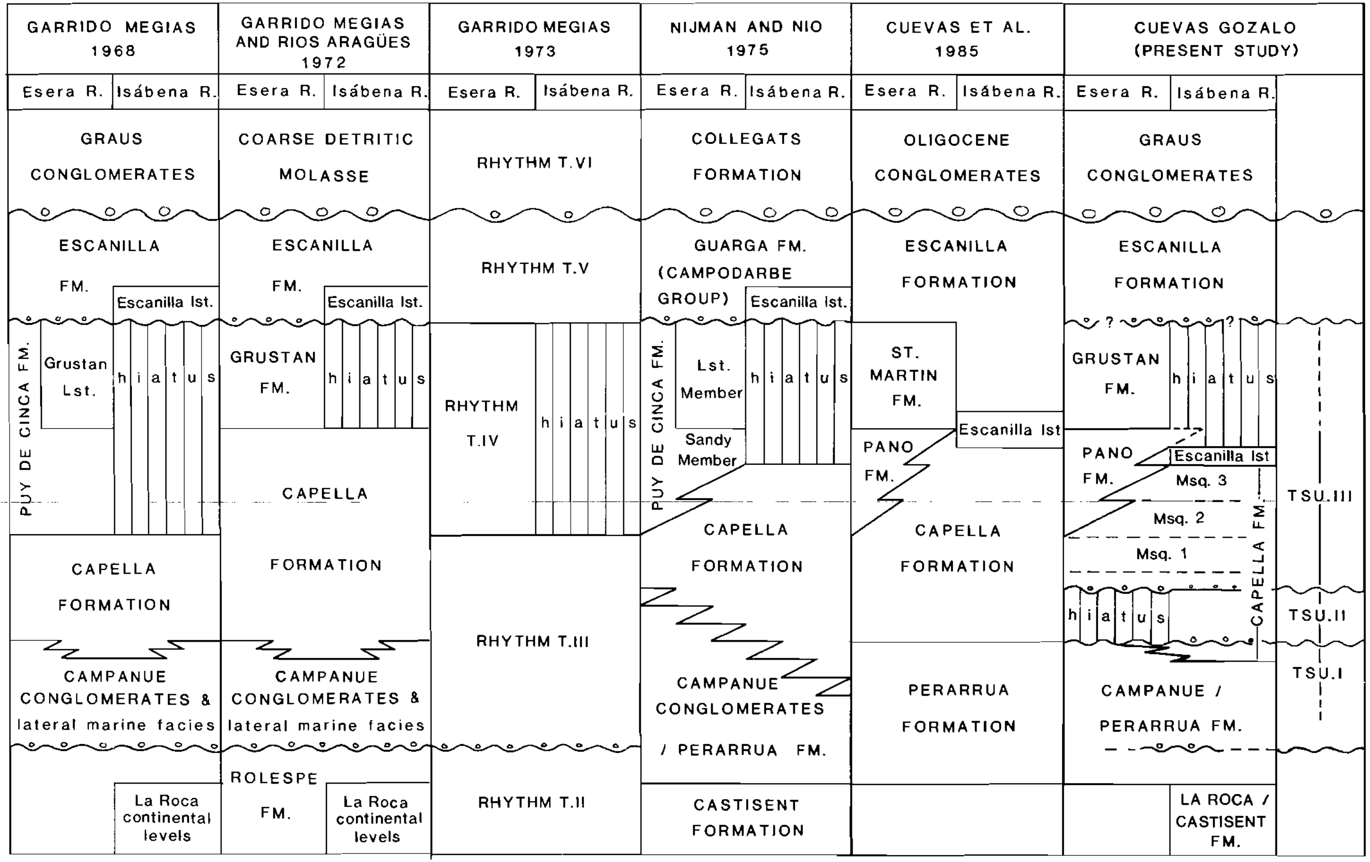
According to Nijman and Nio (1975) the Capella Formation consists of more or less isolated sandstone bodies interbedded with a thick succession of multicoloured mudstones. They interpreted these sandstone bodies as fluvial channels in an alluvial plain setting, representing the upper deltaic plain of the Montañana delta. The westward progradation of the Montañana delta produced the superposition of the Capella Formation on top of the more distal and marine Perarrúa Formation (see fig. 2.3; Nijman and Nio, 1975).

The present study, however, has shown that a large part of the Capella Formation in the Isábena and Esera valleys is not the landward equivalent of the underlying Perarrúa Formation:

1) Within the upper part of the Perarrúa Formation, a gradual change to continental, yellowish facies can be observed (fig. 2.6). This has been the reason for the progradational pattern as interpreted by Nijman and Nio (1975).

2) Field observations during the present study, however, have revealed a distinct unconformity between the Perarrúa and Capella Formations. This shows that the larger part of the Capella Formation probably does not represent the landward equivalent of the underlying Perarrúa Formation as postulated by Nijman and Nio (1975).

Fig. 2.3.- *(next page) Lithostratigraphic position of the Capella Formation in relation to other stratigraphic units in the area of Graus, according to different authors.*


GARRIDO MEGIAS 1968
GARRIDO MEGIAS AND RIOS ARAGÜES 1972
GARRIDO MEGIAS 1973
NIJMAN AND NIO 1975
CUEVAS ET AL. 1985
CUEVAS GOZALO (PRESENT STUDY)
Esera R.
Isábena R.
GRAUS CONGLOMERATES
COARSE DETRITIC MOLASSE
RHYTHM T.VI
COLLEGATS FORMATION
OLIGOCENE CONGLOMERATES
ESCANILLA FM.
Escanilla lst.
RHYTHM T.V
GUARGA FM. (CAMPODARBE GROUP)
ESCANILLA FORMATION
PUY DE CINCA FM.
Grustan Lst.
h i a t u s
GRUSTAN FM.
RHYTHM T.IV
Lst. Member
Sandy Member
ST. MARTIN FM.
Escanilla lst
PANO FM.
Msq. 3
Msq. 2
Msq. 1
CAPELLA FM.
TSU.III
TSU.II
TSU.I
CAPELLA FORMATION
RHYTHM T.III
CAMPANUE CONGLOMERATES & lateral marine facies
CAMPANUE CONGLOMERATES / PERARRUA FM.
PERARRUA FORMATION
CAMPANUE / PERARRUA FM.
La Roca continental levels
ROLESPE FM.
RHYTHM T.II
CASTISENT FORMATION
LA ROCA / CASTISENT FM.

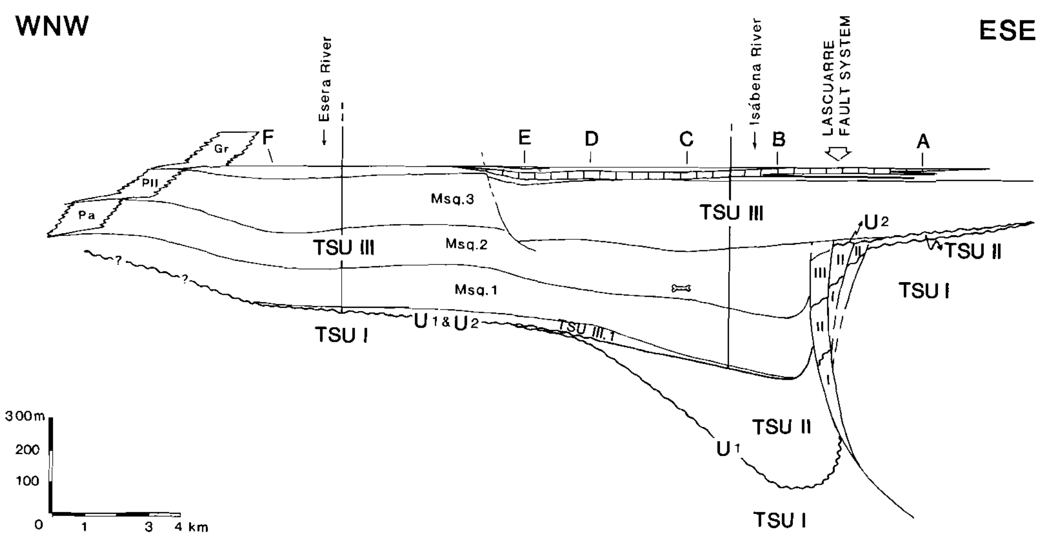


Fig. 2.4.- *Relations between the different tectonosedimentary units; bounding unconformities and thickness distribution.*

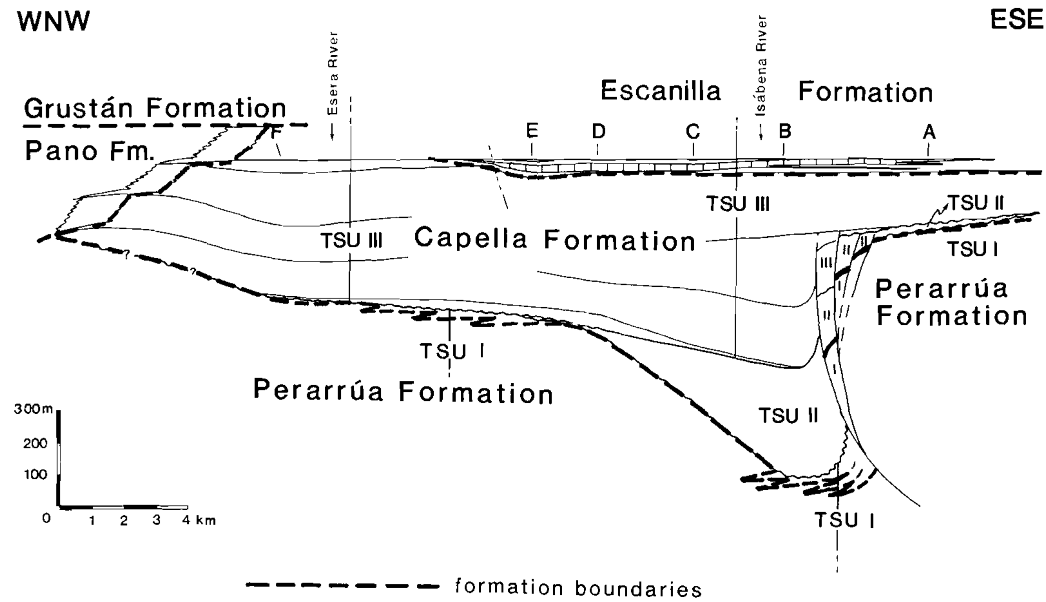


Fig. 2.5.- *Relation between lithostratigraphic formations and tectonosedimentary units.*

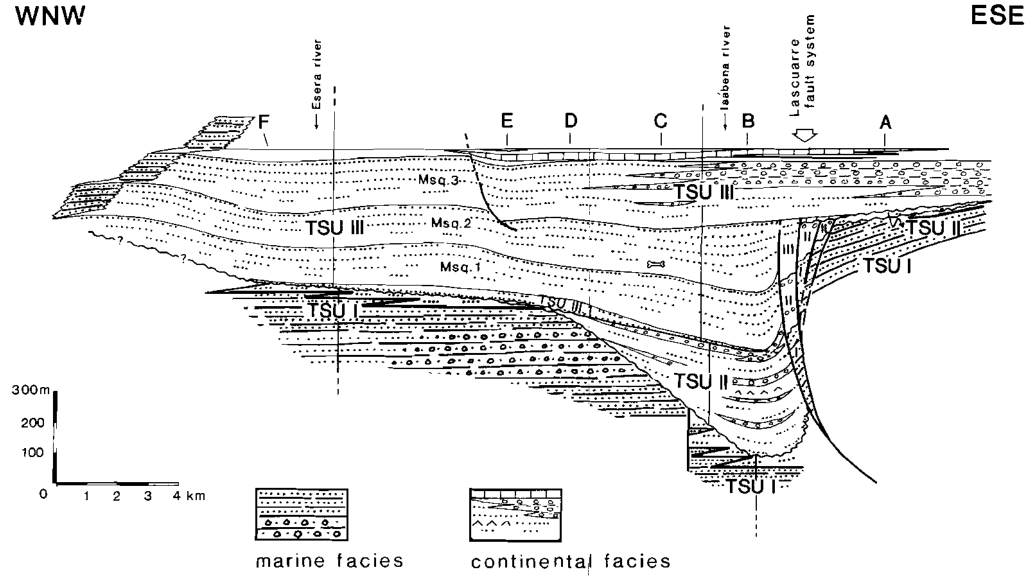


Fig. 2.6.- *Large-scale lithofacies distribution with relation to the tectonosedimentary units.*

Capella Formation upper boundary in the Esera valley

In the western margin of the Esera valley the Capella Formation grades westwards and vertically to a few hundred meters thick succession of sandstones and mudstones with marine fauna. This clastic succession is overlain by marine limestones. Both the clastic and the calcareous successions were included in a single stratigraphic unit, the Puy de Cinca Formation, by Garrido Megías (1968) (fig. 2.3). Later works do not all consider the clastic and the calcareous successions to be part of the same formation (fig. 2.3). The sandstone succession of the Puy de Cinca Formation was renamed the Pano Formation by Donselaar and Nio (1982). The Pano Formation consists of transitional and marine clastic deposits, which are the lateral equivalent of the Capella Formation in the west. The Pano Formation was formed as a barrier island system which onlapped the coastal plain deposits of the Capella Formation during the middle Lutetian transgression (Donselaar, pers. com.). According to Donselaar and Nio (1982) the barrier island system formed a northeast-southwest trending coastline, with land to the southeast.

In the present study the name Pano Formation is kept for the sandstone interval, and the name Grustán Formation (Garrido Megías and Ríos Aragües, 1972) is considered most appropriate for the overlying marine limestones (figs. 1.1 and 2.3).

Capella Formation upper boundary in the Isábena valley

The Capella Formation is overlain in the Isábena valley by lacustrine limestones (Alastrúe et al. 1957; Crusafont et al., 1968). Garrido Megías (1968) included these limestones in the basal part of the younger Escanilla Formation (fig. 2.3). He suggested a regional erosive unconformity at the base of the Escanilla Formation. However, at the scale of outcrop no unconformity has been observed between the Capella Formation and the basal limestones of the Escanilla Formation (fig. 2.3).

Capella Formation upper boundary: correlation problems

The lithostratigraphic relation between the Pano Formation and the Escanilla Formation's basal limestones is difficult to establish in the field: both formations, as well as the upper part of the Capella Formation have disappeared in the Esera valley below the paleovalley-shaped unconformity of the Graus conglomerates (figs. 1.1 and 2.3). Sediments below this unconformity are displaced by faulting. In the present study the tectonosedimentary and the architecture analysis of the Capella Formation and the subsequent interpretation of the formation in terms of basin events have been used to establish the relationship between the facies (fig. 2.3).

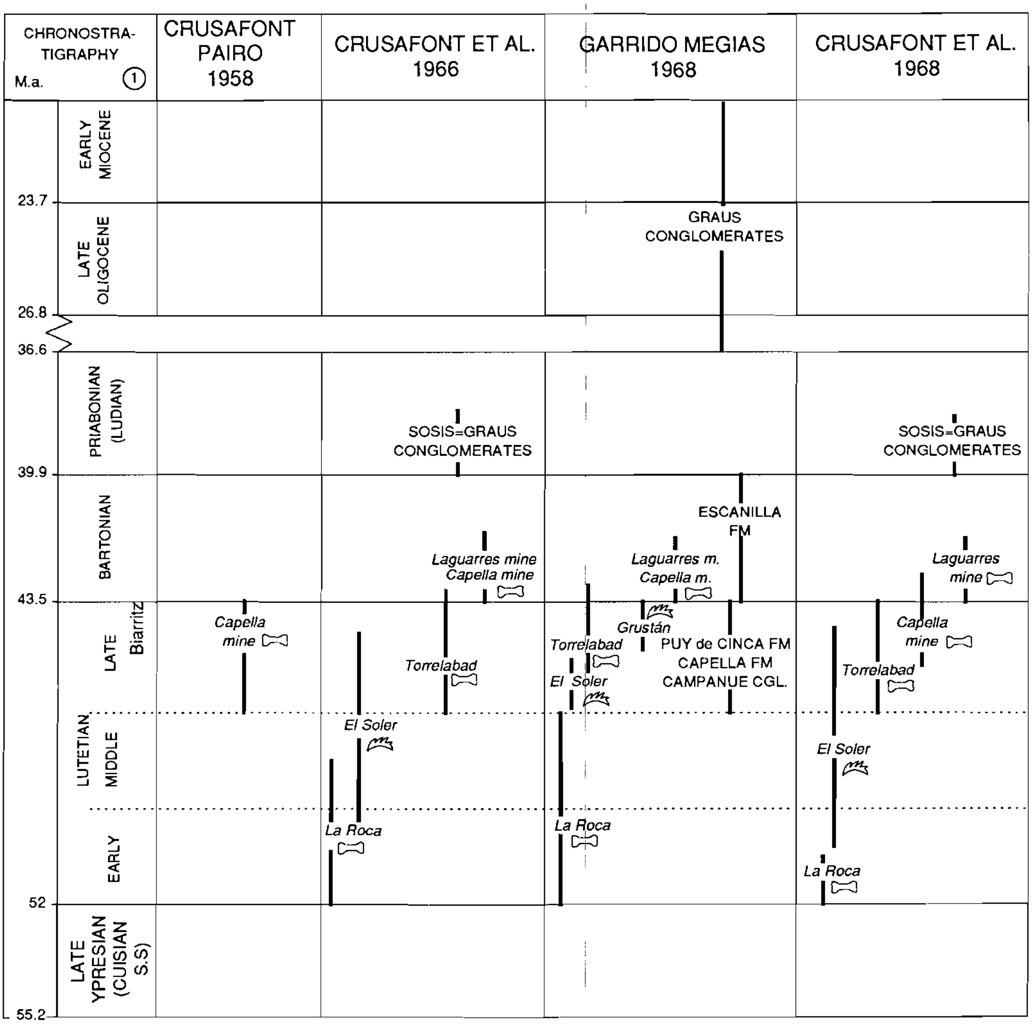
Tectonosedimentary units and lithostratigraphic formations

The three tectonosedimentary units recognized in the present study each include a part of the Capella Formation:

- Tectonosedimentary Unit I comprises the Perarrúa Formation and the lowermost part of the Capella Formation.
- Tectonosedimentary Unit II includes the middle part of the Capella Formation.
- Tectonosedimentary Unit III comprises the upper part of the Capella Formation together with the lower part of the Escanilla Formation, the Pano Formation and probably the Grustán Formation as well (figs. 2.3 and 2.5).

2.3.- Biostratigraphy of the continental Eocene - Oligocene deposits in the Isábena and Esera valleys and associated marine deposits

A large part of the Tremp-Graus Basin is filled in with continental deposits; these grade basinwards into marine deposits. The study area is of special interest because it offers the possibility of linking the biostratigraphical zonation based on marine fauna to the zonation based on continental fauna. The biostratigraphical data serve as a framework for the correlation of the tectonosedimentary units on a regional scale, and are thus important in establishing the tectonic and sedimentary heterogeneities in the basin.

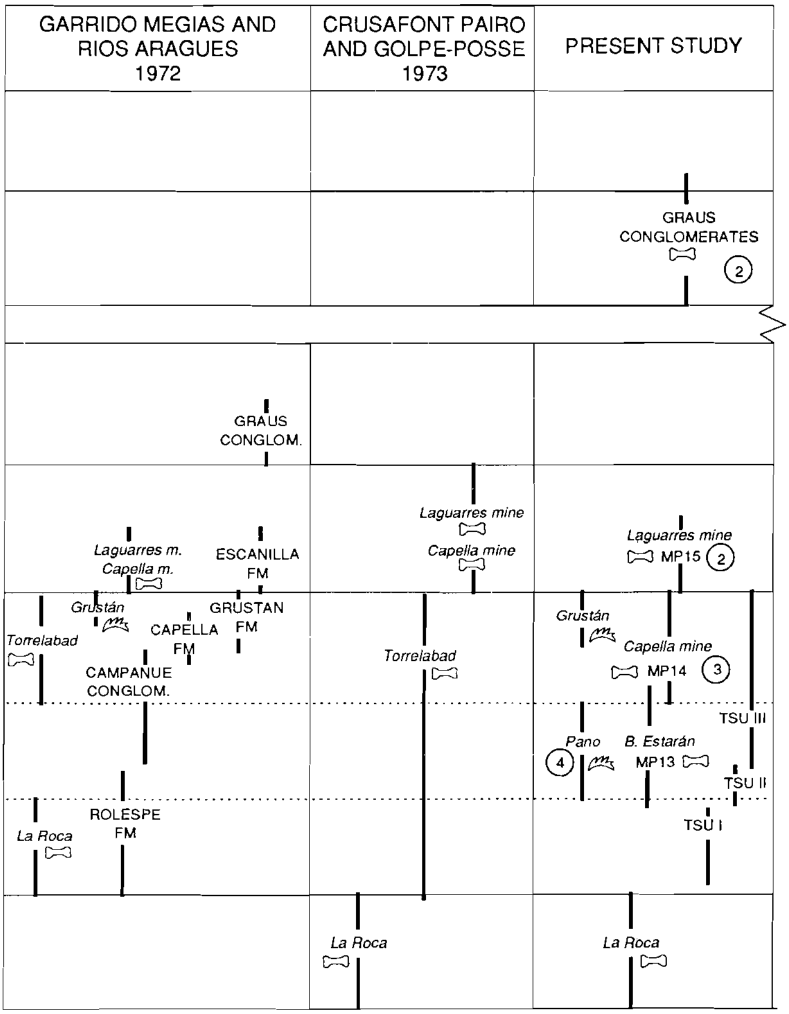


La Roca = vertebrate fossil locality

El Soler = marine invertebrate fossil locality

CAPELLA FM. = stratigraphic unit

Fig. 2.7.- *Chronostratigraphic position of the Capella Formation in relation to other stratigraphic units in the area of Graus, according to different authors.*



(1) After Berggreen et al. 1985

(2) López et al. (pers. com.)

(3) Schmidt-Kittler, 1987

(4) Jiménez (pers. com.)

In this section a historical review of the biostratigraphic dating of Eocene to Oligocene continental deposits in the Isábena and Esera valleys is presented; biostratigraphic data of the interbedded marine deposits is also considered (figs. 2.3 and 2.7).

The continental intervals of **La Roca** are the oldest Tertiary deposits of the Isábena valley that have supplied abundant terrestrial fauna allowing a biostratigraphic dating. The vertebrates association found here has been dated first as "early Lutetian" (Crusafont et al., 1966, 1968) and later as "Cuisian" (Crusafont Pairó and Golpe-Posse, 1973) (fig. 2.7).

This continental succession is overlain by the continental to marine Campanúe/Perarrúa Formation (figs. 1.1 and 2.3). The marine fossils of **El Soler** at the top of the sequence have a "Lutetian-Biarritzian" age (Crusafont et al., 1966, 1968) (fig. 2.7).

The **Torrelabad** vertebrates locality occurs in the transition facies between the Perarrúa Formation and the Capella Formation. This locality has supplied a fauna of *Lophiodon*, chelonids and crocodiles, which has been assigned to different "zones" of "Lutetian" to "Bartonian" age by Crusafont et al. (1966, 1968) and Crusafont Pairó and Golpe-Posse (1973) (fig. 2.7).

A new fossil vertebrate locality was discovered during the field work for this study. The vertebrate locality is named **Barranco Estarán** after the gully where it occurs (fig. 1.1). Stratigraphically it is situated in the middle of Tectonosedimentary Unit III, in Capella facies (fig. 2.6). Biostratigraphic research carried out by López, Daams and Van der Meulen (pers. com.) indicates that the rodent fauna of this locality correlates with the mammal level MP 13 (Lutetian) (fig. 2.7).

In the Isábena valley the Capella Formation is overlain by lacustrine limestones with Mollusca (Escanilla Formation). Some lignite intercalations within the limestones have supplied a mammal fauna. The remains of *Lophiodon*, Palaeodonta, lemurids and artiodactyles of the **Capella mine** have been ascribed to different zones of "late Lutetian" to "Bartonian" age (Crusafont Pairó, 1958; Crusafont et al., 1966, 1968) and Crusafont Pairó and Golpe-Posse (1973). This fossil locality has been recently assigned to the mammal level MP 14 (Schmidt-Kittler, 1987), which according to Hooker (1987) could correlate with the late Lutetian (fig. 2.7).

The vertebrates locality at the **Laguarres mine** also occurs in a limestone interval with lignites. This interval occupies a much higher lithostratigraphic position within the Escanilla Formation than the limestones of the Capella Mine. Crusafont et al. (1968) have already indicated different stratigraphic positions for both vertebrates associations (fig. 2.7). According to López (pers. com.) the fauna of Laguarres correlates with the mammal level MP 15 (Lutetian-Bartonian).

In the Esera valley the Capella Formation grades vertically into the marine detritic Pano Formation; from its lower part (**Pano** village) this formation has supplied a *Nummulites* association of middle Lutetian age (Jiménez, pers. com) (fig. 2.7).

The overlying **Grustán** marine limestone was correlated with the more western Samitier limestone and subsequently a latest Lutetian or Biarritzian age was assigned to it (Garrido Megías, 1968 and Garrido Megías and Ríos Aragües, 1972, respectively) (fig. 2.7).

The youngest Paleogene deposits in the area are the Graus conglomerates. Different ages have been ascribed to these conglomerates depending on the levels with which they are correlated (fig. 2.7). Recent biostratigraphic research on micromammals indicate a late Oligocene to Oligocene/Miocene boundary age for these conglomerates (López, Daams and Van der Meulen, pers. com.) (fig. 2.7).

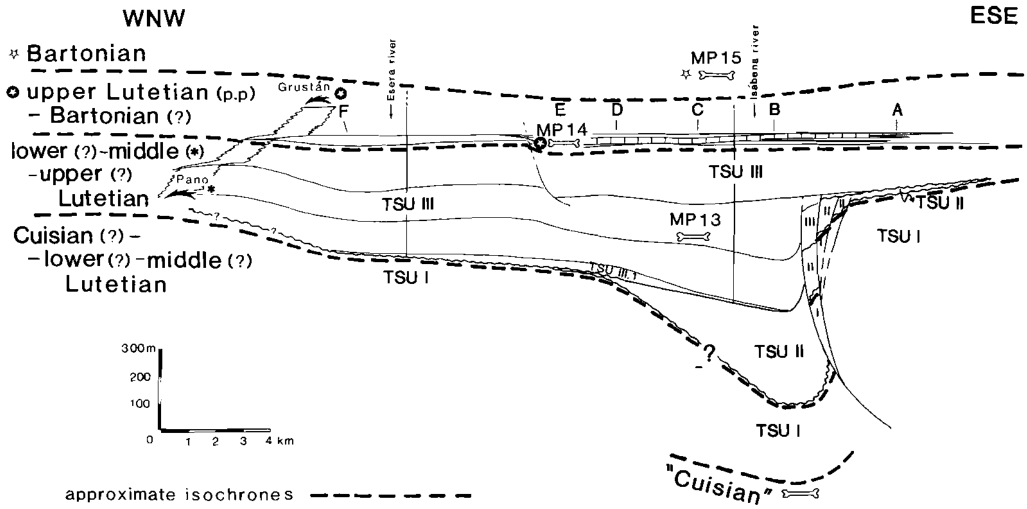


Fig. 2.8.- *Situation of fossil localities in relation to the tectonosedimentary units.*

Discussion on the dating of the tectonosedimentary units

Both the Perarrúa Formation and the Capella Formation were deposited between the Cuisian and the late Lutetian (fig. 2.7 and 2.8). The upper Capella Formation (TSU III) includes the middle Lutetian because of its lateral relation with the Pano fossil locality (fig. 2.8). The Perarrúa Formation (TSU I) lies between the La Roca/Castisent Formation and the upper part of the Capella Formation (TSU III) in the Isábena valley, and below the Pano Formation (TSU III) in the Esera valley. Consequently, a Cuisian (?)- early Lutetian (?)- middle Lutetian (?) age can be assigned to it.

Until now, the Escanilla limestone was assumed to stratigraphically overlie the Grustán limestone. However, both the recent assignation of the Capella mine to the level MP 14 and the stratigraphic lowering of this mammal level in relation to the marine biostratigraphy make this assumption invalid. Based on sequential analysis, the present study proposes a slightly higher stratigraphic position within the upper Lutetian for the Grustán limestone than for the Escanilla basal limestone (figs. 2.3 and 2.8). It is assumed that the upper part of the Pano Formation might also be of late Lutetian age.

Based on these considerations, the following chronostratigraphy of the tectonosedimentary units is proposed (fig. 2.8):

- Tectonosedimentary Unit I (upper part) could be of early Lutetian age.
- Tectonosedimentary Unit II could be of early or/and middle Lutetian age.
- Tectonosedimentary Unit III includes the middle and late Lutetian.

3.- STRATIGRAPHIC UNITS: TECTONOSEDIMENTARY UNITS

3.1- Introduction

Several complete sedimentological logs have been made through the Capella Formation from the underlying Perarrúa Formation to the overlying Escanilla and Pano Formations (fig. 2.9). Three tectonosedimentary units are recognized. In the present study a tectonosedimentary unit is defined as a lithostratigraphic unit bounded by unconformities or their correlative conformities. Tectonosedimentary units are not directly related to facies. A single tectonosedimentary unit may include different facies/formations; a single formation may be subdivided into several tectonosedimentary units (fig. 2.5).

3.2.- Unconformities in the Tremp-Graus Basin

Unconformities in the Tremp-Graus Basin are gradational. In general, the northern basin margin was uplifted and eroded while sedimentation continued in the axial depositional trough. As a result of this, angular unconformities are easy to recognize in the northern part of the basin, while southwards the unconformities grade into conformities. Garrido Megías (1968) described this type of unconformity between Lutetian deposits of the Tremp-Graus Basin, and suggested the southward migration of the depositional trough. Cámara and Klimowitz (1985) related the migration of the depositional troughs to the displacement of thrust-sheets.

According to the model proposed by Riba (1973) this type of gradational unconformity is the result of two stages: a first stage of diastrophic acceleration, during which the tectonic deformation/sedimentation ratio increases, and a second stage of diastrophic slowing down, during which the tectonic deformation/sedimentation ratio decreases. The first stage is characterized by a rapid tectonic deformation and offlaping of the sediments; the second stage is characterized by a mild tectonic deformation and onlapping of the sediments.

3.3.- Tectonosedimentary units and bounding unconformities

Tectonosedimentary Unit I (TSU I) consists of the Perarrúa Formation and of the base of the Capella Formation in the Isábena and Esera valleys (fig. 2.5). The determination of the lower boundary is not an aim of this study. The upper boundary of TSU I is determined by angular and erosional unconformities with the overlying TSU II and TSU III (fig. 2.4):

Unconformity between TSU I and TSU II (U_1):

A cartographic angular unconformity is recognized between TSU I and TSU II by the general difference of tectonic strike: the deposits of TSU II have a strike direction around 110° E, while the directly underlying deposits of TSU I have a strike direction of 140° E.

The unconformity can be observed in detail at various points:

1) In the most eastern part of the study area (east of section A, figs. 1.1 and 2.4), the basal deposits of TSU II are erosively discordant over the caliche horizons forming the top of TSU I. Transported caliche fragments (carbonate nodules) are characteristic of the basal deposits of TSU II.
2) The uppermost sandstone body of TSU I preserved at section B, (figs. 2.6 and 2.10) is tectonically deformed, forming a depression in its southern part. Basal deposits of TSU II, rich in transported carbonate nodules unconformably fill in the depression. Similar changes of tectonic dip, associated with deposits rich in transported carbonate nodules, are observed at the eastern margin of the Isábena River, near Salanova (fig. 1.1).

Unconformity between TSU I and TSU III (U_1+U_2):

In the Esera valley, south of Ventas de Santa Lucía (figs. 1.1, 2.4 and 2.5) a clear angular erosional unconformity is measured between the Perarrúa facies of TSU I (strike 110° E, dip 14°) and coarse subarkosic sandstones at the base of TSU III (strike 130° E, dip 5°). At the base of section E and 2.5 km north of Capella (fig. 1.1), the same unconformity is recognized as a subtle angular and erosional unconformity between the Capella facies of TSU I and the basal conglomeratic subarkosic sandstones of TSU III. Both in the Esera valley and in the area around Capella, the basal subarkoses overlie a system of E-W trending normal faults.

Tectonosedimentary Unit II (TSU II) consists of the middle part of the Capella Formation in the Isábena valley and eastwards of it (fig. 2.5). The upper boundary of TSU II is determined by an unconformity with the overlying TSU III:

Unconformity between TSU II and TSU III (U_2):

Southwest of Lascuarre, conglomerates and sandstones of TSU II are affected by a system of faults (figs. 2.4 and 2.6). A thin level of pedogenic limestones unconformably overlies the deposits and faults. The pedogenic limestones are included in TSU III.

East of Lascuarre (section A) there is a paraconformity between TSU II.1 and deposits of the upper part of TSU III. West of Lascuarre the unconformity grades into a conformity (fig. 2.5); the boundary between TSU II and TSU III is

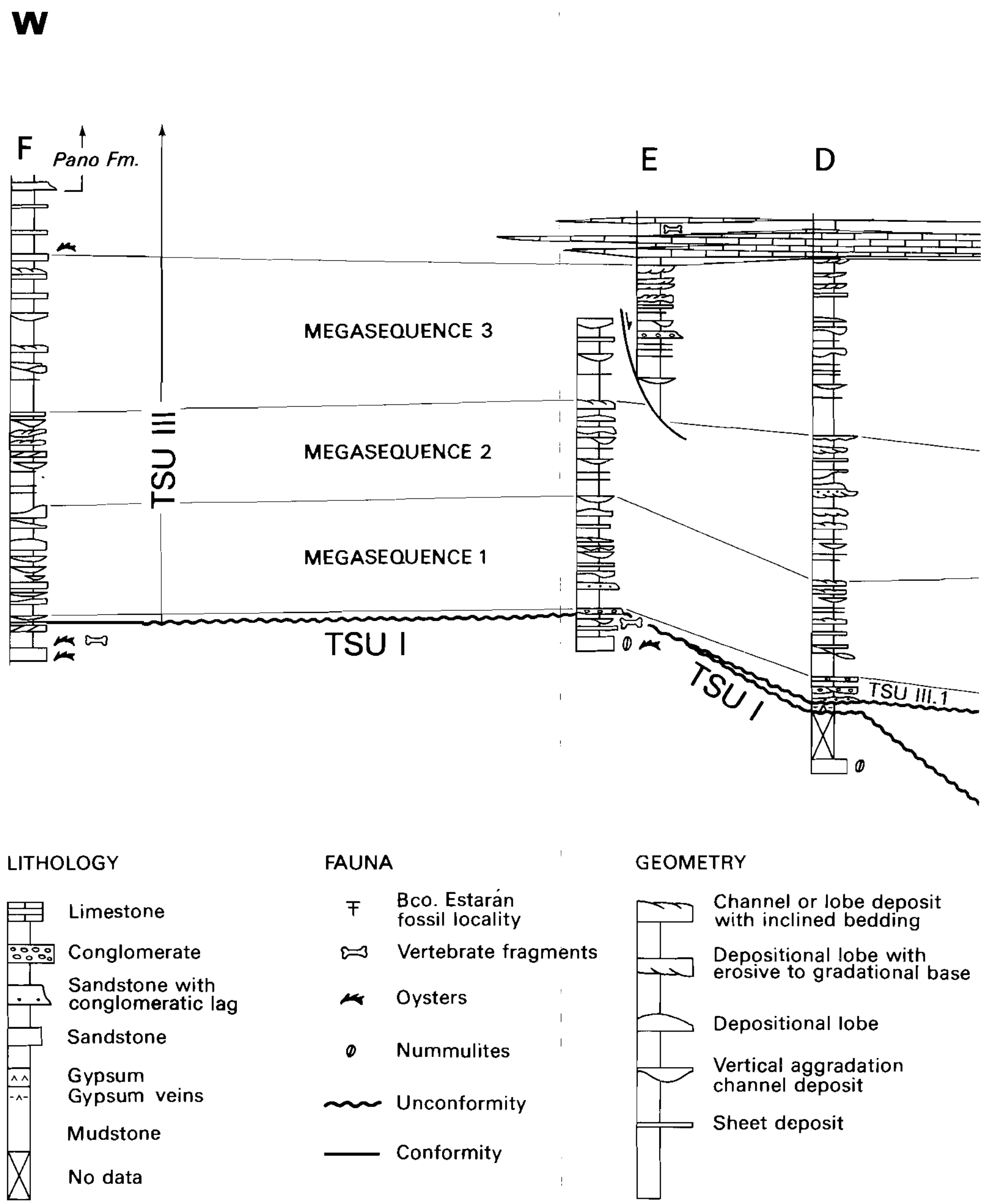
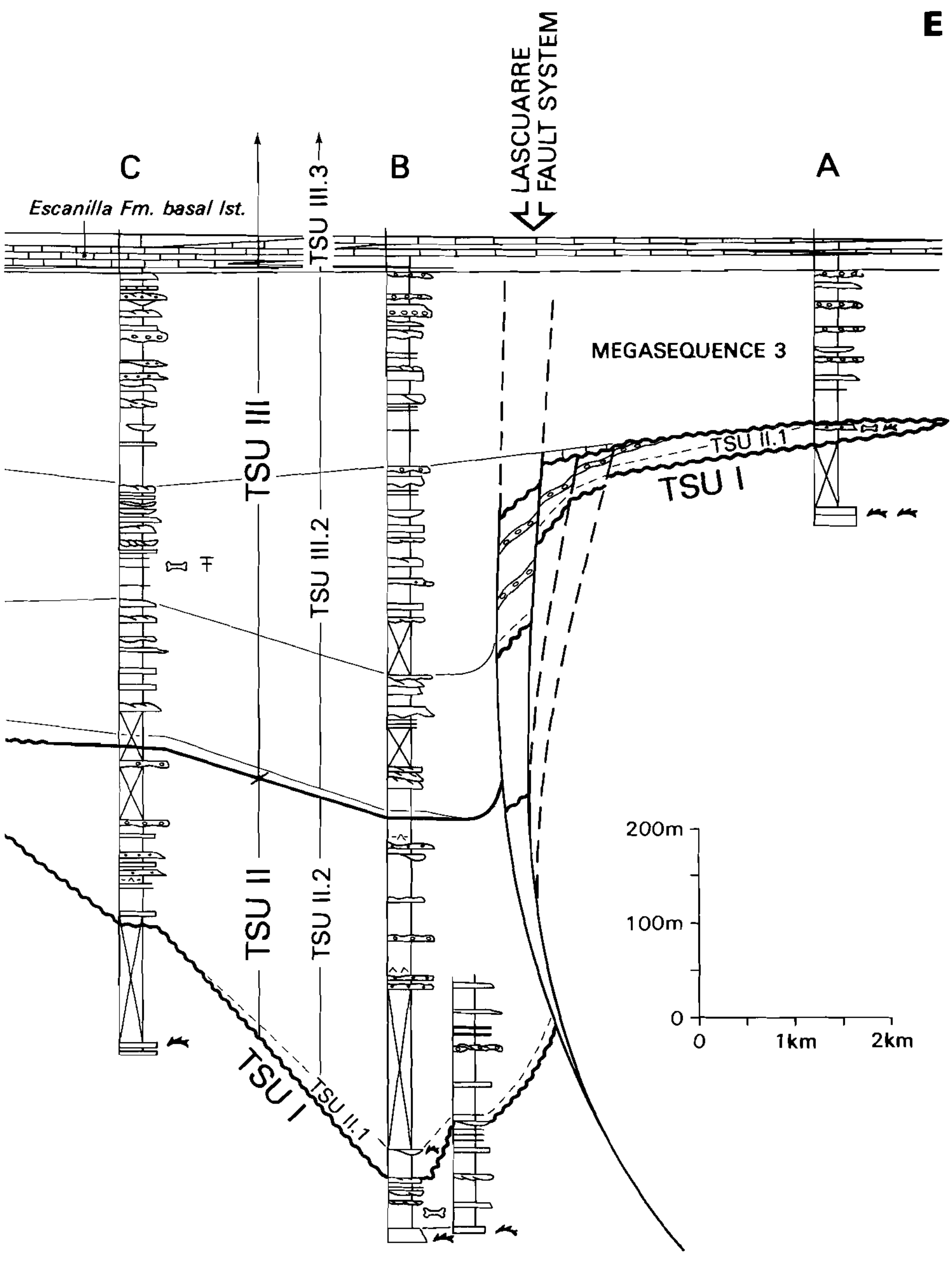


Fig. 2.9.- *General distribution of sediment bodies within the tectonosedimentary units and subunits.*


E
LASCUARRE FAULT SYSTEM
C
B
A
Escanilla Fm. basal lst.
TSU III.3
TSU III
TSU III.2
MEGASEQUENCE 3
TSU II.1
TSU I
TSU II
TSU II.2
TSU I
TSU II.1
200m
100m
0
0
1km
2km

defined by the disappearance of conglomerates interbedded with mudstones with gypsum veins (TSU II.2) and the occurrence of pebbly subarkoses (TSU III).

Tectonosedimentary Unit III (TSU III) comprises the upper part of the Capella Formation together with the lower part of the Escanilla Formation in the Isábena valley, the Pano Formation and probably the Grustán Formation as well (fig. 2.5).

The upper boundary of TSU III is not observed within the Capella Formation. Probably it is represented by an unconformity at the base of the lowest conglomerates of the overlying Escanilla Formation (fig. 2.3). The Escanilla conglomerates would be unconformable over the lower Escanilla Formation in the Isábena valley and over the Grustán Formation in the Esera valley.

3.4.- Tectonosedimentary Unit I. Facies subdivision

Tectonosedimentary Unit I is present in the whole study area (fig. 2.4). Within TSU I, three types of facies associations are recognized. They occur in vertical and lateral transition to each other (fig. 2.10).

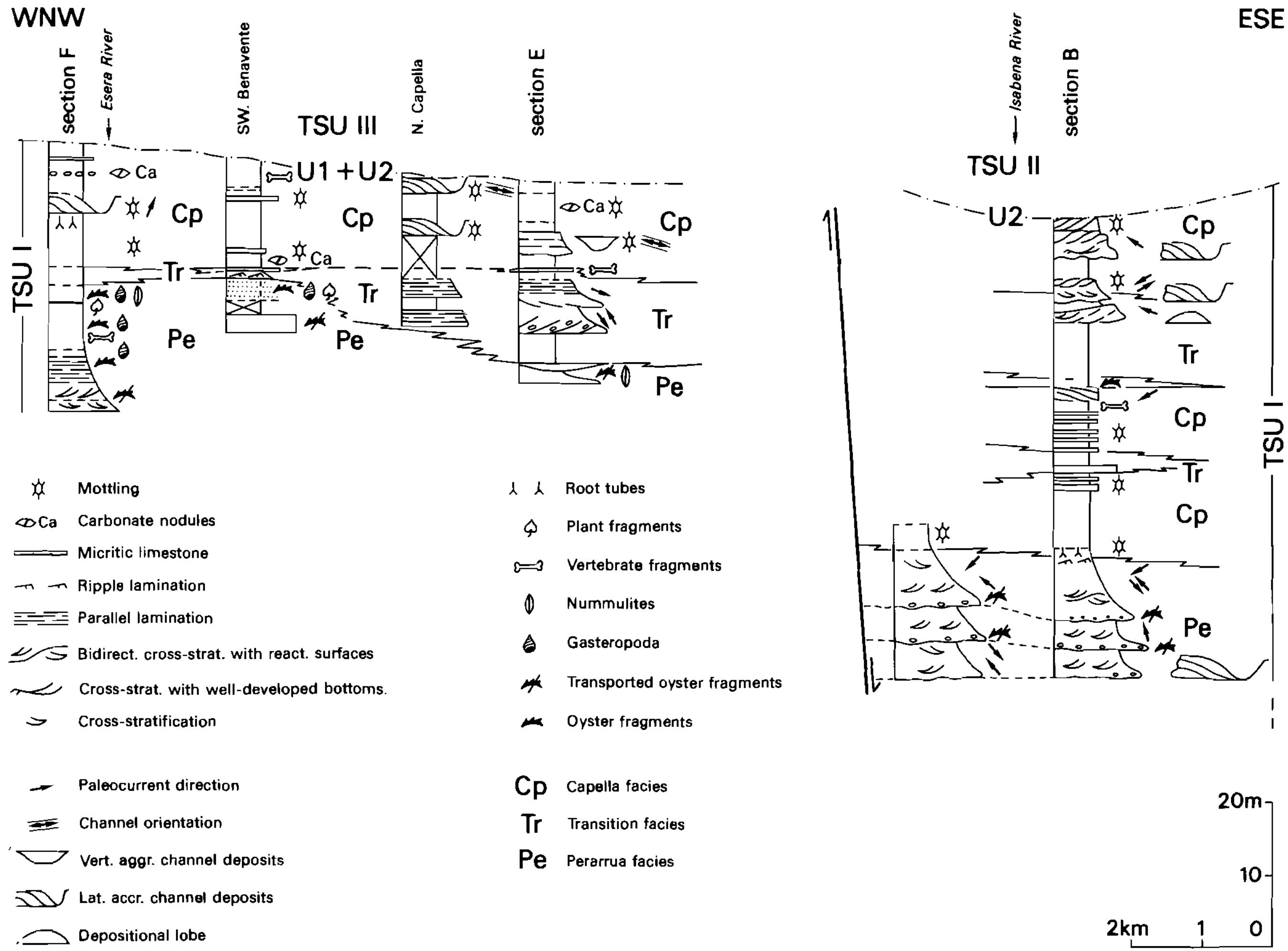
1) *Perarrúa facies association* consists of sandstones and marls with marine fauna (oysters, nummulites and gasteropoda). Characteristic of this facies is the grey colour of the mudstones and the disorganized bioturbation pattern of the sandstones. Sandstone bodies in the Perarrúa Formation are in general thicker and have larger lateral extent than those of the Capella facies association.

   A vertical trend from sandstones to mudstones with oysters, gasteropoda and plant fragments is observed in the upper part of TSU I in the west of the study area (fig. 2.10).

2) *Transition facies association* consists of sandstones and mudstones without marine fauna or ochre colouring. Some patches of bioturbated white micritic limestone occur in the western area, in the upper part of TSU I (fig. 2.10).

3) *Capella facies association* consists of both lenticular and tabular sandy bodies interbedded with mudstones; sandstones and mudstones are predominantly ochre, although green to violet mudstones containing turtle fragments occur near the transition facies. Carbonate concretions occur in ochre mudstones (section E) and in sandstones (section B). Eastwards of section A the Capella facies association is characterized by ruiniform calcitic soil horizons.

Fig. 2.10.- *(next page) Facies distribution in the upper part of TSU I at certain points below the unconformities. In all cases a gradational vertical trend to more continental deposits occurs. Lithofacies in the eastern part (section B) possibly correspond to lower stratigraphic levels than those in the eastern part (sections E, F).*



WNW
ESE
section F
Esera River
SW. Benavente
N. Capella
section E
Isabena River
section B
TSU III
U1 + U2
TSU II
U2
TSU I
Ca
Cp
Tr
Pe
Mottling
Carbonate nodules
Micritic limestone
Ripple lamination
Parallel lamination
Bidirect. cross-strat. with react. surfaces
Cross-strat. with well-developed bottoms.
Cross-stratification
Paleocurrent direction
Channel orientation
Vert. aggr. channel deposits
Lat. accr. channel deposits
Depositional lobe
Root tubes
Plant fragments
Vertebrate fragments
Nummulites
Gasteropoda
Transported oyster fragments
Oyster fragments
Cp Capella facies
Tr Transition facies
Pe Perarrua facies
20m
10
2km 1 0

The maximum observed thickness of Transition and Capella facies associations together is not more than 45 m (section B). The synsedimentary faulting, the thickness variation below the unconformities and the discontinuity of the exposures impede a good lateral facies correlation within TSU I. The uppermost deposits of TSU I preserved in the Isábena River below the unconformity U_2 (section B in fig. 2.10) possibly are older than those preserved in the western area below the unconformity U_1+U_2 (sections E and F in fig. 2.10).

3.5.- Tectonosedimentary Unit II. Lithological subdivision

Tectonosedimentary Unit II includes the middle part of the Capella Formation (fig. 2.5). Two subunits are recognized within TSU II (fig. 2.9):

TSU II.1: It is characterized by the presence of transported carbonate nodules. These occur either within sandstone bodies (east of section A, section A, Lascuarre and section B) or interbedded with yellow-orange mudstones (section B). The carbonate nodules are derived from the erosion of paleosols, probably the calcitic soils at the top of the underlying TSU I. In general only one or two levels, 2-3 m thick, containing transported carbonate nodules are observed at each point.

TSU II.I extends from east of section A to the west of section B. The maximum thickness of the subunit is 42 m at section B (fig. 2.9). Because of the abundance of reworked soil components and its position at the base of the unit, this subunit can be considered as the "basal conglomerate" of TSU II.

TSU II.2: It is characterized by calcareous conglomerates and pebbly carbonatic litharenites interbedded with red and ochre mudstones containing gypsum veins. Some subarkosic levels occur locally; fine lithoarenites are common. Coarse components in the conglomerates are up to 40 cm in diameter (section B). Fragments of laminated yellowish sandy limestone (Upper Cretaceous) and grey calcareous mudstone (upper Cenomanian-Turonian) are the most common components.

Massive, nodular and fibrous gypsum occur at section B. In other areas only nodular and fibrous gypsum is present.

TSU II.2 is well represented from Lascuarre up to Pociello. The maximum preserved thickness occurs at section B (397 m), decreasing gradually to the west (section C). The thickness decreases abruptly eastwards. TSU II.2 is cut by the fault system of Lascuarre (fig. 2.9).

3.6.- Tectonosedimentary Unit III. Lithological subdivision

TSU III comprises the upper part of the Capella Formation, the lower part of the Escanilla Formation, the Pano Formation and probably the Grustán Formation as well (figs. 2.3 and 2.5). Based on lithological criteria, three subunits are recognized within TSU III. The two lower subunits (TSU III.1 and TSU III.2) include the upper part of the Capella Formation, while the third (TSU III.3) includes the lower part of the Escanilla Formation (fig. 2.9). In the following descriptions particularly the subunits of TSU III with facies Capella are referred to; a brief description of the subunit with facies Escanilla is also included. Detailed studies of the Pano Formation have been published by Donselaar and Nio (1982).

TSU III is the thickest and most extensive unit within the Capella Formation. Therefore more logs were obtained from it. East-west correlation of the logs, with the thickness of the different units and subunits is summmarized in fig. 2.9. The thickness of TSU III with facies Capella varies between 173 m (section A) and 592 m (section B).

Lithologically the TSU III with facies Capella consists of sandstones and conglomerates interbedded with mudstones. Mudstones and sandstones occur in the whole unit, while conglomerates appear only towards the top (fig. 2.6). The mudstones vary in colour from red to brown to ochre and are commonly mottled. Grey mudstones occur locally. Sandstones are mostly lithoarenites; coarse fragments are observed in the lag deposits of some sandstone bodies. Pebbly to conglomeratic subarkosic sandstones are concentrated at the base of the unit, just above the unconformity. Several associations of coarse fragments are recognized:

1) Quartzite, black chert and dark grey limestone fragments, eroded from the Mesozoic cover of the Pyrenees. This association occurs throughout TSU III.
2) Yellowish, laminated sandy-limestone fragments (Santonian-Maestrichtian) and alveolina-limestone fragments (late Paleocene-early Ypresian), eroded from the uplifted basin-margins. This association is found a) as lag deposits in the subarkoses of TSU III.I, just above the unconformity, and b) forming conglomeratic levels in the upper part of TSU III.2.
3) Granitic fragments. They are scarce and are only found in the subarkoses at the base of the unit (TSU III.1).
4) Doleritic fragments. They are very rare; occasionally found in the subarkoses at the base of the unit (TSU III.1).
5) White micritic limestone fragments. These are very rare and are only found in the conglomerates in the upper part of TSU III.2.

Description of the subunits:

TSU III.1:

It is characterized by a predominance of sandstone bodies, generally of subarkosic type. The subunit has its depocenter in the central area (sections D and E and north of Capella), with a maximum total thickness of 23 m in section D. Here the subunit consists of up to three subarkosic sandstone bodies (each 2-4 m thick) interbedded with mudstones. The subunit wedges out eastwards: in section B it consists of a 1.5 m thick level of pebbly subarkosic sandstones; in section A no subarkoses are observed. Towards the west (Esera valley and section F), the subunit is represented by one or two sandstone bodies (each 1.5-2 m thick) interbedded with mudstones.

A clear lateral variation in the coarse components of these subarkoses is observed: largest coarse fragments appear in section E: the subarkoses above the unconformity contain clasts of alveolina limestone and Upper Cretaceous sandy-limestone up to 30 cm in diameter as lag deposits. Granitic clasts of smaller size are also found. In section D the coarse fragments are smaller (of the order of centimeters) and less abundant. Besides detrital carbonate granite and dolerite fragments are also observed. In section B only carbonate pebbles are found. Westwards of section E a general decrease of maximum particle size is observed. Near Capella coarser fragments are carbonate clasts of the order of centimeters; in the Esera valley and at section F, the subarkoses are fine-pebbled to coarse sandstones. Pebbly-sized iron nodules occur.

Because of the relative abundance of coarse components and its position just above the unconformity, TSU III.1 can be considered as the "basal conglomerate" of TSU III. The grain size and thickness distribution within this subunit suggest that the depositional system was transverse to the east-west profile studied and that it had its axis in the central area, between Torrelabad and Capella.

The boundary with TSU III.1 is marked by the apparition of a relatively thick interval dominated by mudstones.

TSU III.2:

It is a heterolithic unit (sandstones and conglomerates interbedded with mudstones) characterized by the scarcity of conglomerates at the base. Subdivisions are based on sequential criteria.

This subunit is recognized in the whole area. Sediment thickness varies gradually from section B (590 m) to section E (454 m). A sharp decrease of thickness which is related to the fault system of Lascuarre occurs between section B and section A (173 m) (figs. 2.4 and 2.9).

The upper boundary of the subunit is determined by the upper boundary of the Capella Formation in the Isábena valley.

TSU III.3:

It consists of the lower part of the Escanilla Formation. The subunit includes some of the lithostratigraphic levels recognized by Van den Bergh (pers. com.):

1) lacustrine limestones and mudstones;
2) a thick interval of hydromorphic soils with some interbedded lenticular sandstone bodies towards the top;
3) tabular sandstone bodies interbedded with mudstones.

The upper boundary of the subunit is marked by the erosion surface at the base of the overlying Escanilla conglomerates. This erosion surface probably corresponds with a regional erosive unconformity. A regional study would be necessary to prove this.

This subunit is not discussed in the present study. However, it will be mentioned again in the chapter on architectural analysis, when compared to the uppermost Capella Formation in section F.

CHAPTER III: SEDIMENTOLOGICAL ANALYSIS: ARCHITECTURAL ANALYSIS OF THE TECTONOSEDIMENTARY UNITS

The architectural analysis of the rather homogeneous succession of the Capella Formation consists of two major parts:

1) the definition of the architectural elements; i.e., the geometrical dimensions and variability of the different sediment bodies and their lithofacies development;
2) the reconstruction of the architectural framework formed by the architectural elements; i.e., the occurrences of the different types of sediment bodies within the different tectonosedimentary units.

Both aspects will be discussed in this chapter. Outcrop conditions, however, only allowed a detailed architectural reconstruction for TSU III.

1.- CHARACTERISTICS AND VARIABILITY OF THE GEOMETRY AND INTERNAL ORGANIZATION OF THE DIFFERENT SEDIMENT BODIES

1.1.- Introduction

In this section a general classification of the different types of sediment bodies is established, based on their geometry and lithofacies. The internal structural and sequential organization of the sediment bodies, as well as lateral facies variations are considered. Inferred primary sedimentary processes and ichnofauna are used for the environmental interpretation at the scale of the sediment body.

Two different macroforms can be recognized among the sediment bodies of the Capella Formation:

- *infilling forms*, which are related to channel deposits, and
- *upbuilding forms*, which are related to depositional lobes or sheet deposits.

The upbuilding forms often occur laterally to or interbedded with infilling forms. Upbuilding forms therefore must have a lateral connection with channelized forms. These upbuilding forms, e.g. depositional lobes and most of the sheet deposits, were formed at the mouth of the paleochannels and can be defined as terminal lobes. The progradational or retrogradational evolution of this system can be inferred from the superposition of the two forms. However, what is characteristic of the formation in this area, is the general lack of vertical interconnection between both types of forms, i.e., each terminal lobe is not directly related to the overlying (or underlying) channel. This indicates, besides a relatively high subsidence, that the terminal lobe was not deposited at the mouth of this same channel; more probably it was deposited by a different channel of the same channel system, which rapidly switched to another position lateral to the now abandoned terminal lobe. In the time the terminal lobe was abandoned and before its complete burial, the deposits at the surface of the sediment body were subjected to reworking : tidal and wave action, bioturbation (by animals and plants), mottling and other pedogenetic processes.

The general lack of vertical connection between terminal lobes and channel deposits points to a channel system where avulsion was a frequent process and channels were short-lived; the occurrence in the Capella Formation of totally or partially mud-filled channels supports this idea.

The former aspect together with the lateral variability in dimension, grain size and morphology of the channel deposits can be explained starting from a model of a wet alluvial fan as described by Schumm (1977). A wet alluvial fan consists of a system of multiple channels of distributary type; the surface discharge on the fan is divided into channels of different dimensions. The association of terminal lobes and channels can be explained according to the alluvial fan model proposed by Friend (1978), which is based on the description of Mujerki (1976). In the example described by Mujerki (1976) terminal fans (or lobes) appear near the inland termination of streams; they are formed by reduced discharge due to the continuous multi-furcation of distributary channels in an alluvial fan context. The terminal fans described by Mujerki were deposited in an inland flat plain without the influence of tectonics or marine processes. In the Capella Formation the terminal lobes were deposited at the distal part of a wet alluvial fan, where tides were often active; thereby indicating a fan delta for the distributary system responsible of the lobes. The thinner terminal lobes of the Capella Formation (sheet deposits) do not exhibit evidence of reworking by marine currents. They were probably deposited in a relatively flat area as a result of the reduced discharge of the streams, as proposed by Mujerki (1976). However, evidence of tidal influence is recognized within the thicker terminal lobes (or depositional lobes). In this case the depositional lobes could be formed where the streams reached the local base-level (more or less the sea-level). In contrast to the example described by Mujerki (1976), tectonics could have been a factor controlling the formation of depositional lobes in the Capella Formation. Small synsedimentary faults produced local gradient-changes; localized receptacle basins were created, in which sediment accumulated in the form of lobes.

Published in Geologie en Mijnbouw, 1985, volume 64: 221-231

1.2.- Geometry and lithofacies of sediment bodies in a tidally influenced alluvial area. Middle Eocene, Southern Pyrenees, Spain.

Margarita Cuevas Gozalo

Abstract

Four types of sedimentary macroforms are described from well exposed Eocene deposits in the Tremp-Graus Basin, Spain: channel deposits, depositional lobes, sheet deposits and composite forms of lobe and channel origin. The channel deposits comprise sediment bodies either formed by lateral accretion or by vertical aggradation. Mixed forms also occur. The form of the channel base is used for further subdivision of the channel bodies.

The association of the sedimentary macroforms indicates deposition on distal parts of alluvial fans. Influence of tides is suggested by the bipolar cross-stratification, reactivation surfaces and mud drapes occurring in some of the channel deposits and depositional lobes. At some locations intensive burrowing suggests marine conditions.

Introduction

The dominance of fluvial processes over tidal ones during sedimentation may conceal the evidence of tidal action in the resultant deposits. In ancient deposits this is stressed by the destructive influences of diagenesis.

A description and classification are given of sediment bodies deposited in a tidally influenced alluvial regime. Moreover, criteria for the recognition of tidal influence in such deposits are discussed.

The Capella Formation is of Middle Eocene (Lutetian) age (Crusafont et al. 1966; Garrido Megías 1968). Sediments were deposited in the South Pyrenean foreland basin (Tremp-Graus, Provs. Lérida-Huesca), northern Spain. During the Lutetian the central part of the Tremp-Graus basin was an embayment, protected from the open sea in the west by a barrier island system (Nijman & Nio 1975; Donselaar & Nio 1982). The alluvial sedimentation in the embayment was influenced by tidal action (Capella Formation, Cuevas Gozalo 1985).

The Capella Formation consists of terrigenous mudstones with interbedded conglomerate and sandstone bodies. The high proportion of mud allows individual geometries of the sediment bodies to be clearly recognized.

The present study is based on the exposures of the Capella Formation around the Isabena River and the Esera River valleys (fig. 1). The studied region presents many advantages for the comparative study of sediment bodies since many kinds of sediment morphologies occur in a small area (20 km x 5 km) with a large sediment thickness (up to 1000 m in the Isabena valley). Paleo-geographically, the studied sediment bodies belong to the same basin. They do not form isolated examples of different origin, but were deposited under comparable - if not similar - sedimentary conditions. A number of different types of sediment bodies are spatially related, some of them being the lateral or distal equivalent of each other.

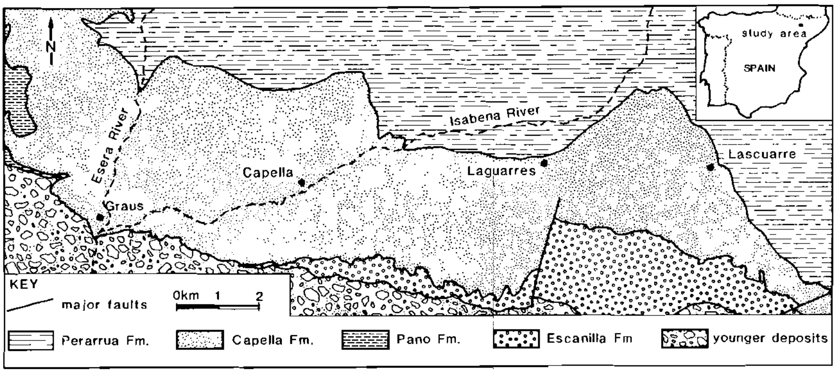


Fig. 1.- *Geographical and geological setting of the Capella Formation in the study area. The sedimentary formations are of Eocene age and young from left to right in the key.*

Mud deposits

Mud deposits form the bulk of the Capella Fm; they represent about 75% of the total volume. Individual mudstone intervals reach thicknesses of more than 20 m. They are brownish and commonly mottled; bluish-grey mudstones occur in the western part of the study area. The intraformational body-fossil content is generally poor. The deposits are, however, intensely bioturbated; vertical cylindrical burrows are predominant and can be observed at the boundaries with coarser grained deposits (fig. 2). The vertical burrowing suggests, together with other criteria to be mentioned later, intertidal conditions where organisms are frequently exposed to fluctuations in temperature, salinity and/or desiccation (e.g. Rhoads 1967, 1975; Crimes 1975; Basan & Frey 1977).

Recent examples of fluvial deposits suggest that the proportion of mudstones in the Capella Formation is too high to be interpreted as only the product of channel overbank deposits in an alluvial plain (cf. Wolman & Leopold 1957). Massive mudstone deposits in a purely continental Tertiary basin have been interpreted as the result of high suspension flows, generated as a consequence of general flooding in a humid alluvial fan (Diaz Molina 1979). Besides the alluvial processes mentioned, processes connected with the tidal environment, such as settling lag and scour lag, as well as the production of fecal pellets by

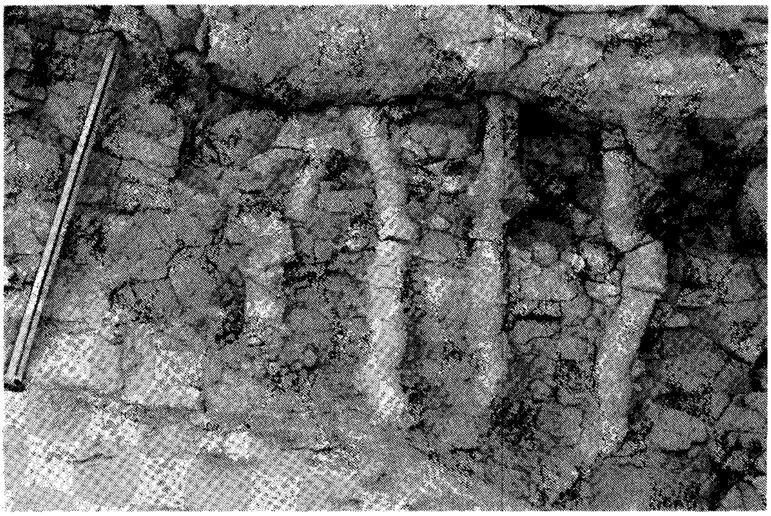


Fig. 2.- Vertical cylindrical burrows in mudstone deposits; upper deposits are fine sandstones. (Decapod burrows ?). Pencil length is 16 cm.

benthic animal populations may be responsible for the accumulation of large amounts of fine grained sediments (Van Straaten & Kuenen 1957; Postma 1967; De Mowbray 1980; Anderson et al. 1981).

Geometry and facies of the sediment bodies

In this study sediment bodies are defined as morphological entities, consisting of conglomerate, sand or coarse silt. The sediment bodies can be easily differentiated from the interbedded mudstone deposits.

The Capella Formation is characterized by a large variety of sediment body geometries; four major macroform types are distinguished:
- Channel deposits
- Depositional lobes
- Sheet deposits
- Composite forms

A further differentiation of these macroform types into various subtypes can be made as shown below.

Channel deposits

A channel is an elongate feature with negative relief, and is defined by a continuous erosive surface cutting into the underlying material; the erosive surface is present at any cross section. A continuous erosive lower surface and a general difference of lithofacies between channel deposits and substratum permit the channel deposits to be recognized.

Based on the internal organization, three types of channel deposits are recognized: bodies created by lateral accretion, by vertical aggradation and by both lateral and vertical aggradation. Vertical aggradation bodies can be subsequently subdivided into bodies with a single concave base, with a multiple concave base or with a base of low erosional relief.

(1) Lateral accretion bodies

The lateral accretion bodies found in the Capella Formation are typically 3-10 m thick. They are characterized by epsilon cross-bedding over a clearly erosive basal surface (A in fig. 3; fig. 4) (cf. Allen 1963). The epsilon cross beds fine upwards from medium sand to silt. Trough and planar cross-stratification are found at the base of the epsilon units; the cross strata often show long, well developed bottomsets (fig. 5). The upper part of the units commonly shows bioturbation and mottling. Where exposures afford a plain view of the units, a scroll-bar topography can be seen (fig. 6) (cf. Sundborg 1956; Puigdefábregas 1973).

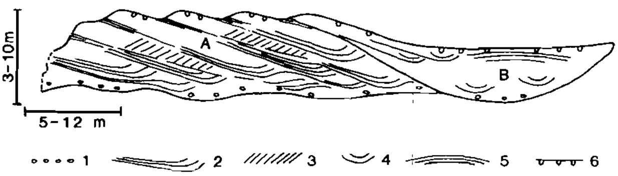


Fig. 3.- Lateral accretion body. A: lateral accretion units. B: channel fill unit; 1: lag deposit; 2: trough cross-stratification with well developed bottomsets (oblique section: o.s.); 3: planar cross-stratification (o.s.); 4: trough cross-stratification (transverse section: t.s.); 5: avalanche foresetting and low-angle cross-stratification (t.s.); 6: vertical cylindrical burrows.

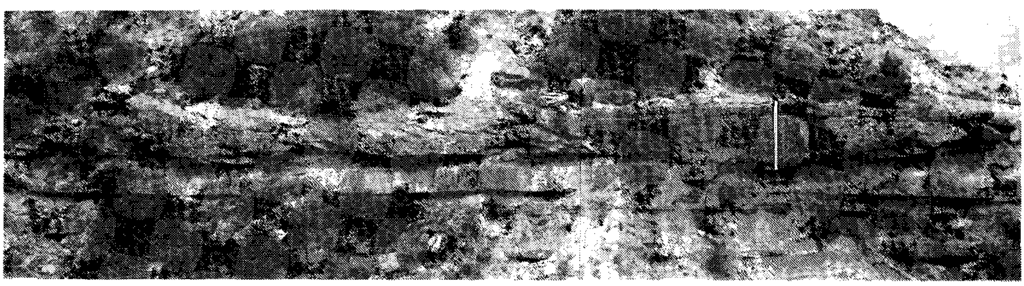


Fig. 4.- Lateral accretion body. Note the lateral accretion units in the left and the channel unit in the right. Scale length is 4.5 m.

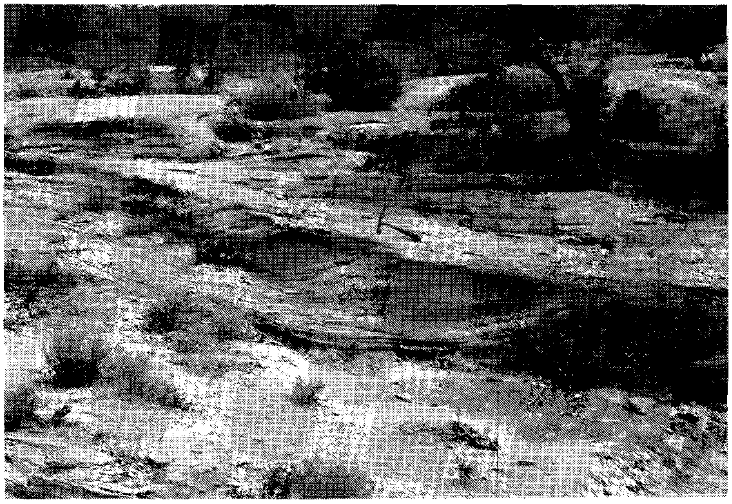


Fig. 5.- *Lateral accretion body: Trough cross-stratification with very well developed bottomsets at the lower part of the lateral accretion units.*

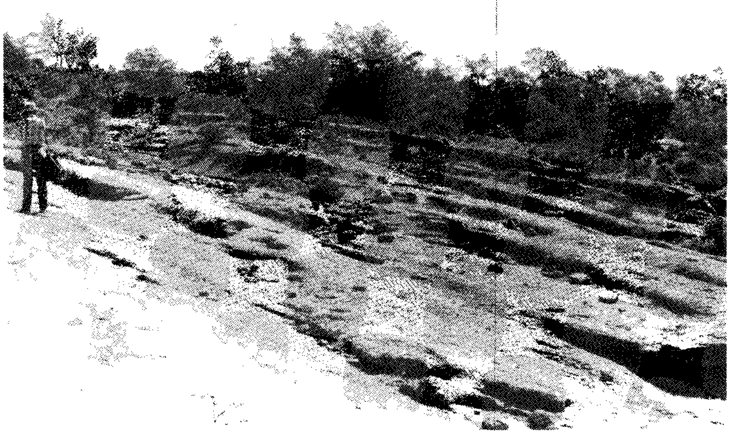


Fig. 6.- *Scroll bar morphology at the upper part of the lateral accretion units of a lateral accretion body.*

Epsilon cross-bedding alone is not sufficient evidence for lateral accretion; examples have been found in depositional lobes where they were formed as a result of lobe progradation and reworking by tidal currents (Cuevas Gozalo

1985). In the deposits described here, however, the association of epsilon cross-bedding with an erosional base, an upward fining in texture, the scroll-bar topography, and the presence of cross-stratification oriented parallel to or upslope the epsilon units provides strong evidence for lateral accretion on a channel point bar.

In addition, the channel relief is sometimes preserved next to the lateral accretion units (B in fig. 3; fig. 4). The channel is commonly filled with sandy deposits. A lag deposit is overlain by trough cross-stratification, avalanche foresetting or low-angle cross-stratification. Bioturbation is well developed, either as hypichnial burrows in the channel floor or as cylindrical vertical burrows at the upper surface of the fill (cf. Martinsson 1970).

Bipolar cross-stratified sandstones associated with reactivation surfaces, and mud drapes at regular intervals are found at the base of the lateral accretion units of some of the bodies (cf. De Raaf & Boersma 1971; Visser 1980; Terwindt 1981; De Mowbray & Visser 1984). Main paleocurrent direction is toward the land - in the northeast. The sediment bodies with these characteristics are interpreted as meandering tidal channel deposits.

(2) Vertical aggradation bodies with a simple concave base

These bodies show a simple concave lower surface and lateral wings (fig. 7 A, B and fig. 8 A).

In most cases, the sediment bodies consist of homogeneous, well sorted medium or fine sand, suggesting that vertical aggradation took place as a single event. Occasionally, low-angle and convex-upward cross-stratification can be observed, but often the deposits appear structureless. The channel depths, inferred from the thickness of the fills and the occurrence of banks covered and

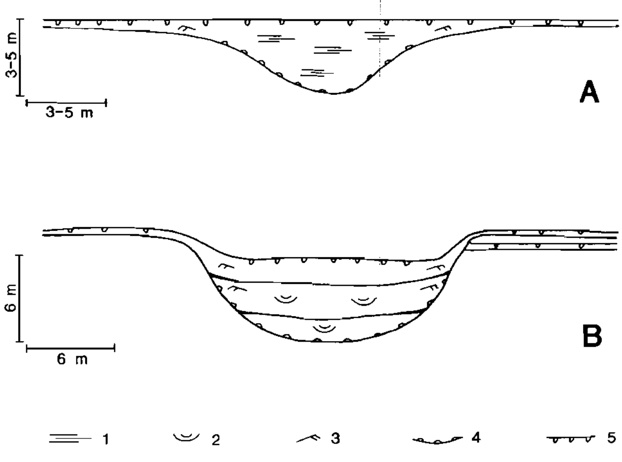


Fig. 7.- *Vertical aggradation bodies with a simple concave base. A: single filled. B: multifilled. 1: low-angle cross-stratification; 2: trough cross-stratification; 3: ripple cross-lamination; 4: hypichnial burrowing at the lower surface of the channel; 5: vertical cylindrical burrows.*

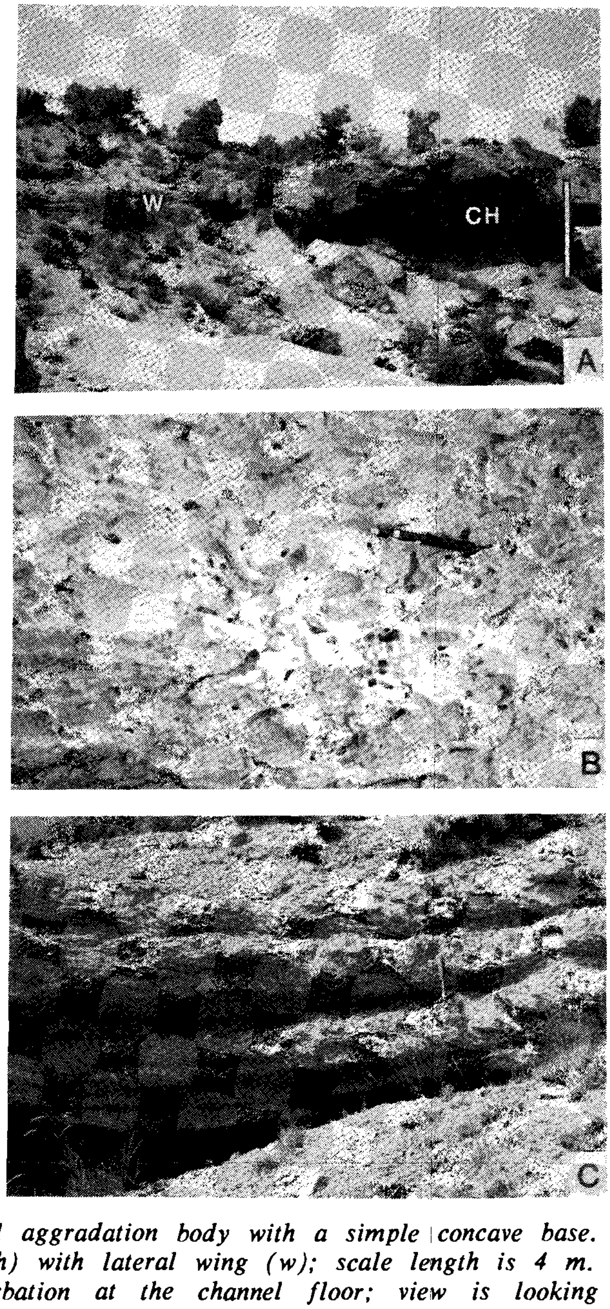


Fig. 8.- *Vertical aggradation body with a simple concave base. A: Single filled channel body (ch) with lateral wing (w); scale length is 4 m. B: Detail of the hypichnial bioturbation at the channel floor; view is looking at the external part of the channel deposit; pencil length is 14 cm. C: Detail of the multibedded lateral wing: hammer length is 28 cm.*

fossilized by overflow deposits, varied from 3 to 5 m. The lower surface of the channel deposit is covered with hypichnial burrows (fig. 8 B). The upper surface of the body is perforated by vertical, 1 cm wide cylindrical burrows.

In a few cases, the sediment bodies show evidence of repeated aggradation episodes. The differentiation of aggradation episodes is based either on the occurrence of lag deposits with large mudstone clasts or on the presence of silt or clay deposits at the top of medium to very fine sandy units (fig. 7 B). Individual aggradation units may be up to 2 m thick, the whole sediment body showing thicknesses of up to 8 m. Typical sedimentary structures are, from bottom to top, crescent casts at the lower surface, cross-stratification, scour and fill structures, low-angle cross-stratification, parallel lamination and ripple lamination. A textural fining and a transition to small-scale structures occur laterally towards the channel banks.

The lateral wings are 0.5 - 1.2 m thick wedges of fine sand and silt deposits connected to the main channel fill; they pinch out away from the channel axis. The wings consist of one or more intensely bioturbated beds (fig. 8 C); occasionally, ripple cross-lamination is preserved. The wings are the result of flow expansion over the banks, produced when the water discharge could not be confined in the channel. This situation may be caused by a reduction of the hydraulic radius after deposition within the channel; in this case the wings represent the lateral expansion of the flow prior to the final abandonment of the channel (Reading 1978, pag. 55; see also fig. 7 A). A lateral expansion of the flow can also be produced during flood periods; in this situation wing-like overbank deposits are formed; these wings are topographically higher than the last channel sandy fill and represent the levee deposits (fig. 7 B).

The bodies described correspond to the simple and multi-storey ribbon bodies of Friend et al. (1979). Moody-Stuart (1966) interpreted similar sediment bodies as being the product of low sinuosity channels. However, there is a similarity of geometry and facies between the bodies with concave base and the channel units associated to the lateral accretion bodies described above (B in fig. 3). The similarity between both channel types suggests that the bodies with concave base may also represent straight sections of meandering channels.

(3) Vertical aggradation bodies with a multiple concave base

These bodies are characterized by a multiple concave basal surface (fig. 9 A, B, C). This surface has an irregular shape and consists of deeper and shallower parts with a difference of depth up to a few meters. The dimension of the channel relief is comparable to the above-mentioned bodies with a simple concave base. The channel fill consists of fine to very fine grained sands with a maximum thickness of 4-6 m. The channel fill consists of several composite cosets over a lag deposit of mud clasts. Observed sedimentary structures are cross-stratification and ripple cross-lamination; opposite paleocurrent directions are occasionally observed in a single body. Bioturbation, burrowing and mottling are very intense in some of the examples. Different types of channel fills are shown in figure 9.

The configuration of the lower surface suggests that at low water stages some islands of muddy substratum emerged between the deeper parts of the channel; in this situation the channel belt consisted of an anastomosing system.

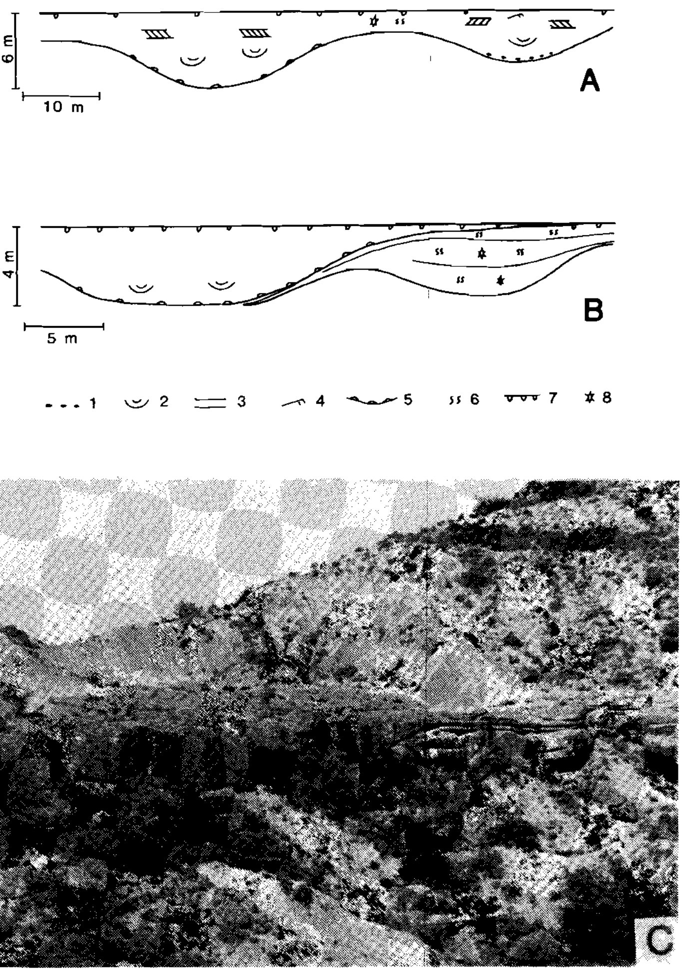


Fig. 9.- *Vertical aggradation bodies with a multiple concave base. The basal surface has an irregular pattern, consisting of deeper and shallower parts. The filling of the different depressions took place either simultaneously (A) or in different stages (B). In B the channel depression to the right was first infilled by very fine sandy deposits, and then highly bioturbated and mottled. The channel depression to the left was infilled in a later stage by fine sandy deposits. 1: mud lag deposits; 2: trough cross-stratification; 3: angular cross-stratification; 4: ripple cross-lamination; 5: hypichnial burrows; 6: intense bioturbation; 7: vertical cylindrical burrows; 8: mottling. C: Example of channel deposit represented in 9 B; scale length is 4 m.*

(4) Vertical aggradation bodies with a base of low erosional relief.

The third type of vertical aggradation channel fill is represented by bodies associated with the coarsest textures, i.e. conglomerates and coarse to very fine sands. These bodies have thicknesses ranging from 3 to 11 m, with lateral extents of several hundreds of meters. The basal surfaces of the bodies are erosive, but the erosional relief of the base has a low amplitude with respect to the body thicknesses (fig. 10). Within the sediment bodies, imbricated scours are observed; they correspond to the bases of smaller subordinate channels. The lower part of a complete sequence consists either of massive conglomerates with sand wedges or of planar cross-stratified conglomerates laterally associated with cross-stratified sands. Higher in the sequence cross-stratified medium sand and low-angle cross-stratified very fine sand with parting lineation occur. The upper part consists of ripple cross-laminated very fine sand, and parallel-laminated silt and clay; mottling is common. The massive conglomerates with sand wedges represent the active part of the channel belt, where probably more sand was deposited, but poorly preserved. The smaller subordinate channels were separated by coarse grained bars. Two types of bars are observed. The first type is erosional, formed by the incision of lateral channels into the previously deposited sediments. The internal structure of the bar is not conformable with the external geometry (fig. 11). The second type consists of a depositional bar formed by avalanching over a slip face; the internal structure in this case is conformable with the external geometry.

The well developed textural and structural sequence of the bodies indicates a progressive and general abandonment of the channel belt. The coarse grained bodies are interpreted as braided channels in an alluvial fan system; they are restricted to the eastern part of the studied area.

(5) Vertical and lateral aggradation bodies

These are coarse to very fine sand bodies with a maximum thickness of a few meters (fig. 12). The erosive lower surface of such a body is concave and slightly asymmetric. The channel deposit consists of several units. Each unit is characterized by a subordinate erosive concave lower surface formed by the partial erosion of the older channel deposits. The arrangement of the units within the channel body reveals a lateral migration of erosive and depositional activity. The scouring depths of the lower surfaces of the successive units decrease in the direction of lateral migration; the last unit splays over the channel margin.

Such a body represents the progressive abandonment of a channel by reduction of the width and depth of the thalweg. The reduction of the cross sectional area is due to a high deposition/erosion ratio in the channel. Deposition occurs mainly as vertical aggradation during lateral migration of the channel. Prior to the final abandonment, the active channel is not deep enough to confine the flow. Thus it expands beyond the channel and invades the adjacent plain.

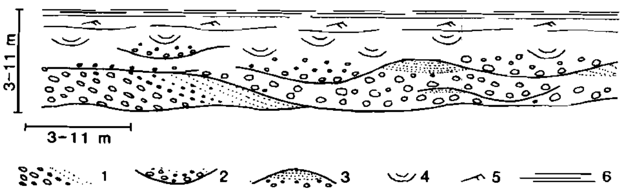


Fig. 10.- *Vertical aggradation body with a base of low erosional relief. 1: avalanche laminae in coarse-grained depositional bars; 2: imbricated, subordinate channels; 3: coarse grained inter-channel bars; 4: trough cross-stratification; 5: ripple cross-lamination; 6: parallel laminated silt and clay deposits.*

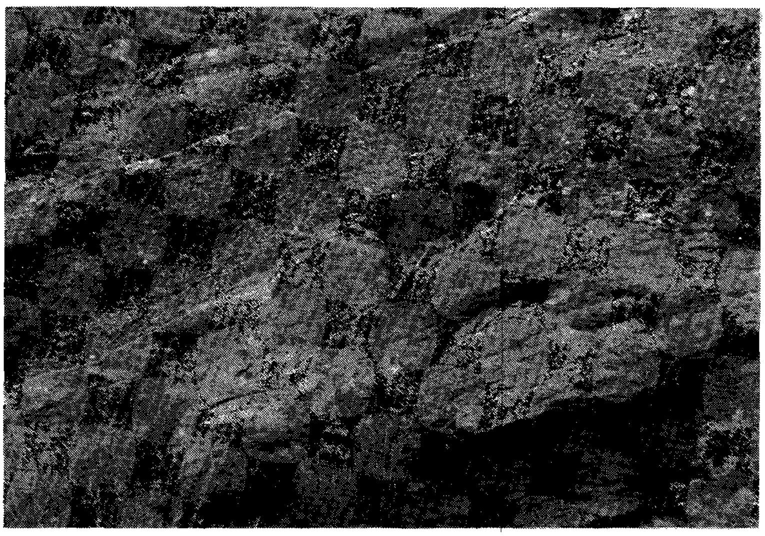


Fig. 11.- *(Vertical aggradation body with base of low erosional relief). Erosional sandy bar bounded by subordinate-channel conglomerates.*

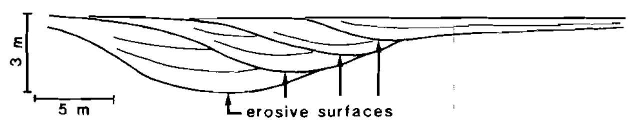


Fig. 12.- *Vertical and lateral aggradation body. The channel deposit consists of several units with erosive base: the units are obliquely arranged within the main channel.*

Depositional lobes

The lobes are bodies with a positive relief; they are not confined to channels. A description and classification of the lobes was made on the basis of geometry and lithofacies (Cuevas Gozalo 1985).

The lobes have thicknesses between 1-10 m; they are characterized by a wedging-out in longitudinal sections. The transverse section geometries range from tabular to plano-convex. The lower contacts of the lobes are either gradational with the muddy substratum, erosive to gradational, or only erosive. Two main lithofacies types are: *lithofacies A with horizontal bedding* and *lithofacies B with inclined bedding*. In transverse sections lobes with lithofacies B may display either a trough bedding pattern or a concentric convex-upward bedding pattern. Primary sedimentary structures include low and high-angle cross-stratification and ripple cross-lamination. The paleocurrent directions as derived from the sedimentary structures, are usually oriented in the dip direction of the inclined beds. Cross-stratification and ripple cross-lamination with countercurrent directions occur in some lobes, together with structures oriented downdip. The countercurrent structures form a ridge and swale morphology in the upper part of the lobes. They are interpreted to be the product of flood tidal currents. Vertical burrowing and mottling are common on the upper surface of the lobes.

Different interpretations are suggested from the different types of lobes: overbank splay lobes, channel terminal lobes, small delta lobes and spillover lobes, i.e. depositional forms in front of a tidally influenced channel (Cuevas Gozalo 1985).

Sheet deposits

The sheet deposits are sediment bodies with a relatively large extent (tens to hundreds of meters), and thicknesses from a few centimeters to about one meter. The sheet deposits consist of fine sand to silt. They occur either as isolated bodies embedded in mudstone deposits or as vertically stacked and amalgamated bodies (fig. 13). The lower surfaces of the sheet deposits are horizontal; the lower contacts are either gradational with the substratum, or irregular and slightly erosive. The upper parts of the bodies grade into the overlying finer sediments. Sedimentary structures are rarely present; ripple cross-lamination is occasionally observed. Bioturbation and mottling are very intense.

The sheet deposits are suggested to be the result of unconfined water flows moving down a slope. The sheet deposits as described here correspond to the sheetflood deposits of Hogg (1982) and Friend (1983). According to Hogg the hydraulic characteristics of the flow change from place to place on the slope, and also in time. Consequently, erosion and scouring at the sheetflood base either may occur or not; a sheetflood may end in a mudflow. These factors explain the variability of characteristics of the sheet deposits of the Capella Formation.

Sheetfloods are events of intermittent nature: they typically occur across fans, with erosion on the upper slopes and deposition in the lower parts (Hogg 1982). Sheetfloods occur at the downstream ends of channels (Bull 1972). It is assumed that the sheet deposits of the Capella Formation represent distal alluvial fan deposits.

From the point of view of the geometry, the sheet deposits have to be considered as a distinct macroform, occurring independently from the other macroform types. Nevertheless there is a strong similarity of features between sheet deposits, horizontal bedded lobes and the wings of the channel deposits with a simple concave base (compare fig. 8 C and fig. 13).

In the extreme case, the sediment bodies with coarser texture, large lateral extent and irregular, scouring basal surface can be considered as sheet deposits. The body thicknesses are from some decimeters up to 1 m. Occasionally preserved relicts of cross-stratification suggest that streamflood conditions occurred (cf. Hogg 1982).

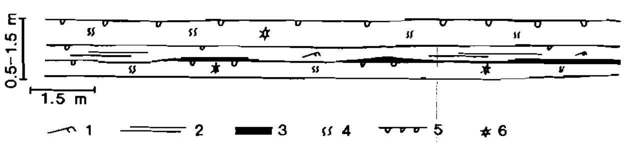


Fig. 13.- *Vertical stacked sheet deposits. The lower contact of the sheet deposit may be scouring or not. 1: ripple cross-lamination; 2: parallel lamination; 3: intercalated mud deposits; 4: intense bioturbation; 5: vertical cylindrical burrows; 6: mottling.*

Composite forms

The composite forms are defined by an association of two of the earlier described geometrical types, channels and lobes. An interpretation of the relationships between channel and lobe forming processes is necessary to describe the composite geometries. The succession or the alternation of both sedimentary processes generates relatively simple forms in which the identification of lobe and channel geometries is possible. However, if both sedimentary systems are simultaneously active, a body with mixed geometry is generated.

The following three relationship types can be recognized:

a) Lateral infilling of a slightly active channel by a lobe. The channel geometry is preserved; the channel deposits are fossilized by the lateral influx of sediment (fig. 14).
b) Alternate episodes of lobe and channel activity in front of the lobe. The lobe sedimentation generates a morphology that determines the position of the channel. In turn, the channel relief controls further progradation of the lobe (fig. 15).
c) Simultaneous activity of lobe and channel. The resulting form is more complex. There are no boundaries between channel and lobe; the sediment

supplied by the lobe is immediately transported along the channel. The lobe progradation provokes the lateral migration of the complex channel-lobe (fig. 16).

The composite forms represent the association of a lobe and a channel fill showing paleocurrent directions oblique to perpendicular with respect to each other. The lobe may have relatively little importance in relation to the channel (fig. 14); then it is interpreted either as an overbank splay lobe or a small terminal lobe derived from an adjacent channel. Whenever the amount of sediment laterally supplied into the channel is large, the channel can be plugged and backfilling occurs.

In other cases (fig. 15 & 16), lobe and channel are of comparable size and their deposits are laterally related. Lobe forming processes and channel forming processes are closely related to the extent that both geometries cannot be distinguished individually (fig. 16). The model of spillover lobes (terminal lobes in front of a tidally influenced channel) with marginal channels may be applied to these composite forms (fig. 16 in Cuevas Gozalo 1985).

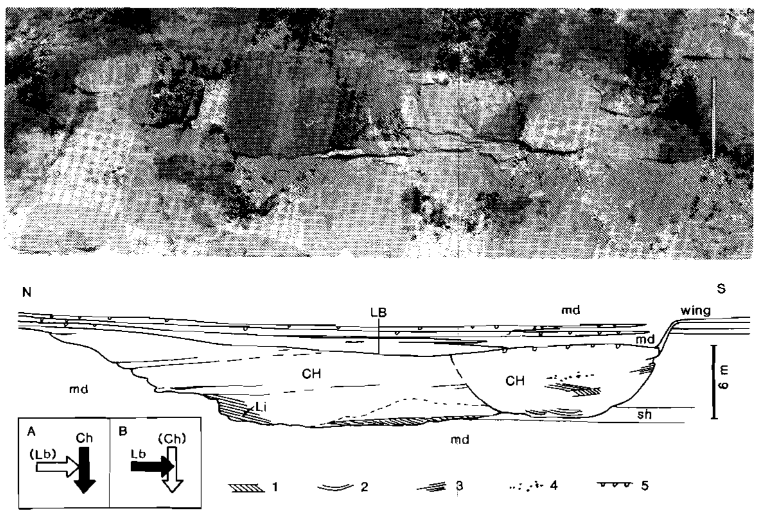


Fig. 14 A & B.- *Composite form produced by lateral infilling of a slightly active channel by a lobe. The channel deposit (CH) was formed during an active channel period (site A, Ch.). The channel deposit is covered by a horizontally bedded lobe (LB) coming from the north (left) (site B, Lb.). The lobe beds wedge out before reaching the southern (right) channel bank. Initial lateral infilling (Li) of the channel can be observed at the northern lower channel bank; the lateral fill consists of avalanche cross-laminae. Sedimentary structures of the channel deposit are planar (1), trough (2) and low-angle (3) cross-stratification. Lag deposits consisting of mud clasts (4) occur in the channel fill. The lobe deposits are highly bioturbated and burrowed (5). md: mudstone deposits; sh: sheet deposits. Scale length is 6 m.*

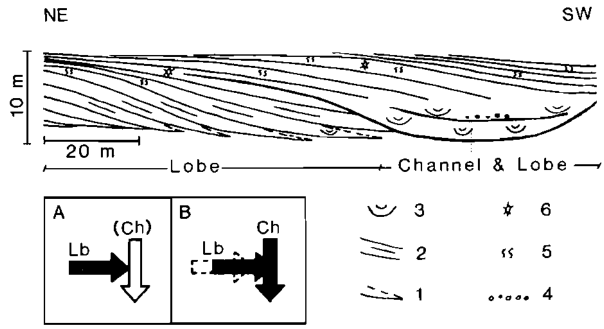


Fig. 15.- *Composite form created by the alternation of lobe activity and channel activity in front of the lobe. In a first phase (A) the lobe (Lb) progrades from NE to SW (left to right); the channel ([Ch]) is practically inactive and only manifested by some reworking structures at the base of the lobe. In a second phase (B) the channel is highly active and erodes the base of the prograding lobe. Lobe (Lb) and channel (Ch) are simultaneously active; the lobe ends in the channel; the channel fill consists partially of lobe deposits. 1: transition to finer deposits; 2: inclined stratification; 3: trough cross-stratification; 4: lag deposits; 5: intense bioturbation; 6: mottling.*

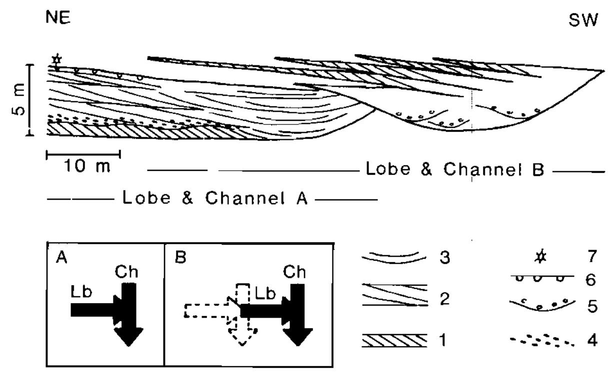


Fig. 16.- *Two composite forms as the result of lateral migration. Phase A: The simultaneous activity of a lobe (Lb) and a channel (Ch) creates a composite form. The lower contact of the composite form is erosive in response to the channel scouring. Paleocurrents in the composite form exhibit two main directions: the first one is from left to right and corresponds to the lobe deposits (Lb); the second one is perpendicular to the picture and corresponds to the channel deposits (Ch). A thinning of the lobe deposits towards the channel can be discerned. The lobe-channel complex (A) is active in the position A until the channel relief is infilled. As a consequence of this, the lobe-channel complex migrates in the direction of progradation of the lobe to a new position; a new phase B starts. 1: angular and 2: low-angle cross-stratification in longitudinal section; 3: low-angle cross-stratification in transverse section; 4: low-angle cross-stratification in mud pebbles (bottomsets; long. sect.); 5: subordinate channels with lag deposits; 6: vertical cylindrical burrows; 7: mottling.*

Conclusions.

The sediment bodies of the Capella Formation are classified on the basis of the external geometry and the internal organization. Four major types can be recognized: channel deposits, depositional lobes, sheet deposits and composite forms of lobe and channel. A further differentiation of these groups is based on similar criteria.

The understanding of the relationship between the geometrical properties and the internal organization of the sediment bodies provides information on the mechanism of their formation.

The morphology of channels was considered to be controlled by the bank strength (Schumm 1968; Friend et al. 1979). The banks of most of the channel deposits of the Capella Formation have similar lithology (mudstones), but different morphologies are inferred. The morphology of the channels was influenced by factors as the hydrology, lithology and tectonic activity of the source area, the relative proximity of the channel to the source area and the position of the channel with reference to base level and flow strength.

The deposits of the Capella Formation, channel deposits, depositional lobes, sheet deposits and composite forms were deposited either from confined or unconfined flows. The results of both types of flow are spatially associated: depositional lobes and sheet deposits (unconfined flow) are related to the channel deposits (confined flow) either as lateral products (overbank splay lobes) or as distal products (sheetflood deposits, terminal lobes). Both confined and unconfined flow and their resulting products may be integrated in a model of deposition on the lower, tidally influenced part of an alluvial fan system.

Among the channel deposits two main textural groups are observed. The first group is coarse, conglomeratic and is associated with braided channel deposits. The second group is represented by fine to medium sandy deposits. It includes most of the channel deposits described, comprising meandering, straight and anastomosing channels. In general they are of less lateral extension than those of the first group. The textural and morphological separation of these two groups of channel deposits suggests that they belong either to different areas (proximal to distal) of the same alluvial fan system or to different alluvial fan systems.

Opposite paleocurrent directions accompanied by reactivation surfaces and mud drapes at regular intervals are present in some of the channel deposits and depositional lobes. These features are interpreted to be the result of tidal influence during sedimentation. The vertical burrowing of the mud deposits suggests that the mud plain in which channel deposits and lobes were formed was also affected by intermittent subaquatic conditions. An intertidal environment is inferred from these deposits; supratidal and subtidal conditions could be locally dominant in respectively elevated or deeper parts.

Acknowledgments

The author wants to thank the members of the Comparative Sedimentology Division, especially S.D. Nio, R. Donselaar, A. van Gelder and T. de Mowbray for their help in the field, discussions, and critical reading of the manuscript. Without the moral and scientific help of M. Diaz Molina this work could not have been accomplished. This study, which forms part of the author's Ph.D. research carried out at the University of Utrecht, The Netherlands, was written during the tenure of a postgraduate scholarship of the Spanish Ministry of Education which is gratefully acknowledged. The author is also indebted to the Spanish Foreign Office and the Ministry of Industry and Energy for their financial assistance at the beginning of this study.

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1.3.- Sedimentary lobes in a tidally influenced alluvial area, Capella Formation, Tremp-Graus Basin, Southern Pyrenees, Spain.

Margarita Cuevas Gozalo

Abstract

Among the coarse clastic deposits of the mainly alluvial Capella Formation several sediment bodies with characteristics of depositional lobes are recognized. These lobes are characterized and distinguished from other macroforms by their geometry, basal contact and lithofacies.

A classification of the lobes is proposed on the basis of the geometry and the lithofacies. The most common *geometrical type* is tabular to wedge-shaped; a second geometrical type is sigmoidal. Two main *lithofacies types* are distinguished on the basis of the bedding pattern: a horizontally and an inclined bedded type.

The lobes were deposited in the lower part of an alluvial fan system. The sedimentation of some of the lobes was influenced by tidal currents. Several models of sedimentary lobes are recognized: overbank splay lobes, terminal lobes, debris flow lobes and spillover lobes.

Introduction

The Tremp-Graus Basin is located in the Southern Pyrenees, in the provinces of Lérida and Huesca, Spain (fig. 1). During the Paleocene this basin constituted a marginal trough, to the south of the tectonically active Pyrenean axial zone.

Recent studies of the basin (Nio et al. 1984) have shown the sedimentation pattern to be complex, probably tectonically controlled, with an alternation of transgressive and regressive sequences.

The Capella Formation was defined by Garrido Megías (1968) as a continental formation of Late Lutetian age. During the Lutetian the central part of the Tremp-Graus Basin was a large embayment (Nijman & Nio 1975), protected from the open marine part of the basin in the west by a barrier island system (Donselaar & Nio 1982). Locally the Capella Formation grades upwards into a lagoonal facies (Garrido Megías & Ríos Aragües 1972). Nijman & Nio (1975) interpreted the Capella Formation as the fluviatile upper part of a general regressive megasequence. Generally, the Capella Formation consists of alluvial sediments deposited under the influence of tidal action (Cuevas Gozalo 1984).

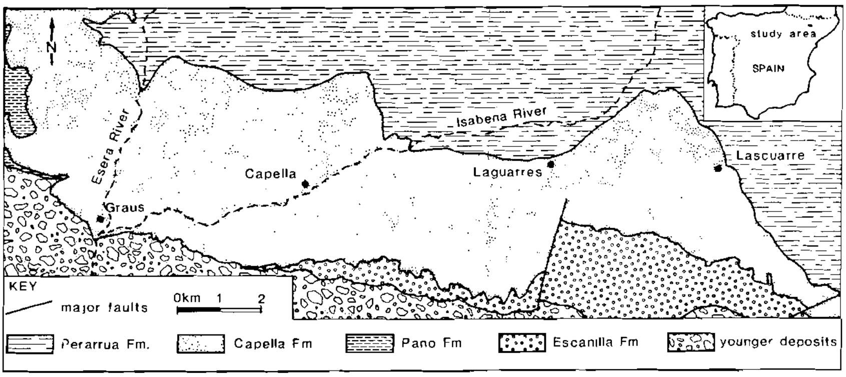


Fig. 1.- *Geographical and geological setting of the Capella Formation in the study area. The sedimentary formations shown in the legend form a sequence and young from left to right.*

The Formation consists mainly of fine-grained deposits, i.e. ochre, brown or red terrigenous mudstones. Most of the interbedded sandstone bodies are laterally discontinuous. Towards the top of the Formation, however, more continuous, sheet-like sandstone bodies appear. Conglomerates also occur, but they are mainly restricted to the eastern part of the basin.

The studied area, around the Isábena and the Esera rivers, province of Huesca (fig. 1), contains many good exposures which allow a three-dimensional study of the sediment bodies.

The study of the sandstone and conglomerate bodies in both transverse and longitudinal sections allows the recognition of *two main morphological types:*

- Sediment bodies with a distinct *channel geometry* (infilling forms), which result from deposition over a continuous erosive concave surface. These channels are formed and infilled by alluvial and tidal systems. Besides the simple type which has a concave-planar geometry in transverse cross section (ribbon type of Friend et al. 1979) other infilling forms are also present as vertical aggradation bodies of large lateral extent and as lateral accretion bodies.
- Sediment bodies with a *lobate geometry*, i.e. lobes (upbuilding forms).

This paper will discuss the geometry, lithofacies and genesis of the latter.

General geometry and lithofacies of the lobes

Several types of lobes can be distinguished on the basis of their geometry in longitudinal (axial) section as well as on their lithofacies.

The geometry of the lobes is characterized by a wedging out in downcur-

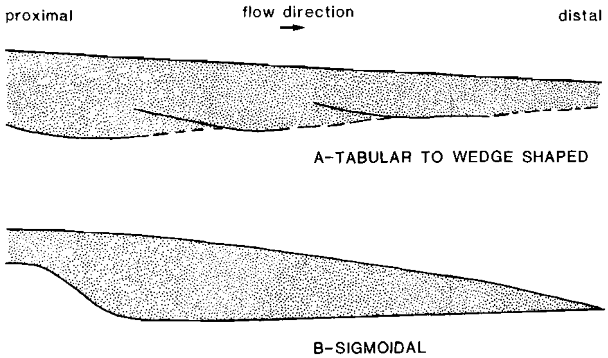


Fig. 2.- *External geometries of the lobes in axial section. Two main types are differentiated: (A) tabular to wedge-shaped and (B) sigmoidal.*

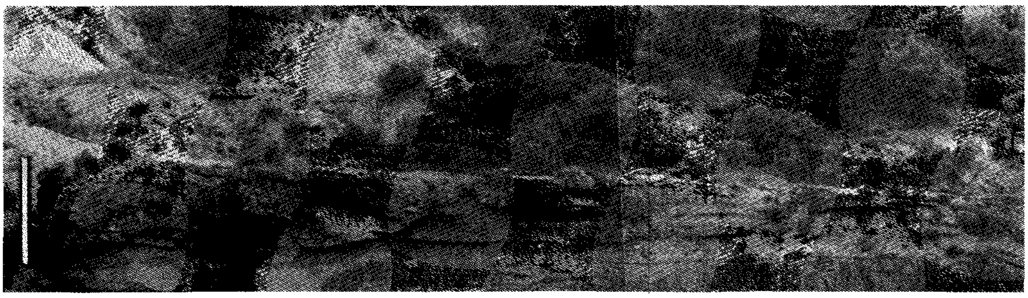


Fig. 3.- *Wedge-shaped lobe in axial section; scale length is 5 m.*

rent direction (longitudinal section). *Two types* can be distinguished: *tabular to wedge-shaped* and *sigmoidal* (fig. 2).

Lobes with a *tabular to wedge-shaped* geometry are characterized by a continuous decrease in thickness from the proximal to the distal part (fig. 2,A and 3).

Lobes with a *sigmoidal* geometry are characterized by a thickening in their proximal part and a progressive thinning towards the distal part (fig. 2,B).

In transverse section the geometry of both types of lobes is more complex. The proximal parts show a tabular to slightly plano-convex geometry; the distal part on the other hand is plano convex (figs. 4 and 5).

Besides the two lobe geometries, *two major lithofacies types* can be differentiated on the basis of the internal bedding (figs. 6 and 7 and table 1):

Lithofacies type A is characterized by horizontal bedding, formed by vertical stacking of subhorizontal, tabular to wedge-shaped beds, whereas lithofacies type B shows inclined bedding, formed by lateral superposition of beds. The

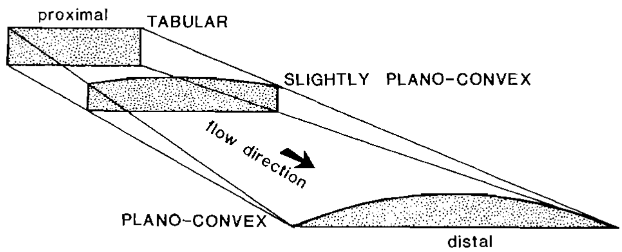


Fig. 4.- *External geometries of the lobes in transverse section. A gradual change can be observed from a tabular geometry in the proximal part of the lobe to a plano-convex geometry in the distal part.*

Fig. 5.- *Plano-convex geometry of a lobe in transverse section of the distal part. Scale length is 1 m.*

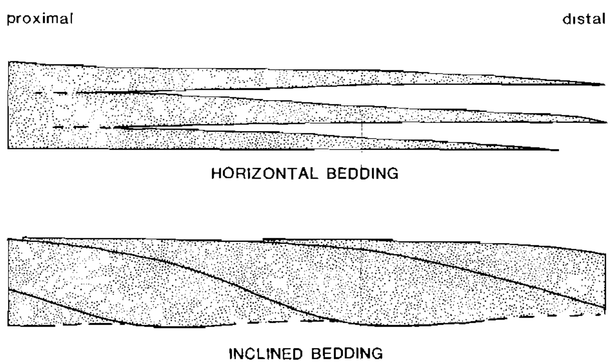


Fig. 6.- *The two main lithofacies types of the lobes, based on the bedding pattern: Lithofacies A or horizontal bedding, Lithofacies B or inclined bedding.*

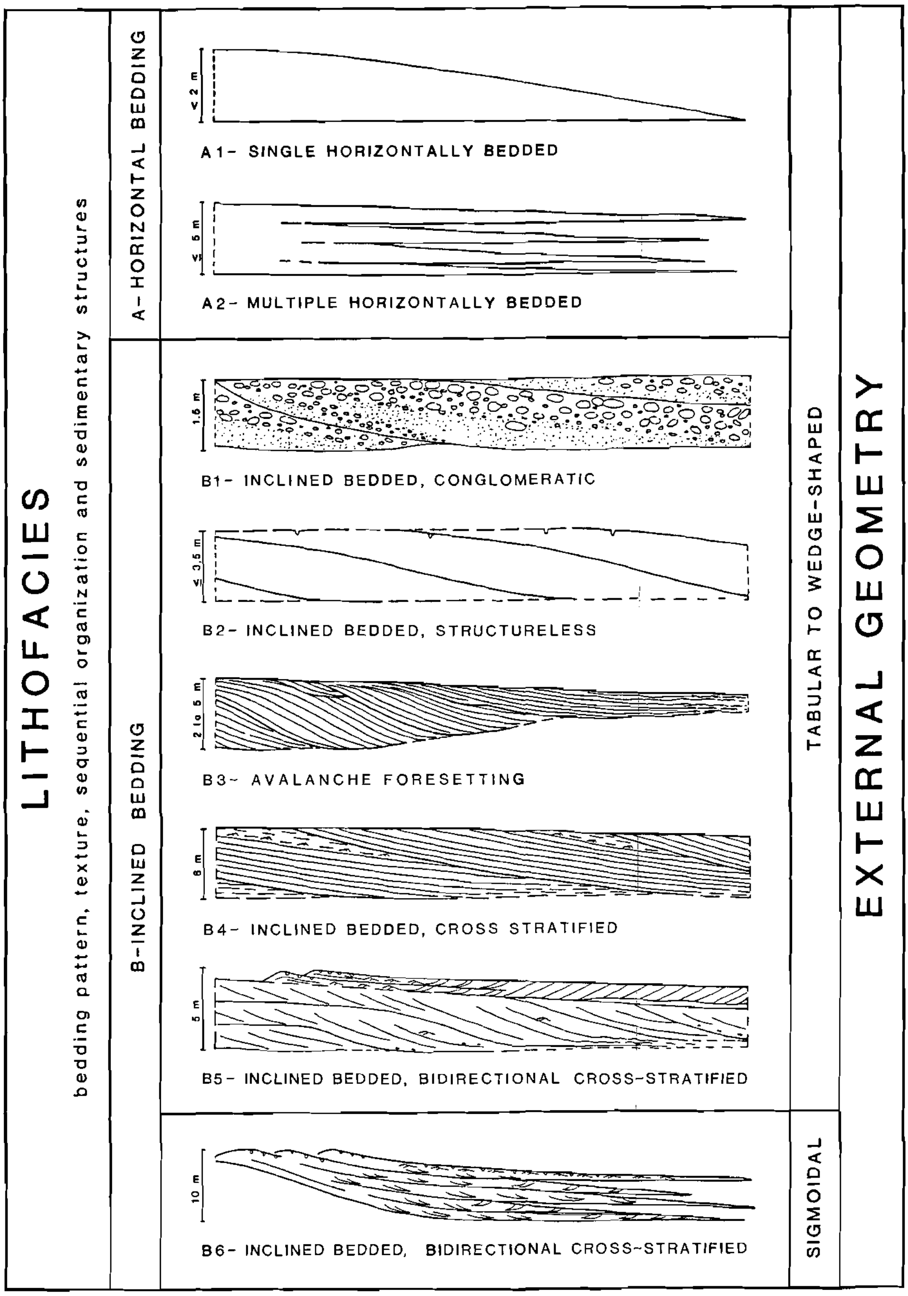


Fig. 7.- *Classification of the Capella Formation lobes. The characteristics used for the classification are the external geometry in axial section and the lithofacies. The horizontal scale is slightly reduced relative to the vertical scale.*

Lobe type	Geometry (1)	Lower & upper Boundaries (2)	Lithofacies		
			Bedding pattern (1)	Texture	Grain size sequence
A1- Single horizontally bedded	L: wedge-shaped T: plano-convex or multiple plano-convex	Lo: slightly erosive to gradational	single horizontal bed	fine sand to silt or very coarse sand	fining upwards
A2- Multiple horizontally bedded	L: wedge-shaped T: tabular to plano convex	U: gradational to finer deposits	vertical stacking of single horizontal beds	fine sand to silt	Repeated fining upwards
B1- Inclined bedded, conglomeratic	tabular	Lo: erosive U: sharp	inclined bedding	sand and a poorly sorted mixture of sand, pebbles & cobbles	coarsening-upwards
B2- Inclined bedded, structureless	L: tabular T: tabular to slightly plano-convex	Lo & U: gradational	inclined beds, tangential to lower & upper lobe boundaries	fine to very fine sand	coarsening fining upwards
B3- Inclined bedded; avalanche foresetting	L: wedge-shaped T_p: tabular T_d: plano-convex	Lo: erosive to gradational U: sharp or gradational	L: inclined bedding defined by sets of avalanche laminae T_p: trough bed. T_d: concentric		coarsening-fining upwards
B4- Inclined bedded, cross-stratified	L: wedge-shaped T_p: tabular T_d: slightly plano-convex	Lo: erosive to gradational U: gradational	L: inclined bedding T: trough bedding Ridges in the upper part (B5)	medium sand to silt	coarsening-repeated fining-upwards
B5- Inclined bedded, bidirectional cross-stratified		Lo: erosive to gradational U: sharp			
B6- Inclined bedded, bidirectional cross-stratified, sigmoidal	L: sigmoidal T_d: plano-convex	Lo: erosive U: gradational	L: inclined bedding; ridges in the uper part T_d: concentric bedding		alternating sand and silt deposits; upper part is fining upwards.

	Primary sedimentary structures	**Countercurrent sedim. struct.**	**Bioturbation & Mottling**
	?		very intense
	matrix supported conglomerate		
	?		very intense
	avalanche foreset lamination	occasionally tangential cross stratification and ripple cross-lamination	occasionally on the upper surface
	low-angle cross-stratification; subordinate cross-stratif. & ripple cross-lamination	very rarely cross-stratif. & ripple cross-lamination	intense on the upper surface
	lag deposits; cross-stratification	cross-stratif. & ripple cross-lamination in the upper part	very intense on the ridges
	cross-stratification	very common: cross-stratif.; ripple cross-lamination on the upper surface	very intense on the upper surface and ridges

Table 1.- *Geometry and lithofacies of the sedimentary lobes in the Capella Formation. The different lobe types represented in figure 7 are indicated here in the left column. (1) L: longitudinal section; T: transverse section, Tp: of the proximal part, Td: of the distal part. (2) Lo: lower boundary; U: upper boundary.*

beds are inclined with respect to the upper and lower surfaces of the lobe. The lower boundaries of the beds occasionally have an erosive character: the erosive surfaces are generally well developed in the lower part of the lobe. Generally, this erosive lower boundary passes distally into a gradational boundary with a mudstone substratum (figs. 8 and 9).

Lithofacies type B is the most common and shows a very variable bedding pattern in transverse section, depending on the type of lobe and on the position within the lobe. The proximal parts are characterized by a 'trough' bedding pattern, while a 'concentric' bedding pattern occurs in distal parts (figs. 10 and 11).

The tabular to wedge-shaped lobes are associated with either lithofacies type A or lithofacies type B. The sigmoidal lobes display only lithofacies type B.

Lithofacies of the tabular to wedge-shaped lobes

The tabular to wedge-shaped lobes display either the lithofacies type A (horizontal bedding) or the lithofacies type B (inclined bedding) described above. A subdivision of these lithofacies types is made on the basis of texture and sedimentary structures (fig. 7 and table 1). Unless an other composition is indicated, the lobes consist of sand.

Lithofacies A1. Single, horizontally bedded lobes

Lithofacies A1 consists of a single horizontal bed up to 2 m thick, and is characterized by a plano-convex or multiple plano-convex geometry in transverse section; the bed wedges out towards the distal part.

Lithofacies A2. Multiple, horizontally bedded lobes

Lithofacies A2 consists of several vertically stacked horizontal wedge-shaped beds. The proximal part of the lobes generally shows an amalgamation of beds. The distal pinching out of a bed is accompanied by a fining of the grain size. The lobes have an average thickness of 1-2 m with a maximum of 5 m.

Lithofacies B1. Inclined bedded, conglomeratic lobes

These lobes are about 1.5 m thick and a few tens of meters in longitudinal extent. They are built up of a series of inclined, partly overlapping conglomeratic beds, each up to 1 m in thickness (fig. 12 a). The conglomeratic beds show reversed grading: the lower part is sandy, the upper part consists of a poorly sorted matrix-supported conglomerate (fig. 12 b).

Lithofacies B2. Inclined bedded, structureless lobes

The lobes of lithofacies type B2 are typically about 2 m thick with a maximum of 3.5 m; the longitudinal extent is several tens of meters. The sandy lobes grade down- and upwards into silty deposits. Primary sedimentary structures are absent, probably because of intense bioturbation.

Lithofacies B3. Avalanche-foresetting lobes.

The dominant sedimentary structure is avalanche foresetting. The dip

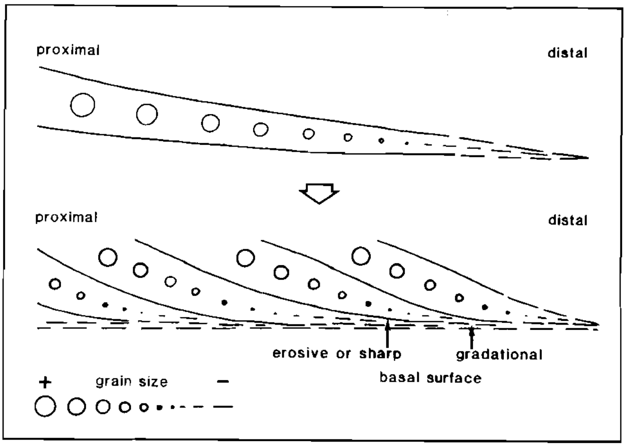
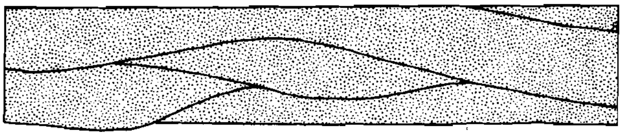


Fig. 8.- *Erosive to gradational lower contact formed by the superposition of beds with fining texture in downcurrent direction.*

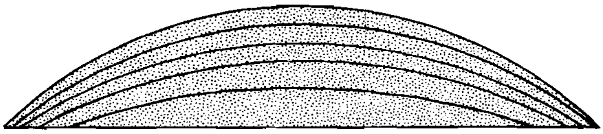
Fig. 9.- *Change in the nature of the lower contact of a lobe in the downcurrent direction, from erosive (E.B) to gradational (G.B). Hammer length is 28 cm.*

angles of the avalanche foreset laminae are 8° to 11° in the upper part of the lobe increasing to 33° in the middle part. The bottomsets are angular to tangential (fig. 7, B3).

Together with a thinning of the lobe towards the distal part, there is a change from high to low inclination in the foreset laminae. The structures



TROUGH BEDDING



CONCENTRIC BEDDING

Fig. 10.- *Two characteristic bedding patterns of lithofacies type B in transverse section, the trough bedding pattern and the concentric bedding pattern.*

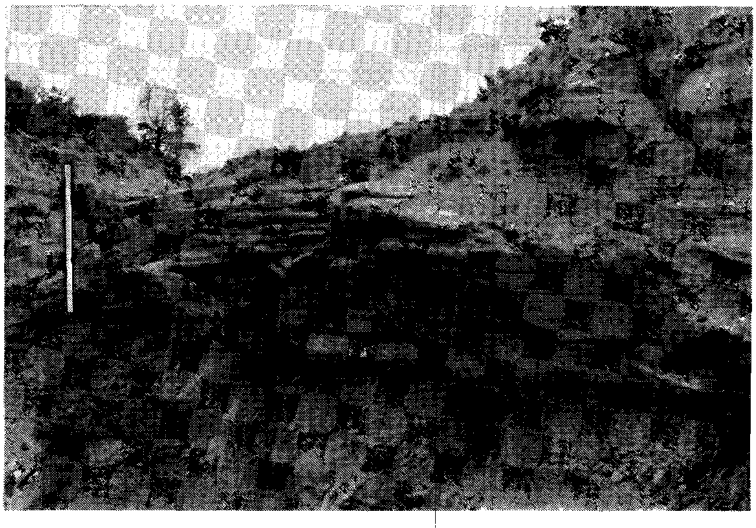


Fig. 11.- *Transverse section of the distal part of a lobe of lithofacies B6. Note the plano-convex geometry and the concentric bedding pattern. Scale length is 7 m.*

finally grade into ripple cross-lamination.

Counter-current structures were also observed. They appear as sets of tangential cross-stratification, less than 20 cm thick, and ripple cross-lamination. Evidence of counter-current cross-stratification is restricted to the middle and lower part of the avalanche foresetting, whereas ripple cross-lamination is more widely distributed.

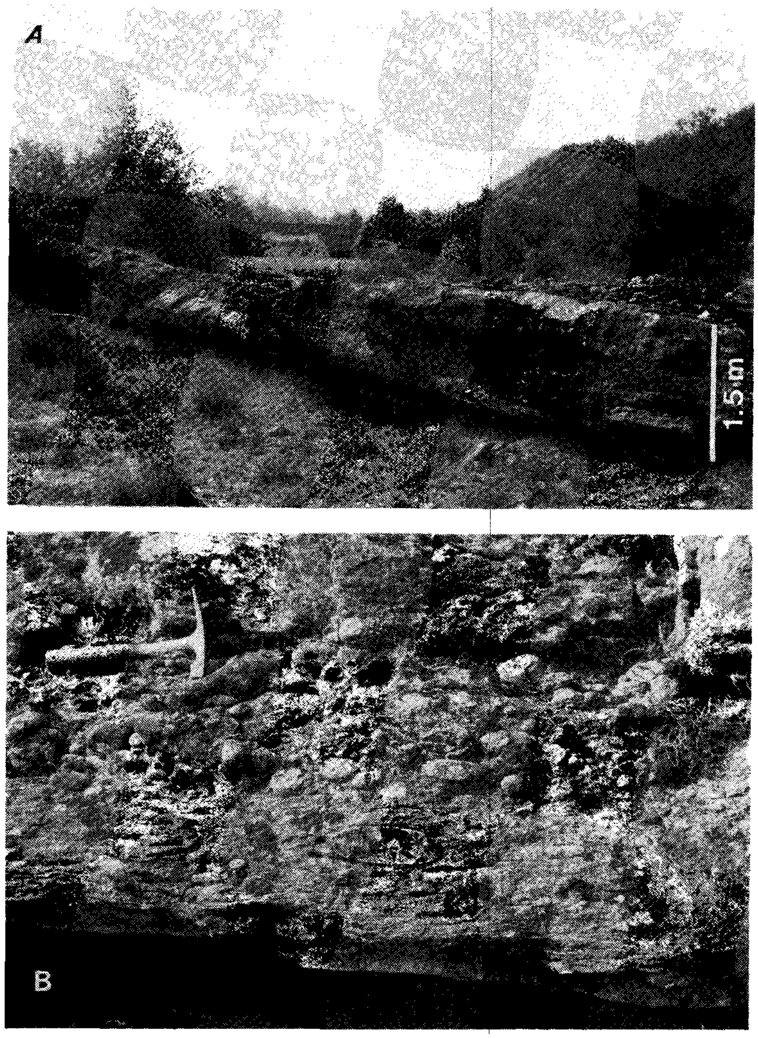


Fig. 12.- *Conglomeratic lobe of lithofacies type B1. A: Note the tabular geometry and the inclined bedding (left). B: coarsening upward sequence from sand to poorly sorted, matrix-supported conglomerate. Hammer length is 28 cm.*

One of the studied avalanche foresetting lobes exhibits, in transverse section, a shallow concave erosive form in the upper part (fig. 5). This erosive form represents the feeder channel that prograded over the lobe. The erosive surface is overlain by low-angle cross-stratification. The erosive surface passes laterally into avalanche foreset laminae.

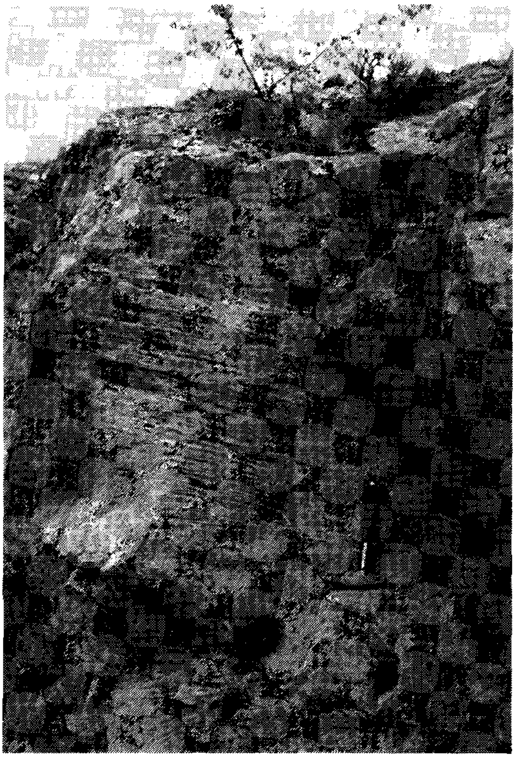


Fig. 13.- *Low-angle cross-stratification in a lobe of lithofacies type B4. Hammer length is 30 cm.*

Lithofacies B4. Inclined bedded, cross-stratified lobes

Lobes of lithofacies B4 are typically up to 6 m thick with a longitudinal extent from several tens of metres up to a few hundred metres.

In transverse section the proximal part of the lobe displays a trough bedding pattern. The troughs are several metres wide and up to 2 m thick; in longitudinal section they can be seen to be built up of low-angle cross-stratification (dip angles 2°-12°) (fig. 13). Subordinate structures are trough cross-stratification and ripple cross-lamination. The sedimentary structures indicate transport parallel to the wedging direction of the lobe. The grain size and the thickness of the sedimentary structures decrease upwards within each bed.

Lithofacies B5. Inclined bedded, bidirectional cross-stratified lobes

In longitudinal section the bedding surfaces of this lobe type show a slight inclination at the upper part of the lobe. They tend to steepen towards the lower part of the lobe and are occasionally erosive at its base. The erosive character of the bedding surfaces passes into a gradational contact with the muddy substratum in the downcurrent direction; this change is associated with a textural fining of the deposit (fig. 9).

A transverse section of the proximal part of the lobe, however, exhibits a trough bedding pattern in the lower part. The troughs are relatively small, have an erosive base and fine upwards from medium sand to silt. Lag deposits of small pebbles and oncolites are common. Oncolites occur both as imbricated fragments, i.e. clearly transported, or as more extensive planar fossil bodies, horizontally interstratified in the lag, suggestive of an in-situ formation. Oxidized plant fragments are present in the silt deposit.

The dominant sedimentary structure is angular cross-stratification; it is generally oriented in the direction of the bed inclination. Cross-stratification and ripple cross-lamination with opposite orientations are also present. They occur mainly at the top of the lobe, overlying sets of downdip oriented cross-stratification.

The sets of these counter-current structures pinch out in the migration direction. The thinning of the sets is associated with a transition from angular cross-stratification in the thicker part of the set into tangential cross-stratification, ripple cross-lamination and poorly developed parallel lamination in the thinner part (fig. 14). Mud laminae and repeated discontinuity surfaces are commonly intercalated within the foresets of these structures. The amalgamation of several thin, poorly laminated sets creates a form with a convex-up morphology. These convex forms have an elongation parallel to the strike of the lobe surface and a slip face facing in the downdip direction. Parallel lamination is poorly preserved in these forms because of the intense bioturbation. In the field these convex-up forms are arranged parallel to each other revealing a ridge and swale topography. These features show similarities to the scroll bar topography described at the top of point bar deposits (Sundborg 1956, Puigdefábregas 1973, Nami 1976, Nanson 1980, Van der Meulen 1982).

Lithofacies of the sigmoidal lobes

The sigmoidal lobes have been found to be always associated with lithofacies type B (inclined bedding) (fig. 7 and table 1). A characteristic sigmoidal lobe is described here:

Lithofacies B6. Inclined bedded, bidirectional cross-stratified lobes.

This type of lobe has a sigmoidal geometry in longitudinal section. The contact with the substratum is erosive. The beds are inclined and, characteristically, some of them do not grade into the finer deposits of the distal part but pinch out with sharp contacts.

The lobe shows a plano-convex geometry with a locally deeply scoured lower surface in transverse section. The bedding pattern is concentric and convex-upwards.

The primary sedimentary structures are cross-stratification of both the angular and the tangential type and ripple cross-lamination. The cross-stratification is indicative of migration directions both down and up the inclined bedding planes; reactivation surfaces are common. The inclined upper surface of the lobe is covered with small scale ripples. Convex-up ridges as described for lithofacies B5 occur in the upper part of the lobe.

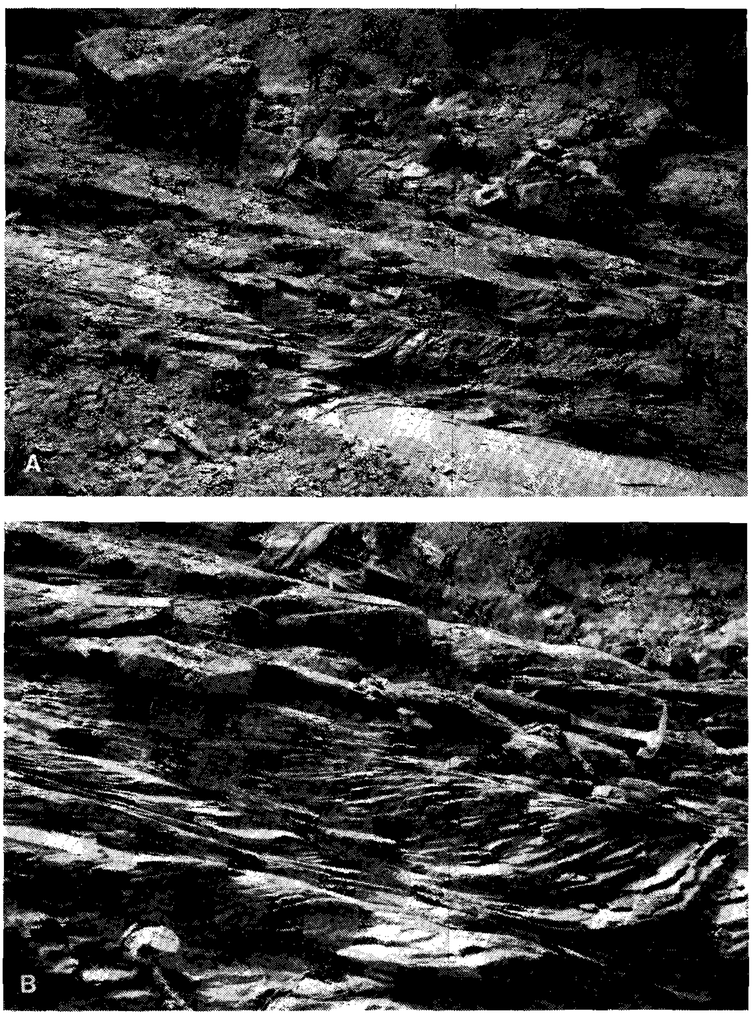


Fig. 14.- *Counter-current cross-stratification in the upper part of a lobe of lithofacies type B5. The sets of counter-current structures pinch out in the migration direction. A: the thinning of the sets is associated with a transition from tangential cross-stratification to ripple cross-lamination and poorly developed parallel lamination towards the higher part of the inclined depositional surface. B: detail of the tangential cross-stratification and the discontinuity surfaces (mud drapes) within the foreset laminae. Hammer length is 28 cm.*

Discussion and interpretation.

Sedimentary lobes with lithofacies A have been found associated with channelized bodies, either as overbank deposits of the channel (overbank splay lobes) or as channel terminal lobes. The overbank splay lobes, with a thickness of up to 1.20 m, occur at the sides of the main channel deposit. Topographically

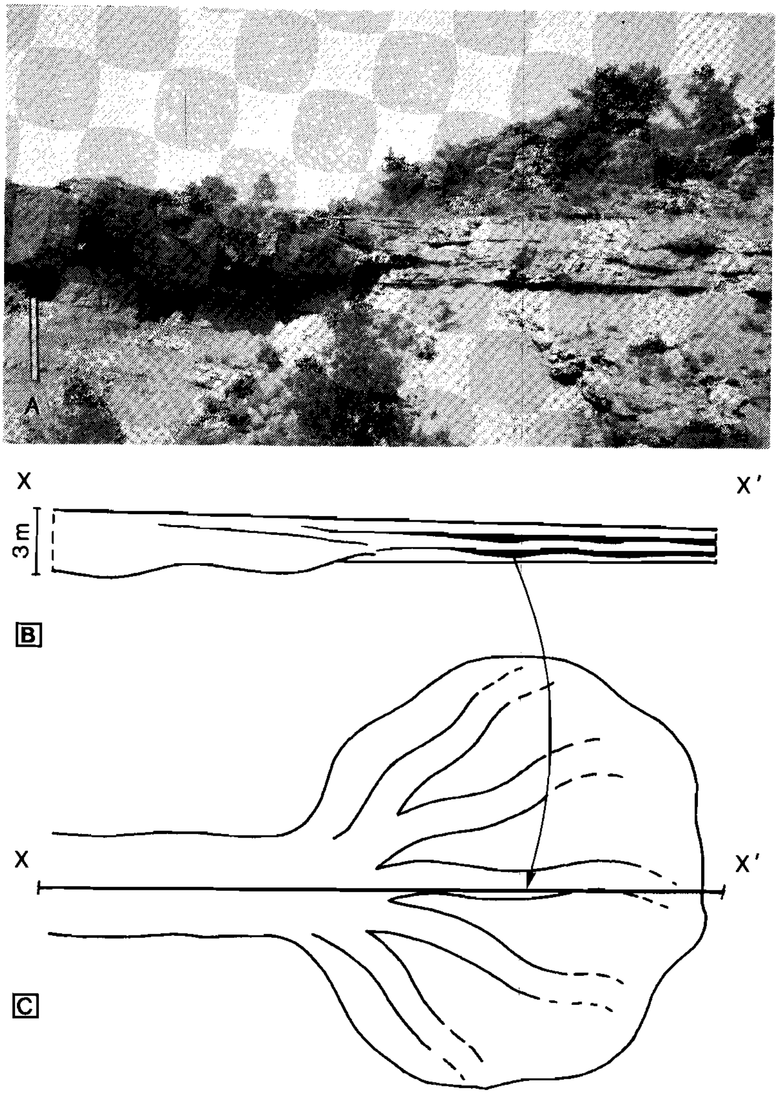


Fig. 15.- *Terminal lobe with lithofacies type A2 (right) at the mouth of a channel (left). A and B: axial section; the depth of the channel decreases towards the lobe; the erosive lower surface of the channel passes into the terminal lobe. C: Inferred plan view of the channel - terminal lobe complex.*

they are elevated several metres in relation to the channel floor. In the case of channel terminal lobes (fig. 15) the lobe occurs as the distal continuation of the channel deposit; the depth of the channel decreases towards the lobe. The lobe is approximately at the same topographic height as the channel deposit; the erosive lower surface of the channel passes into the terminal lobe. The channel terminal lobes are comparable to the terminal fans of Friend (1978) which are suggested to be the result of a decrease in transport capacity. This results in a channel pattern of distributive type, and the ultimate disappearance of all channels.

The conglomeratic lobes with lithofacies B1 are interpreted as debris-flow lobes on the basis of the matrix-supported character, the poor sorting of the conglomerate and the reverse grading. This lithofacies is very restricted within the studied deposits.

For the lobes with lithofacies B2, an origin similar to that of lithofacies A (overbank splay lobes and terminal lobes) is inferred.

The lobes with lithofacies B3, which consist of avalanche foresetting, are interpreted as small deltas deposited in a standing water body such as a lagoon. They are formed as a result of radial overbank flooding of a feeder channel, i.e. the lobe consists of both a terminal fan and lateral levees, both made up of avalanche foresetting laminae. The feeder channel is not always present over the delta lobe, probably because of channel shifting over the deltaic complex.

The more complex lobes are those with lithofacies B4, B5 and B6. They are relatively large, with thicknesses from 5 up to 10 m, and an axial extent from several tens of metres up to 400 m. No feeder channels have been observed overlying these lobes and they are therefore interpreted as *spillover lobes*. Spillover lobes are formed at the mouth of a channel; the lobe is higher than or at the same level as the base of the feeder channel. Consequently, progradation of the channel results in the destruction of the proximal part of the lobe. Spillover lobes were described first by Ball (1967) from the marine sand belts of the Bahamas; these are lobate accumulations of sediment with plano-convex geometry in transverse section. Sediment bodies comparable to the 'spillover lobes' have been observed in the Parker River estuary (Atlantic coast of the U.S.A.) by DaBoll (1969 a, fig. 8b-1) and Hayes (1969, fig. 25-1). These are intertidal, lobate, small sand bodies, with either ebb or flood orientation. The lobes are covered by sandwaves oriented in the direction of the locally dominant current. The lobes are fringed by shields; the sedimentary structures in the shields are oriented in the direction of the subordinate current. The flood tidal deltas described in the same area by Boothroyd (1969), DaBoll (1969 b), Greer (1969), Hartwell (1969) and Boothroyd & Hubbard (1975) are comparable to the spillover-lobe sand deposits described above in that both sedimentary forms consist of a sandwave flat fringed by shields, spits and channels. The feeder channel, a tidal inlet in the case of the flood tidal deltas, shallows towards the lobe flat in both cases. Shields and marginal channels of tidal deltas are dominated by the tidal counter-current (Hayes 1975, 1980). The basic difference between spillover lobes and tidal deltas is the scale; a flood tidal delta usually consists of several spillover lobes (Ball 1967, fig. 3; Hine 1975), while a spillover lobe is a single sedimentary macroform.

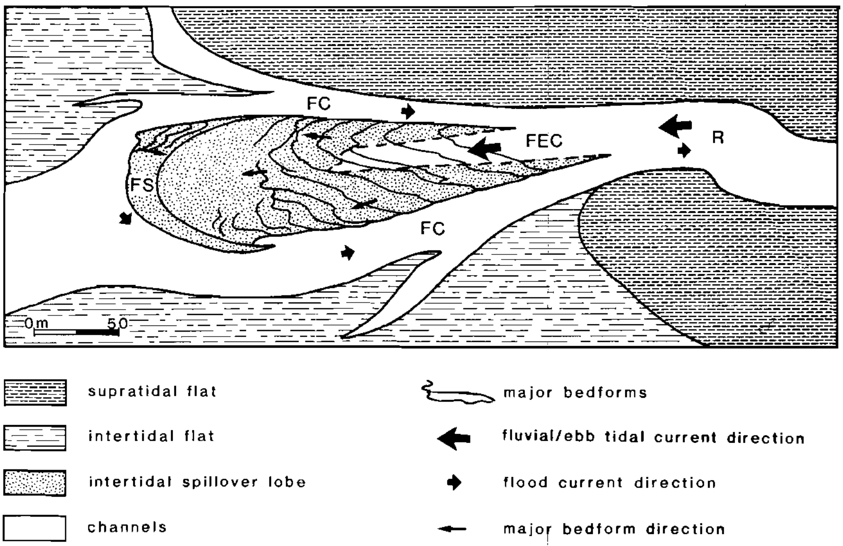


Fig. 16.- *Sedimentary model for spillover lobes in an intertidal flat. R: tidally influenced river; FEC: fluvial-ebb tidal channel; FS: flood shield; FC: marginal flood channel. The spillover lobe geometry is adapted from Hayes' (1969) photography of an intertidal sand body in the Parker River Estuary.*

A model of fluvial-ebb tidal spillover lobe is proposed for the lobes of the Capella Formation with lithofacies B4, B5 and B6 (fig. 16). They show a SW direction of progradation and are therefore ebb-oriented (the hinterland lay to the NE; Nijman & Nio 1975). The fluvial-ebb spillover lobes were formed at the mouth of a tidally influenced alluvial channel. Horizontal segregation of ebb and flood currents occurred in the intertidal area over and around the lobe. The spillover lobe bedforms are mainly ebb-oriented since the ebb current was strengthened by the river discharge. Flood-dominated structures, which are found in minor amount, probably form the lobe shields. During low river water stages, the river discharge diminishes and the supplied sediment becomes finer. At the same time, a greater symmetry of ebb and flood current intensities occurs, which allows a better development of the flood-oriented bedforms on the shields. The definitive abandonment of the lobe in consequence of the shifting of the feeder channel may lead to the reworking of the lobe by tidal currents and to the final dominance of flood bedforms on the upper surface of the lobe.

The lobes with lithofacies B4 consist of several fining upward beds. The sedimentary structures in these lobes are mainly ebb-oriented. Each fining upward bed represents a period of river discharge reactivation and subsequent waning. The scarcity of flood-oriented structures indicates a dominance of river/ebb currents, even during stages of low river discharge. This fact may be explained either by a special protected location of the lobe in the tidal flat or by weak tides during its formation.

The lobes with lithofacies B5, although ebb-dominated, show clear evidence of flood currents, which are even dominant on the upper part of the lobe. The sets of flood oriented cross-stratification and ripple cross-lamination on the upper part of the lobe represent the flood shield deposits. In the lobes with lithofacies B6, the influence of flood currents is manifested throughout the whole lobe. The upper surface of the lobe is covered by flood-oriented small scale ripples, associated with finer grained deposits and with an intense burrowing. These small scale structures represent the reworking of the lobe by flood currents after abandonment of the lobe.

The comparison of lithofacies B4, B5 and B6 suggests that deposition occurred under different tidal conditions. Lithofacies B4 appears to have been deposited in a system with, locally, a considerable asymmetry between flood and ebb, while particularly lithofacies B6 suggests a more symmetrical system of tidal currents.

Conclusions

The coarse deposits of the Capella Formation have been considered to consist mainly of fluvial channel deposits, among which point bar sequences prevail (Nijman & Nio 1975, Atkinson 1983). However, many of these 'point bar' lithosomes, when carefully examined, suggest a different interpretation. The inclined bedding pattern (epsilon stratification of Allen 1963), associated with a fining upward sequence of the deposit, is not characteristic only of point bar deposits. Sediment bodies with the same characteristics and down-dip oriented sedimentary structures can be recognized in the Capella Formation, and are interpreted as depositional lobes. In some cases the lobes show bar ridges similar to those described for the upper part of point bars (Sundborg 1956), with abundant burrowing and mottling. In the case of the lobes, the ridges are generally oriented perpendicular to the main paleocurrent direction. The lobe ridges are generated by reworking of the lobe by water currents opposite to the direction of lobe progradation.

From the plano-convex geometry of some of the sandstone bodies in terminal transverse section it can be inferred that these bodies are depositional lobes. They can be clearly distinguished from the channel features.

The grain size variations in each bed of some sandy lobes (typically a fining in the younging direction) attest to a periodicity in the water and/or sediment discharge. This periodicity may be related to climatic factors such as seasonal periods. It is inferred that rapid deposition and a high preservation potential characterize these lobes.

The sedimentary lobes of the Capella Formation are related to an alluvial system characterized by relatively fine grained deposits, mainly mudstones and sandstones. An alluvial fan environment is suggested by the presence of debris-flow lobes. Besides these, other types of lobes were deposited in the alluvial fan, i.e. terminal lobes at the distal part of streams, or small deltas when the feeder channels reached a standing water body. Where the alluvial fan was affected by tidal currents a more complex system of lobes developed: the spill-over lobes.

Acknowledgements

Dr. M. Díaz Molina and Dr. S.D. Nio are thanked for their help in the field and critical reading of the manuscript. Thanks are also due to the members of the Comparative Sedimentology Division for their suggestions and technical assistance, especially to Dr. T. de Mowbray for reviewing the English text. This study, which forms part of the author's Ph.D. research carried out at the University of Utrecht, The Netherlands, was written during the tenure of a postgraduate scholarship of the Spanish Ministry of Education which is gratefully acknowledged. The author is also indebted to the Spanish Foreign Office and the Ministry of Industry and Energy for their financial assistance at the begining of this study.

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1.4.- Annexes to the facies analysis

1.4.1.- Annex 1

Besides the channel deposits described in chapter III, part 1.2, other channelized forms are observed:

A) The channel deposits described in chapter III, part 1.2, as type 4 and type 2a are end terms of a continuous spectrum of vertical aggradation channel deposits.

In this spectrum a general evolution arises in a) the geometry, from channel deposits with high width/thickness ratio (type 4) to channel deposits with low width/thickness ratio (type 2a); b) the internal bedding, from a complex of subordinate channels and bars (type 4) to a subtle, almost parallel, internal bedding (type 2a); c) the grain size, from conglomerates and pebbly sandstones (type 4) to medium-fine sand (type 2a). This spectrum of vertical aggradation channels corresponds with a distribution of facies from high discharge to low discharge (in a general way from proximal to distal) in a distributary channel system.

In between these end terms, intermediate forms occur. The channel deposits of type 3 and 2b occupy a position slightly more proximal than that of type 2a. Another type of channel deposits (type 4a) represents a slightly more distal position than those of type 4. The channel deposits of type 4a are several meters thick and have a smaller lateral extent than those of type 4; their internal bedding indicates repeated episodes of scouring and deposition; their coarsest facies are conglomeratic to pebbly sandstones. The channel deposits of type 4a represent the transition from braided to straight channels.

B) Paleochannels of metric scale fully infilled by mudstones. They are interpreted as channels which were not connected with the source of coarse terrigenous material, for instance because their thalweg was topographically higher than the active channel of the distributary system.

C) Paleochannels infilled by sandstones in their lower parts and by mudstones in their upper parts. These forms are interpreted as channels abandoned by avulsion.

1.4.2.- Annex 2

The depositional lobes of types B3 to B5, which have incline beds and present a wedge-shaped geometry in the field, could represent the distal part of depositional lobes with sigmoidal geometry. The sigmoidal geometry could result from a slope break at the stream mouth.

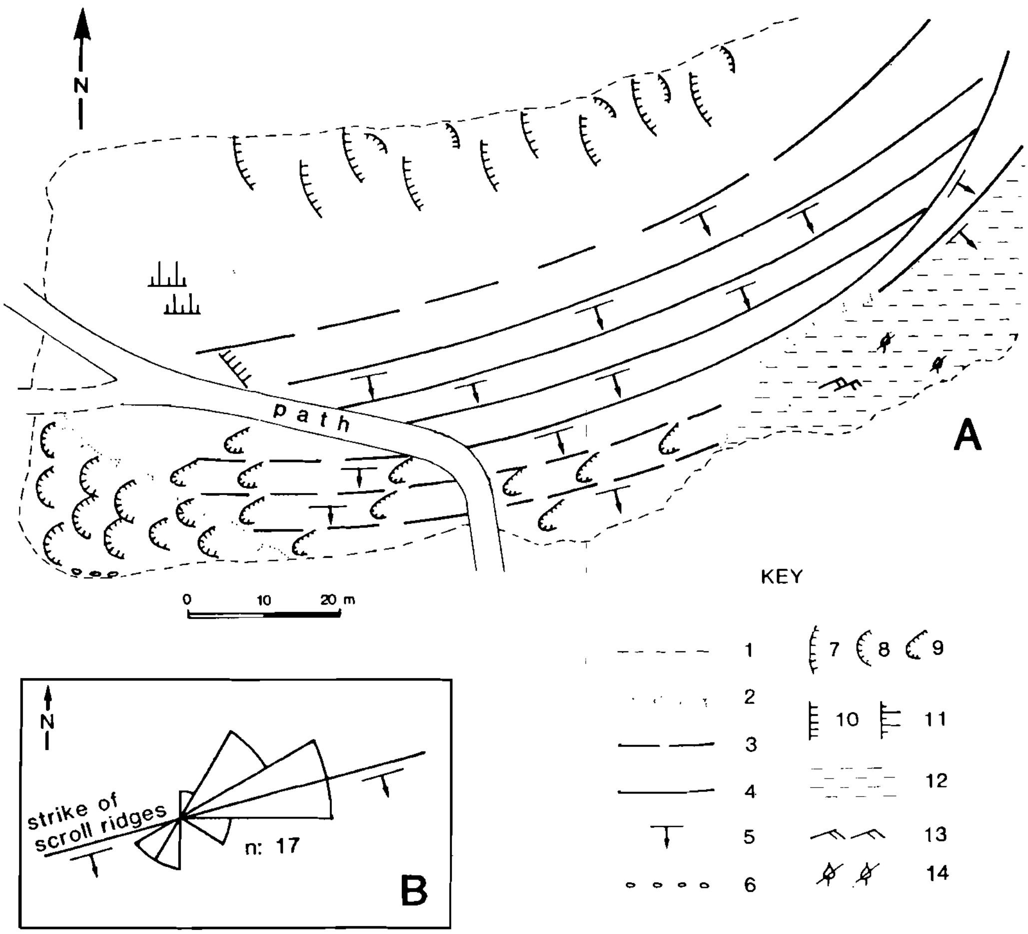


Fig. 3.1.- *Tide-influenced lateral accretion channel deposits (point bar). A: Schematic plan view of the outcrop. 1: outcrop limits; 2: stratigraphic step; 3: lateral accretion units; 4: ridges in the upper part of the lateral accretion units; 5: lateral accretion dip direction; 6: lag deposit; 7 to 9: trough cross-stratification, 7: wide, shallow trough, 8: symmetrical trough, 9: asymmetrical trough; 10 and 11: straight crested, cross-stratification, 10: with angular foresets, 11: with long bottomsets; 12: silty channel fill (abandonment facies); 13: ripple cross-lamination; 14: oxidized plant fragments. B: Distribution diagram of cross-stratification foreset dip directions; the distribution is bipolar; main current direction is parallel or slightly oblique to the scroll ridges in the upper part of the lateral accretion units (from Cuevas et al., 1985).*

1.4.3.- Annex 3

Discrimination between lateral accretion channel deposits and sandy depositional lobes with inclined-bedding and cross-stratification (types B3 to B6) is sometimes difficult, especially if a morphology of ridges and swales is present at the top of the depositional lobes and the paleocurrents were bidirectional.

Two examples of sediment bodies, a lateral accretion channel (case a) and a depositional lobe (case b), both with tidal influence, are analyzed in order to illustrate some of the differences and common points between both forms. The relation between paleocurrent directions and strike of the ridges is discussed for each model. Both examples are taken from Cuevas et al. (1985), an excursion guide to the Eocene tidal deposits of the Tremp-Graus Basin (6th IAS meeting, Lleida, Spain).

Case 1: Lateral accretion channel deposits with tidal influence (fig. 3.1).

The sediment body consists of:
a) a lower part characterized by trough cross-stratification produced by metric-wide megaripples (7 and 8 in fig. 3.1); bipolar distribution of foreset dip directions occurs in association with reactivation surfaces and mud drapes (7); a monopolar distribution of foreset dip directions is observed for the clean, coarsest sandstones (8).
b) a middle part characterized by lateral accretion units (3); the accretion units consist mainly of asymmetrical trough cross-stratification (9) parallel to the accretion units. Cross-stratification produced by straight-crested megaripples (10 and 11) occurs locally; foreset dips are oblique to the accretion units;
c) an upper part characterized by highly bioturbated lateral accretion units (4); the ridges in the upper part of the lateral accretion units exhibit a scroll-bar topography in plain view.

The distribution of cross-stratification foreset dip directions is bipolar. Main current direction is parallel or slightly oblique to the scroll ridges of the upper part of the lateral accretion units.

Case 2: Depositional lobe with inclined bedding and bidirectional cross-stratification (type B6) (fig. 3.2)

The sediment body consists of:
a) a lower part characterized by bidirectional cross-stratification of both angular and tangential types (3); reactivation surfaces and mud drapes are common;
b) an upper part consisting of:
   b1) a series of highly bioturbated sandstone ridges running parallel to each-other (2); the ridges grade downslope into
   b2) an inclined upper surface built up by small-scale ripples (4), and

b3) a frontal part characterized by a lobate topography (5) and built up by avalanche-foresetting.

The distribution of cross-stratification foreset dip directions is bipolar. The main current direction is perpendicular to the ridges in the upper part of the depositional lobe.

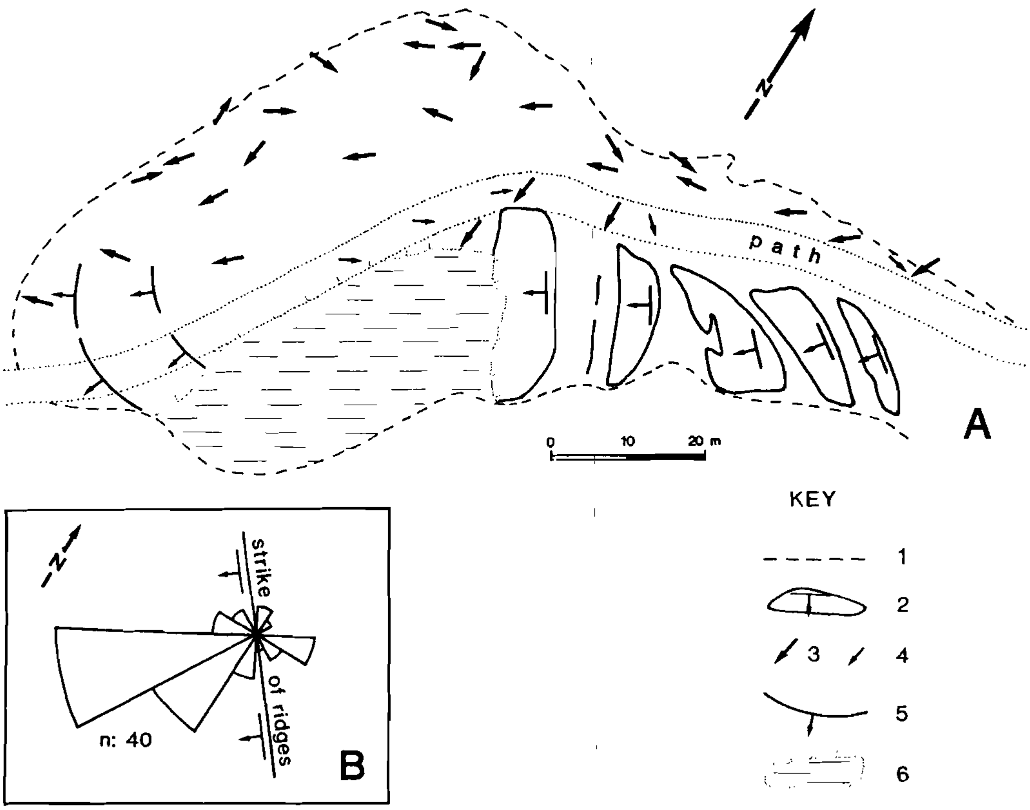


Fig. 3.2.- *Depositional lobe with bipolar distribution of cross-stratification (lithofacies B6). A: Schematic plan view of the outcrop with indication of paleo-current directions. 1: outcrop limits; 2: ridges in the upper part of the lobe, with indication of dip direction; 3 and 4: foreset dip directions of: 3: cross-stratification, 4: ripple cross-lamination; 5: strike and dip direction on the upper surface of the lobe; 6: silty abandonment facies. B: Distribution diagram of cross-stratification foreset dip directions; the distribution is bipolar; main direction is perpendicular to the ridges in the upper part of the lobe (from Cuevas et al., 1985).*

2.- SEQUENTIAL ARCHITECTURE AND DEPOSITIONAL SYSTEMS

2.1.- Introduction

Fluvial sedimentation is governed by a variety of partly independent controls. It is possible, conceptually, to isolate each one of these controls and vary its effects while keeping other controls fixed (Miall, 1985). The most important controls governing sedimentation in a coastal alluvial area are the climate, the basin tectonism and subsidence rate, and the sea-level variations. In this study the variations observed in the depositional systems have been interpreted in terms of geological variations, i.e., tectonism and relative sea-level variations; these factors controlled the relief of the source area and the basin depth. Evidence of synsedimentary tectonic instability is recorded in the deposits of the Capella Formation. The climatic influence on the sedimentation has not been considered; the collected data are not suitable for a study of the climatic variations.

Most of the channels were actively infilled (by sandstones or conglomerates) which indicates a process of backfilling of the channel system due to the relative rise of the base level (Schumm, 1977). Relative sea-level rise conditions in the sense of Vail et al. (1977) are suggested for the sedimentation of the Capella Formation. However, the unconformities bounding the tectonosedimentary units represent periods of relative sea-level fall.

In the following sections each tectonosedimentary unit and its subunits are considered separately. The different types of sediment bodies appearing within each tectonosedimentary unit are indicated and in some cases the vertical distribution of sediment bodies is used for the further subdivision of the sedimentary succession.

2.2.- Tectonosedimentary unit I

The upper part of TSU I consists mainly of sedimentary successions within the transition of the Perarrúa and Capella Formations (fig. 2.10).

The sequence within this transition is characterized by a decrease of grain size and of sediment body thicknesses. As mentioned before (chapter II, part 3.4) the deposits of TSU I occurring at different localities below the unconformities may be diachronous. However, this sequence is recognized in both the eastern and the western areas (fig. 2.10). This general trend can be related either to the retrogradation or the lateral shifting of the Montañana deltaic system and the Campanúe fan-delta system, which were responsible for the supply of clastic sediments to the Perarrúa Formation (see chapter I, part 2).

Generally a three-fold subdivision of this transitional zone can be recognized: - the Perarrúa facies association at the base, followed by the Transition facies association s.s. and finally the Capella facies association (see chapter II, part 3.4). Besides the general fining upwards, the occurrence of different types of sand bodies can be recognized in each of the three facies associations in the two main areas.

In the area of the Isábena River (section B and west of the Isábena River in fig. 2.10) the upper part of the Perarrúa facies association shows distinct thick channel sand bodies. These are lateral accretion channel deposits characterized by the occurrence of oyster debris as lag deposits. Sedimentary structures indicate sediment transport was strongly influenced by tidal currents.

The Transition facies association comprises depositional lobes of variable thickness. The depositional lobes are overlain by lateral accretion channel deposits of the Capella facies association. These channel deposits show some indication of sediment transport by tidal currents.

The lateral accretion channel deposits, both in the Perarrúa and in the Capella facies associations, are mainly oriented northwest (fig. 2.10). They form part of the Montañana axial deltaic system of Nijman and Nio (1975). A general facies shallowing is suggested by the transition from estuarine, oyster-bearing channel deposits of the Perarrúa facies association to the channel deposits of the Capella facies association. This shallowing is associated with a decrease of the channel depth (as inferred from the body thickness) and with a decrease of grain size. Thus, it is suggested that there was a lateral shifting or the total abandonment of the Montañana axial system.

In the western part of the study area (sections E to F in fig. 2.10) depositional lobe-like sand bodies can be observed in the Perarrúa facies association. In the Transition facies association depositional lobes with indications of tidal reworking can be distinguished (section E, fig. 2.10). The top of the Transition facies association consists of correlatable thin patches of a bioturbated micritic limestone, which changes westwards into a sandy limestone (eastern margin of the Esera River) (fig. 2.10). The Capella facies association is characterized by channel deposits and sheet deposits. Most of the channel deposits are of the lateral accretion type (type 1). Vertical aggradation channel deposits (type 2; section E) and vertical and lateral aggradation channel deposits (type 5; north of Capella) occur locally.

The depositional lobe-like sediment bodies within the Perrarúa and the Transition facies associations are mainly oriented southwest. They form part of the northern Campanúe fan delta as described by Nijman and Nio (1975). The small, generally meandering channels within the Capella facies at the top of TSU I, do not present an uniform orientation, although there is some evidence of an E-W trend. These channels may represent either the abandonment facies of the Montañana deltaic system or a new (tidal?) channel system which drained the abandoned coastal plain.

That a shallowing of the basin happened simultaneously with the abandonment of the depositional systems (or their shifting to a part of the basin outside the study area) is indicated by the sequential disappearance of depositional lobes of the northern system and the progradation/appearance of an axial system consisting of small meandering channels.

The shallowing cannot be explained by a progradational mechanism as suggested by Nijman and Nio (1975); rather, it probably resulted from a decrease of the rate of relative sea-level rise. The angular unconformities at the upper boundary of TSU I suggest that this decreased rate had a tectonic cause. The progressive uplifting of the basin resulted first in a facies shallowing which ended in the subaerial exposure of the deposits. Sedimentation ceased when or where the rate of relative sea-level rise reached a value of zero. With a subsequent relative sea-level fall erosion commenced and a paleovalley was incised in the eastern part of the study area.

2.3.- Tectonosedimentary unit II

No specific large-scale sequences can be observed within this tectonosedimentary unit.

Generally there is a random distribution of the different sediment bodies in the vertical succession. The conglomeratic sediment bodies, however, occur in the middle-upper part of the succession (fig. 2.9).

2.3.1.- Subunit TSU II.1

TSU II.1 is characterized by a thin vertical succession of one or two channel sand bodies (type 2 B) interbedded in mudstones. The channel sandstones contain carbonate nodules derived from erosion of paleosols. Fossil oysters occur in the abandonment facies of some of the channel deposits (sections A and B in fig. 2.9 and fig. 3.3). In the eastern part (section A in fig. 2.9) the channel sand bodies amalgamate to form multi-storey channels. This suggests a relatively low subsidence rate for the basin east of the Lascuarre fault zone.

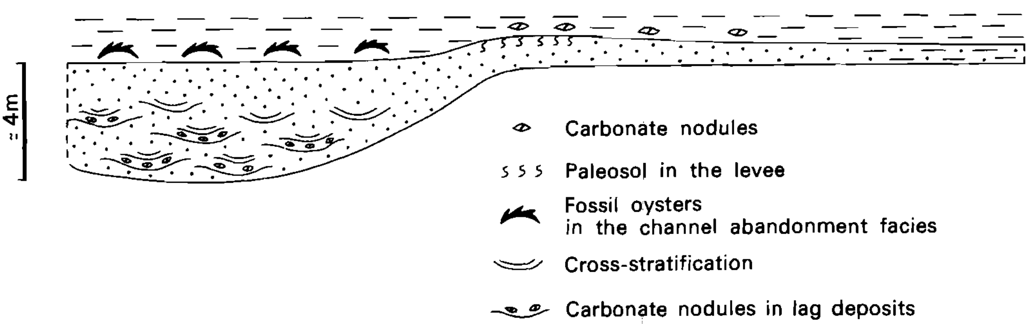


Fig. 3.3.- *Channel deposits of type 2B with oysters in the abandonment facies and carbonate nodules in the levee. Subunit TSU II.1 at Salanova.*

2.3.2.- Subunit TSU II.2

TSU II.2 is characterized by both conglomerates and coarse sandstones interbedded with mudstones which contain gypsum veins.

The conglomerates occur both as depositional lobes (type B1, see fig. 12 of Cuevas Gozalo, 1985, in chapter III, part 1.3) and as channel deposits.

Channel deposits (one to a few meters thick) showing a distinct vertical aggradation and a weak erosional base (type 4) consist of conglomerates and coarse-medium grained sandstones. Tidal influence in some of these channels is indicated by a distinct bipolar paleocurrent pattern, frequent reactivation surfaces and clay laminae in the foresets. The top of these channel deposits shows erosional reworking, probably by tides.

Sheet deposits often occur laterally away from these type 4 channel sand bodies.

Finally, small subarkosic channel sand bodies, showing vertical aggradation and a simple concave base (type 2) can be observed.

2.3.3.- Depositional system of Tectonosedimentary Unit II

Interpretation of the depositional system of TSU II is based on the following field observations:

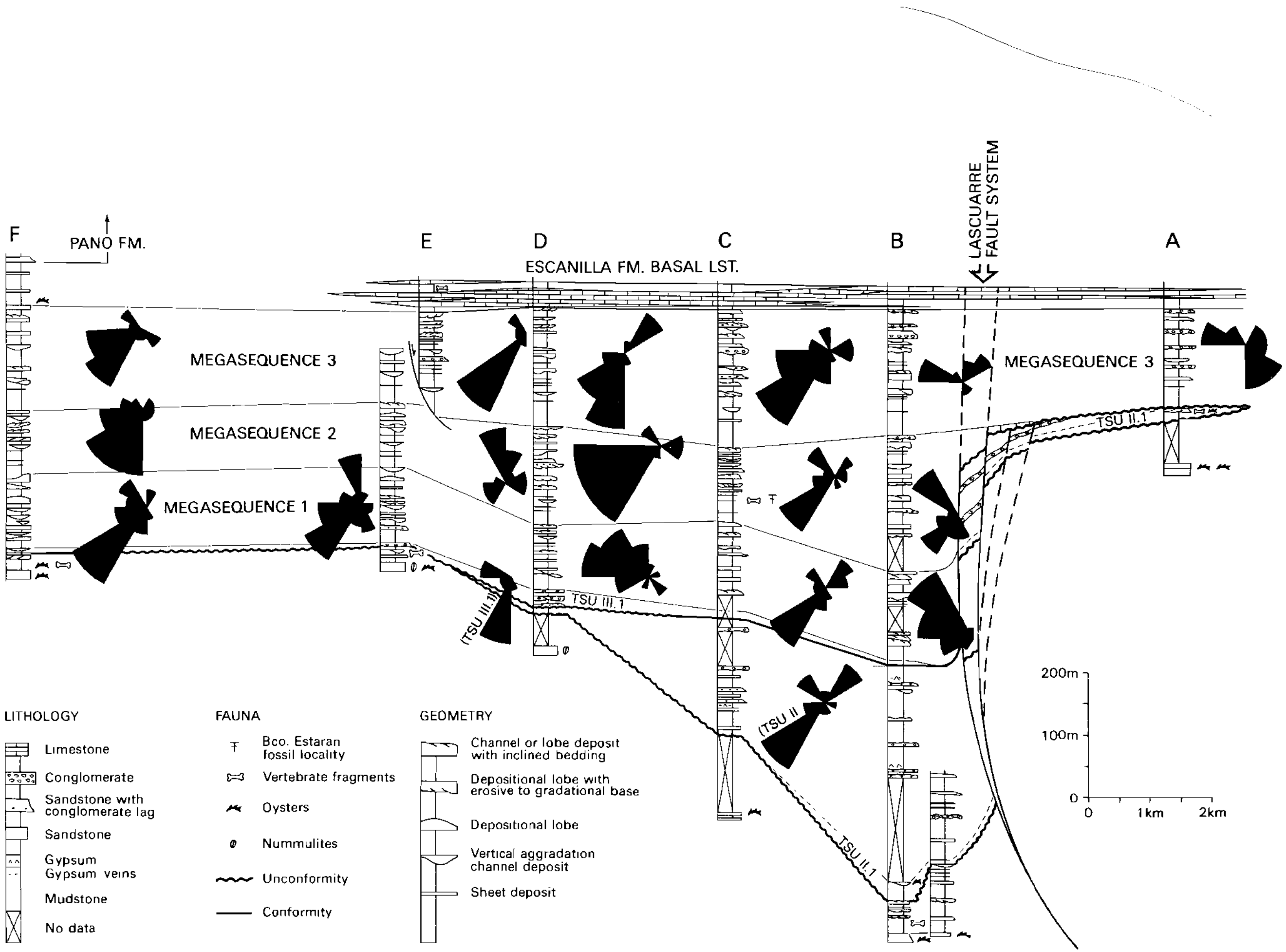
a) lateral as well as vertical grain-size variations,
b) the lateral and vertical association of sediment bodies of different dimensions and geometries,
c) paleocurrent directions.

The lateral and vertical variability in dimension, grain size and morphology of the channel deposits and their association with sheet deposits suggests the proximity of an alluvial fan (see chapter III, part 1.1). Transport of sands and conglomerates in the alluvial fan was mainly by a system of relatively shallow and wide channels, which were probably short-lived. Overflooding of the channels across the fan surface was a common process, resulting in the formation of sheet deposits laterally to the channels.

The occasional presence of debris-flow deposits (depositional lobes of type B1) suggests the relative proximity of the source area. Tidal reworking and the particular bioturbation pattern (see chapter 3, part 1.2) also suggest deposition in a low coastal plain. Tectonosedimentary Unit II can be therefore interpreted as a tide-influenced fan delta system.

Paleocurrent directions suggest a dominant depositional trend towards the SW and a subordinate (flood ?) current towards the NE (fig. 3.4). The main paleo-drainage system was towards the SW.

Fig. 3.4.- *(next page) Distribution of paleocurrent directions in the different stratigraphic units of the Capella Formation. A western-central area dominated by NE-SW current direction can be distinguished from an eastern area were SE-NW current directions are more important.*



F
PANO FM.
E
D
C
B
A
LASCUARRE FAULT SYSTEM
ESCANILLA FM. BASAL LST.
MEGASEQUENCE 3
MEGASEQUENCE 2
MEGASEQUENCE 1
MEGASEQUENCE 3
TSU II.1
TSU III.1
(TSU III.1)
(TSU II
TSU II.1
200m
100m
0
0
1km
2km
LITHOLOGY
Limestone
Conglomerate
Sandstone with conglomerate lag
Sandstone
Gypsum
Gypsum veins
Mudstone
No data
FAUNA
Bco. Estaran fossil locality
Vertebrate fragments
Oysters
Nummulites
Unconformity
Conformity
GEOMETRY
Channel or lobe deposit with inclined bedding
Depositional lobe with erosive to gradational base
Depositional lobe
Vertical aggradation channel deposit
Sheet deposit

A mega-sequential analysis of TSU II does not reveal any distinct trend. The occurrence of conglomeratic intervals in TSU II, however, indicates a rejuvenation of the relief and the formation of a new fan delta system. The occasional presence of both marine fauna and paleosols at the base of TSU II (fig. 3.3, subunit TSU II.1) indicates that TSU II commenced with a transition facies from marine to continental. Towards the top of TSU II, the occurrence of reddish mudstones, sometimes associated with gypsum, suggests a slightly regressive sequential trend. The upper part of TSU II can therefore be interpreted to have been deposited on an extensive intertidal to supratidal muddy plain, dissected by numerous, small, tidally-influenced channels.

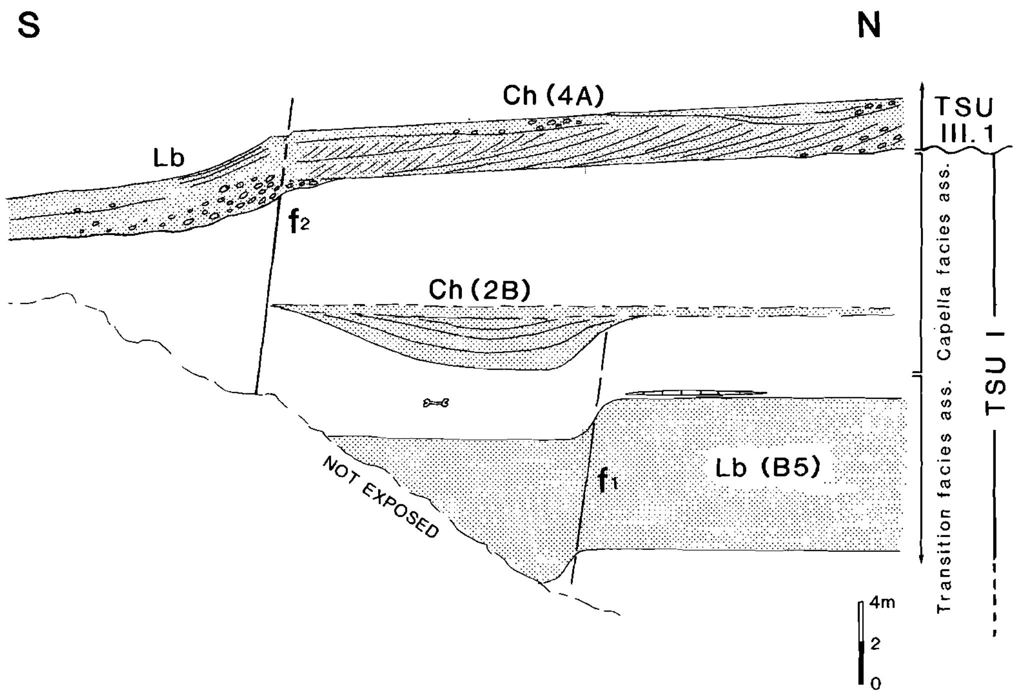


Fig. 3.5.- *Synsedimentary faults in the upper part of TSU I* (f_1) *and in the lowest part of TSU III* (f_2). *A small depositional lobe was formed on the downthrown block of fault* f_2. *Lb: depositional lobe; Ch(2A): channel deposits of type 2a; Ch(4a): channel deposits of type 4a.*

2.4.- Tectonosedimentary Unit III

Based on lithostratigraphic criteria Tectonosedimentary Unit III is divided into three subunits (for detailed information see chapter 2, part 3.6). The basal subunit TSU III.1 is a thin interval (up to 23 m) characterized by the predominance of conglomeratic and coarse subarkoses. Subunit TSU III.2 is up to 590 m thick and consists of sandstones and conglomerates interbedded with mudstones. Both subunit TSU III.1 and subunit TSU III.2 are part of the Capella Formation. Subunit TSU III.3 includes the lower part of the Escanilla Formation.

2.4.1.- Subunit TSU III.1

This relatively thin subunit is characterized by the occurrence of subarkosic conglomeratic to coarse sandy channel deposits, which form vertical aggradation bodies of type 4a (see chapter III, part 1.4.1.A). The maximum thickness of these sand bodies is around 4 m.

An interesting feature is the formation of small depositional lobes on the downthrown blocks of synsedimentary faults (fig. 3.5).

The dominant paleocurrent directions are southwards (fig. 3.4); the subordinate paleocurrent directions are oriented towards the west and were probably controlled by the E-W trending synsedimentary faults.

2.4.2.- Subunit TSU III.2

Within TSU III.2 a certain sequential organization has been recognized. The recurrence and lateral continuity of determined sequential trends is used for the definition of megasequences. Each megasequence is characterized by a lower fine-grained part, in which thin sandstone bodies are interbedded, and an upper coarse part with a prevalence of coarser, thicker and more complex sediment bodies. Here intervals with a predominance of fines alternate with sandstone- and conglomerate-dominated intervals. Generally, a gradual transition between the lower fine-grained part and the coarse-grained upper part is observed.

Three megasequences which can be correlated throughout the study area have been recognized. They each have a thickness of 100 to 250 m. East of the Lascuarre fault system, however, only one megasequence is present (section A, fig. 2.9).

The megasequences can clearly be defined in a series of diagrams, which show the vertical development of the textural composition and the thickness of the sediment bodies (fig. 3.6). More specifically, the coarse-fraction percentage, the maximum grain size and the maximum thickness of the sediment bodies within each 20 m of the sedimentary succession are plotted together.

The coarse-fraction percentage represents the occurrences of conglomerates and sandstones versus siltstones and mudstones.

Generally, a coarsening- and thickening-upward sequential trend characterizes each megasequence. This trend is very obvious for Megasequences 2 and 3. Within this general trend, smaller coarsening and fining sequences can be recognized.

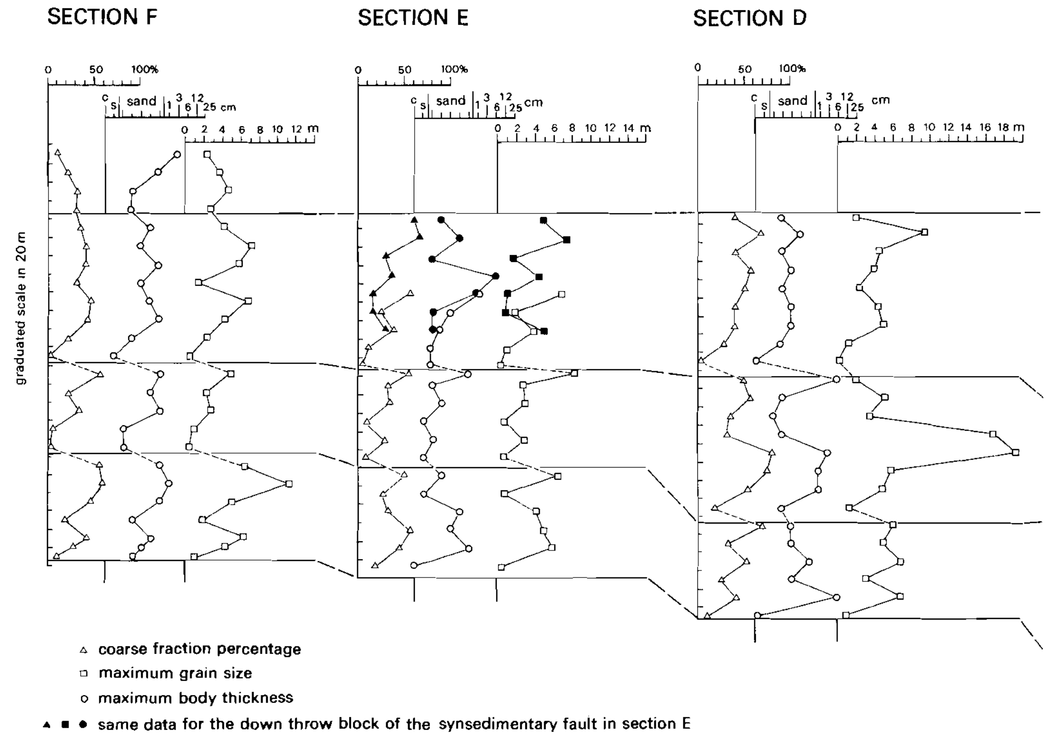
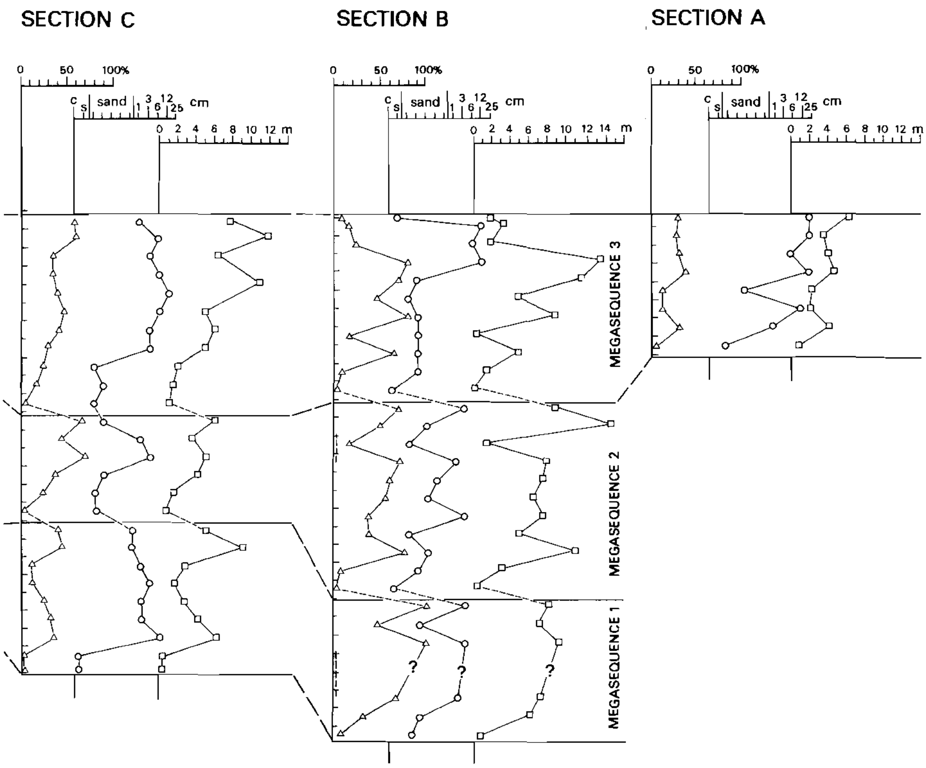


Fig. 3.6.- *Diagrams showing the vertical development of the texture and the thickness of the sediment bodies in the different sections. Each point represents the values of a 20 m thick interval. When the coarse-fraction percentages are plotted, sequential trends with a series of minimums (intervals dominated by fines) and a series of maximums (intervals dominated by coarse deposits) are recognized. Several of the fine-grained intervals are characterized by a thickness equal to or above 20 m (one exception with 15 m), a coarse-fraction percentage below 10 % (one exception with 20 %), very fine sand as maximum grain size (some exceptions with fine sand) and a maximum body thickness of 1 m. To form a megasequence a fine-grained interval with these characteristics has been grouped with the overlying coarser interval. In general, the transition between these two intervals is gradational.*


SECTION C
SECTION B
SECTION A
0 50 100%
c s sand 1 3 6 12 25 cm
0 2 4 6 8 10 12 m
0 2 4 6 8 10 12 14 m
MEGASEQUENCE 3
MEGASEQUENCE 2
MEGASEQUENCE 1

Each megasequence comprises a different stage of progradation of the depositional systems into the basin. Alternating with the progradational stages, retrogradational stages took place. These retrogradational stages are represented in the transitions between megasequences. The more or less basin-wide synchronous development of these megasequences shows that synsedimentary tectonic activities were the main conditions for these specific successions. The repeated progradational-retrogradational character of the depositional systems is related to a discontinuos sediment supply, suggesting several stages of tectonic uplifting in the sediment source area. The retrogradational stages resulted from a low rate of sediment supply during periods of low tectonic activity. The upbuilding of these megasequences also indicates a relative sea-level rise, which enabled the relatively thick successions to be accommodated.

The definition of megasequences as sequential units of large scale and lateral continuity allows the establishment of a lithostratigraphic subdivision which makes the lateral facies correlation easier (fig. 2.9). Both a detailed description and the architectural analysis of each megasequence is presented later in this chapter.

2.4.3.- Discussion and interpretation of depositional systems in TSU III.2

Based on paleocurrent directions (fig. 3.4) and sediment-body geometries two depositional systems are recognized:

a) a NE-SW distributary system, which built up a fan delta complex;
b) an axial channel system with paleocurrent directions towards WNW. This axial channel system changed in time into a fan delta system with a similar direction of progradation.

The interpretation in terms of different depositional models is based on the different associations of sediment bodies occurring within each depositional system. Differences between the depositional models are mainly due to the variability of channel morphology, which is controlled by different fluvial parameters.

Channel width (b), depth (d) and meander wave length (w) are directly related to discharge (Q), whereas slope (S) is inversely related to discharge. A generalized relation is (Schumm, 1977):

$$Q \sim b, d, w \,/\, S$$

From which it can be concluded that sinuosity (s), the inverse of meander wave length, is directly related to the channel size (b, d) and inversely related to discharge and slope:

$$s = 1/w \sim b, d \,/\, Q, S$$

According to Schumm (1977) the channel morphology of a channel system is controlled by the stream power. Stream power is directly related to channel size (hydraulic radius) and stream slope. The empirical model of Schumm and Khan (1973) related a change of channel morphology from braided to meandering to straight, to a decrease in stream slopes. The downstream evolution of channel morphologies in the northern system as well as in the axial system differ from Schumm and Khan's model:

Northern system

This NE-SW trending distributary system is characterized by low sinuosity channels (sediment bodies of types 2, 3, 4 and 5). Most of the depositional lobes (types B4-B6) are related to this system. Generally the channels do not show marine reworking but the depositional lobes, however, often show tidal reworking.

No meandering channels have been observed between the braided channels and the straight channels, as would be expected according to the fluvial model of Schumm and Khan (1973). A high frequency in the subdivision of the channels would lead to a large loss of discharge in the distributary channels, resulting in the transition from braided to straight channels, without the occurrence of meandering channels in between.

A fan delta model with a distributary channel pattern is proposed for this depositional system. Relatively low discharges in the distributary channels in the distal part of the fan resulted in straight channel morphologies. When the straight distributary channels reached sea-level, depositional lobes were formed at the channel mouth. The presence of depositional lobes at the channel mouth prevented the tidal reworking of the channel deposits (see later figs. 3.11, 3.16 and 3.19).

Two different associations of depositional lobes and channels are observed for this depositional system; they are interpreted as different morphological stages of a distributary fan delta complex.

a) Low-slope fan delta complex (fig. 3.7a):

This facies association is characterized by small straight distributary channels (types 2 and 5) and thin terminal lobes, i.e., sheet deposits, horizontally bedded lobes (types A1 and A2) and thin inclined bedded lobes (type B2), interbedded in thick fine-grained deposits. The low slope of the alluvial system will enhance a lateral dispersal of the sediment across the fan. The very shallow embayment where sedimentation took place was too restricted to allow the tidal waves to enter. However, animal bioturbation was encouraged by the fluctuating water level and the low current velocities on the tidal flat. The interaction of these mechanical and biological processes favoured the sedimentation of fine-grained deposits as well as the hydromorphic mottling and the formation of soils in topographically elevated areas (salt marsh).

It is suggested that the low-slope profile of the fan delta complex is related to periods of low tectonic activity and low basin subsidence, during which the bay was filled up and the distributary system approached its morphological equilibrium.

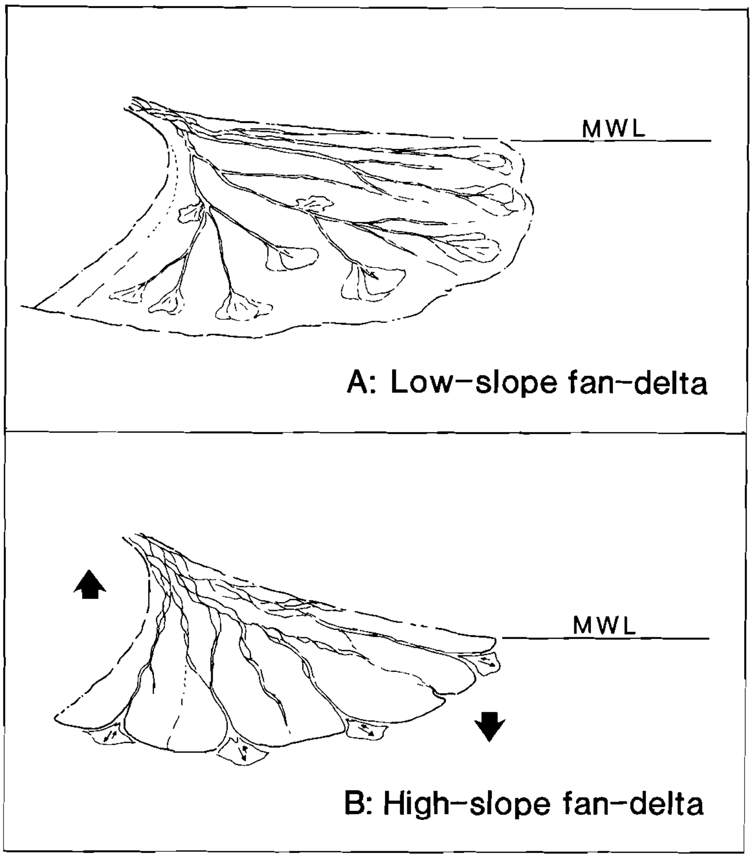


Fig. 3.7.- *Fan delta models for the northern distributary depositional system. A: Low-slope fan characterized by small channels and thin terminal lobes. B: High-slope fan characterized by larger channels and thick terminal lobes. The slope of the fan delta is controlled by tectonics. Further information in the text.*

b) High-slope fan delta complex (fig. 3.7b):

This facies association is characterized by straight and braided channels (types 3 and 4) and by relatively thick depositional lobes (inclined bedded lobes of types B3, B4, B5 and B6). Tidal reworking of the depositional lobes is common. This facies association is interpreted to have been formed in a high-slope fan delta complex, during a phase of morphological disequilibrium. A relatively deep embayment in the distal part of the fan delta system enhanced both tidal action and the deposition of thick depositional lobes at the mouths of the channels.

The high-slope profile of the fan delta complex resulted from the uplifting of the basin margins and the deepening of the basin during periods of tectonic instability.

Since the tractive force and velocity of a stream are proportional to the slope (cf. Schumm, 1977; page 129), coarser facies are expected for the high-slope fan morphology than for the low-slope one. The connection of the bay with the open sea is expected to be larger when the bay is deep and tides are more active. A better seawards dispersion of fine-grained deposits is suggested for the model of a high-slope fan delta as a result both of the higher tractive force of the alluvial channels, and the better connection of the bay with the open sea. A more condensed vertical stacking of sediment bodies is expected for the sedimentary sequence produced by such a depositional model.

Axial system

This ESE-WNW trending depositional system is dominated by both sandy and conglomeratic braided channels (sediment bodies of type 4) and meandering channels (sediment bodies of type 1). Occasionally straight channels (sand bodies of type 2 and 3) occur together with depositional lobes.

An interesting feature is that the high-sinuosity channels often show clear evidence of tidal influence.

According to the model of Schumm and Khan (1973) fluvial systems exhibit straight channel morphologies in their distal part. However, channels in the distal part of the axial system frequently present a meandering morphology and show clear indications of tidal influence (upper part of Megasequences 1 and 2, see later figs. 3.11 and 3.16). The frequent absence of straight channels in the distal part of the system could result from a lack of space in longitudinal section for developing a profile with this type of channels. The lack of space possibly resulted from a relative rise of the base-level. Alternatively the meandering morphology of the channels could also be the result of tidal action in the distal part of the fluvial system.

The axial depositional system is interpreted to form a channel belt with channel morphologies from braided to meandering at the time of deposition of Megasequences 1 and 2 (see later figs. 3.11 and 3.16). However, the lateral variations in the size and lithofacies of the channels of this system suggest that at the time of deposition of Megasequence 3 the axial system was a wet fan delta system (see later fig. 3.19). There is also a change in the petrology of the coarse components of the axial system in Megasequence 3. While in Megasequences 1 and 2 the coarse components consist mainly of grey limestone, black chert and quartzite, in Megasequence 3 new components appear, such as sandy limestone (Santonian-Maestrichtian), alveolina-limestone (Late Paleocene-early Ypresian) and white micritic limestone. This variation suggests a change in the petrology of the source area.

In general the northern system is finer than the axial system. However, a predominance of the northern marginal reliefs at the time of Megasequence 2 is manifested by the presence of pebbles and cobbles in some of the deposits of the northern distributary system, while a decrease in the stream power of the axial channel system is indicated by the scarcity of pebbles and the dominance of meandering channels (see later fig. 3.16).

From the interpretation of the lateral distribution of sediment bodies, an eastern landward area dominated by the axial channel system and a western seaward area dominated by the northern system are inferred. The interaction of the depositional systems resulted in the formation of composite forms in Megasequence 2 (section C, see later figs. 3.12 and 3.16).

2.4.4.- Architectural analysis of the Megasequences

The distribution of the different types of sediment bodies is summarized in the correlation diagrams of the megasequences (see later figs. 3.8, 3.12 and 3.17). A cyclic occurrence of depositional lobes and channel deposits is observed within the megasequences. Based on the association of sediment bodies and lithology three units are distinguished within each megasequence.

Unit 1 is characterized by the abundance of mudstones and sheet deposits and by the presence of small channel deposits of either the northern or both depositional systems. Evidence of tidal reworking either in the depositional forms or in the channel deposits is rare.

The most characteristic feature for unit 1 is a coarsening- and thickening-upward sequence from mudstones to sheet deposits to small channels. This sequence was produced by the progradation of either the northern system or both depositional systems over a low-slope coastal plain (see table 2).

The transition to *unit 2* is marked by the occurrence of thick depositional lobes of the northern system, frequently in association with coarser facies. This vertical change of facies resulted from a deepening of the basin and the associated reactivation of the northern marginal reliefs; i.e., a tectonic-induced increase of the rate of relative sea-level rise (see table 2). As a consequence of the tectonic movements the profile of the northern system changed from a low-slope model to a high-slope model (fig. 3.7).

In some cases in the eastern area, pebble-lagged channel deposits of the axial system occur laterally to the depositional lobes of the northern system, suggesting the reactivation and progradation of this system. Channel deposits of either the northern system or both depositional systems prograded over the depositional lobes, producing a regressive succession (see table 2).

Unit 3, like unit 2, is characterized by the occurrence of thick depositional lobes at the base. In two of the three megasequences relatively coarse channel deposits of the axial system occur laterally to the depositional lobes in the eastern part. The facies at the base of unit 3 also resulted from tectonic movements which reactivated the marginal reliefs and produced a deepening of the basin. In this way the high-slope profile of the northern system was maintained (see table 2).

In contrast to unit 2 the depositional lobes are not always overlain by channel deposits. Instead, the sequence may continue with progressively thinner depositional lobes. In the most frequent case, when channels of one or other system prograded over the depositional lobes, the channel deposits exhibit

Table 2.- *Interpretation of the architectural framework of Tectonosedimentary Unit III in terms of episodes of high and low tectonic activity, related episodes of progradation/retrogradation of the depositional systems, changes in the slope of the coastal plain and finally variations in the ratio between the rate of relative sea-level rise and the rate of sedimentation. For further information see the text.*

Litho-stratigraphic subdivisions		GRAIN-SIZE TREND	FACIES	Sequential units	TECTONIC EPISODES	COASTAL-PLAIN SLOPE in transversal direction (SW – NE)	PROGRADATION / RETROGRADATION EPISODES: Northern system	PROGRADATION / RETROGRADATION EPISODES: Axial system	Rate of relative sea-level rise / Rate of sedimentation (>1 · 1 · <1)
TSU III.3	lower Escanilla Fm.		paleosols fossiliferous grey mudst & limestones		low tectonic activity	filling-up	Retrogradation	abandonment	
SUBUNIT TSU III.2	Megasequence 3		fossiliferous grey mudst & depositional lobes depositional lobes on channel deposits sheet deposits	UNIT 3		H.W.L. infilling slope / high slope	Progradation	largest progradation	
						small retrogradation			
			small channel deposits conglomeratic channels depositional lobes grey mudstones	UNIT 2		H.W.L. infilling slope / high slope	Progradation	progradation insinuated	
			small channel deposits sheet deposits mud deposits, paleosols	UNIT 1	low tectonic activity	H.W.L. low slope filling-up	Progradation Retrogradation	?	
	Megasequence 2		tide-influenced channels depositional lobes	UNIT 3		H.W.L. infilling slope / high slope	Progradation	progradation	
			channel deposits depositional lobes	UNIT 2		H.W.L. infilling slope / high slope	Progradation	progradation	
			small channel deposits sheet deposits mud deposits, paleosols grey mudstones	UNIT 1	low tectonic activity	H.W.L. low slope filling-up	Progradation Retrogradation	progradation	
	Megasequence 1		tide-influenced channels & depositional lobes	UNIT 3		H.W.L. infilling slope / high slope	Progradation	large progradation	
			channel deposits depositional lobes	UNIT 2		H.W.L. infilling slope / high slope	Progradation	large progradation	
			sheet deposits mud deposits	UNIT 1	low tectonic activity	low slope filling-up	(progradation) Retrogradation	?	
TSU III.1	B. Cg.		conglomeratic-sandy channel deposits		tectonic uplifting		Progradation	?	

evidence of tidal influence. The tidal influence on the channel deposits, as well as the thinning sequence of depositional lobes resulted when the rate of relative sea-level rise exceeded the rate of sedimentation rate on a coastal plain with a slope built up by infilling (see table 2).

These features constitute the architectural framework of the megasequences. However, each megasequence shows small deviations from the general pattern or presents particular features of importance for the interpretation. In the following section, the detailed distribution of sediment bodies within each megasequence is analyzed.

Megasequence 1 (figs. 3.8 - 3.11)

The thickness of Megasequence 1 varies between 100 m (section D) and 155 m (section B); the lower fine-grained part is relatively thin (15-31 m) (fig. 3.8). The most common coarse facies are fine to medium sandstones; pebbles and cobbles occur both as lag deposits or are intercalated in the foreset cross-stratification of coarser sandstones. The megasequence consists of a thin coarsening upwards sequence at the base, followed by two fining upwards sequences in coarser facies. In a general way each of these sequences corresponds with a unit.

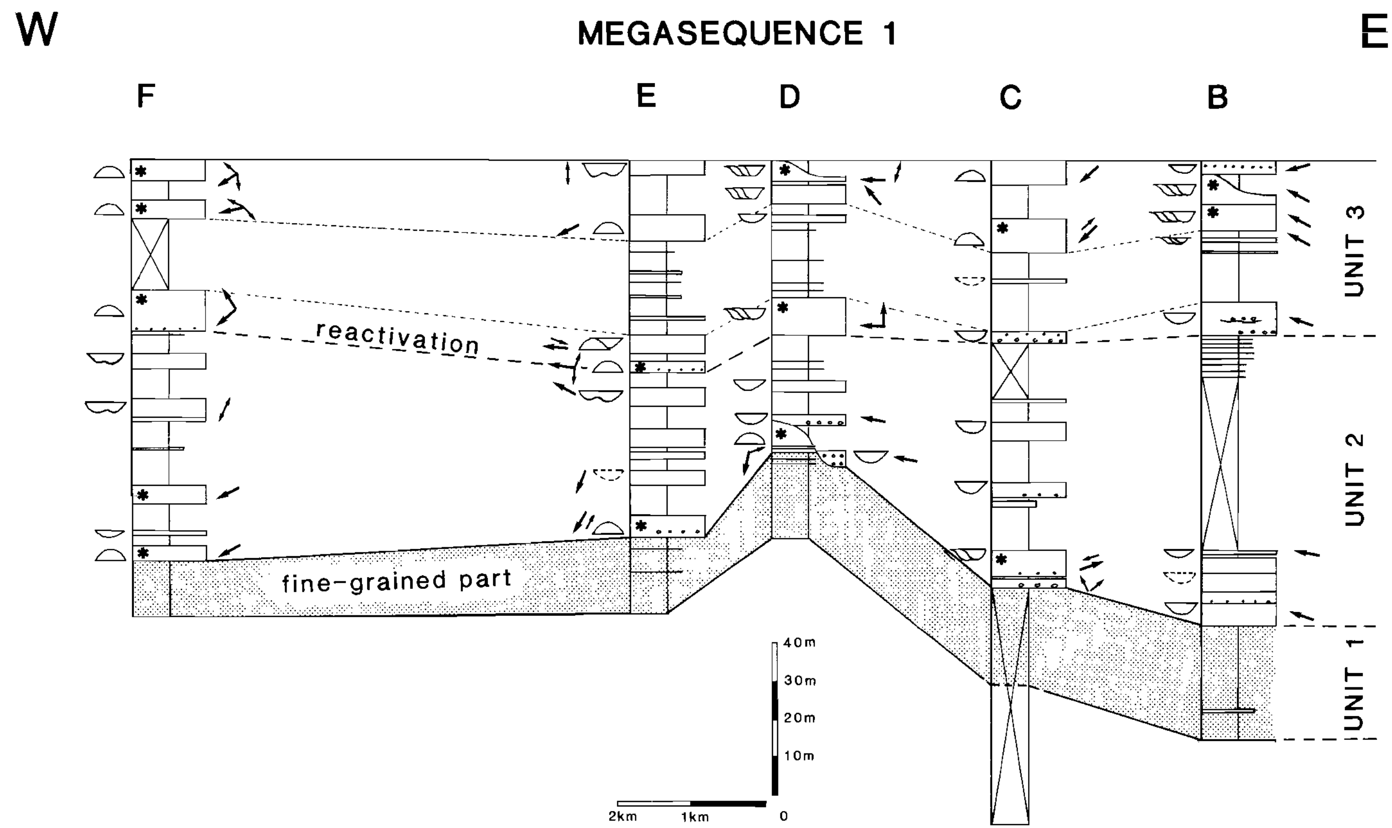
Unit 1

Unit 1 shows an incomplete sequence consisting of sheet deposits interbedded in ochre mudstones. A clear but limited coarsening- and thickening-upwards trend is recognizable at section D from sheet deposits at the top of unit 1 to a thick depositional lobe at the base of unit 2 (fig. 3.8).

Unit 2

The basal part of the unit is characterized in the eastern area (section C) by thick meandering channel deposits of the axial system. In the western area (sections D to F) thick depositional lobes of the northern system occur (types B4 and B5). Both meandering channels and depositional lobes show clear evidence of tidal influence.

Higher in the sequence low-sinuosity channel deposits of both depositional systems are the main feature (fig. 3.8). In the eastern area (sections B to D) the axial system remains dominant; sandy, pebble-lagged braided channel deposits (type 4) are common. Sandstones in this type of channel are relatively clean of fine fraction; no mud drapes or other indication of tidal influence are observed. The channel deposits of the axial system become finer in the vertical succession. Sandy channel deposits of type 3 are found at the top of the unit (section E, fig. 3.8). In the western area (sections E and F) channel deposits of the northern system prevail (channel deposits of type 3).



W
MEGASEQUENCE 1
E
F
E
D
C
B
UNIT 3
UNIT 2
UNIT 1
reactivation
fine-grained part
40m
30m
20m
10m
0
2km
1km
0

Unit 3

The beginning of the unit is marked both by the appearance of depositional lobes and of slight coarser facies (fig. 3.8).

At the base of the unit the deposits of the axial system show a westwards gradation from pebble-lagged channel deposits (type 4; sections B and C) to lateral accretion channel deposits with tidal influence (section D) and to a depositional lobe of type B6 (bidirectional cross-stratified, section E). The northern system is represented by a thick depositional lobe in section F.

Higher in the sequence an interval dominated by fines, sheet deposits and very thin channel deposits occurs (fig. 3.8).

In the upper part of the sequence lateral accretion channel deposits with a main paleocurrent direction similar to that of the axial system appear in the eastern area (sections B and D). The northern system consists mainly of thick depositional lobes of types B4, B5 and B6 (sections C, E and F). Both lateral accretion channels and depositional lobes show evidence of tidal influence.

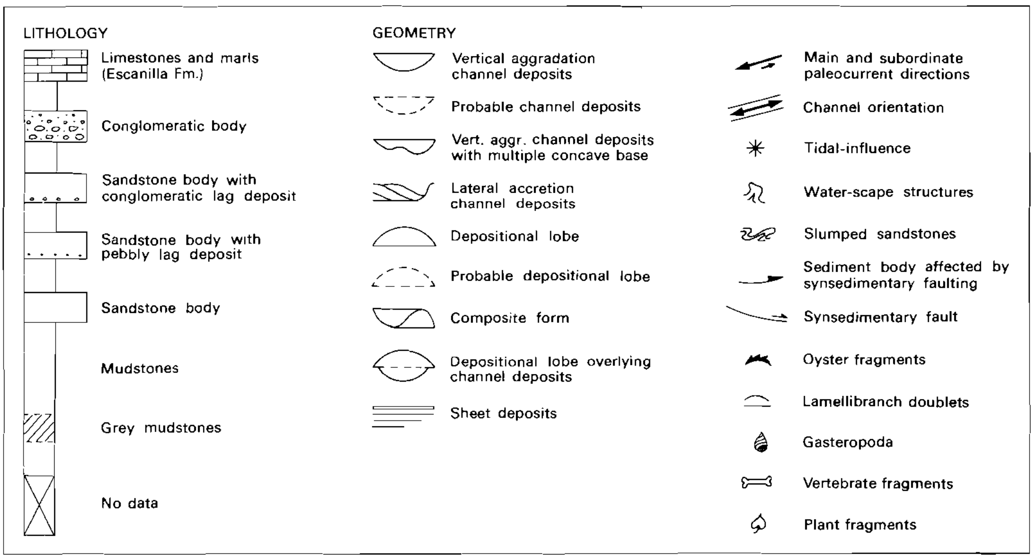


Fig. 3.9.- *Legend to figs. 3.8, 3.12 and 3.17.*

Fig. 3.8.- *(left page) Distribution of sediment bodies within Megasequence 1. See legend in fig. 3.9. Further information in the text.*

MEGASEQUENCE 1

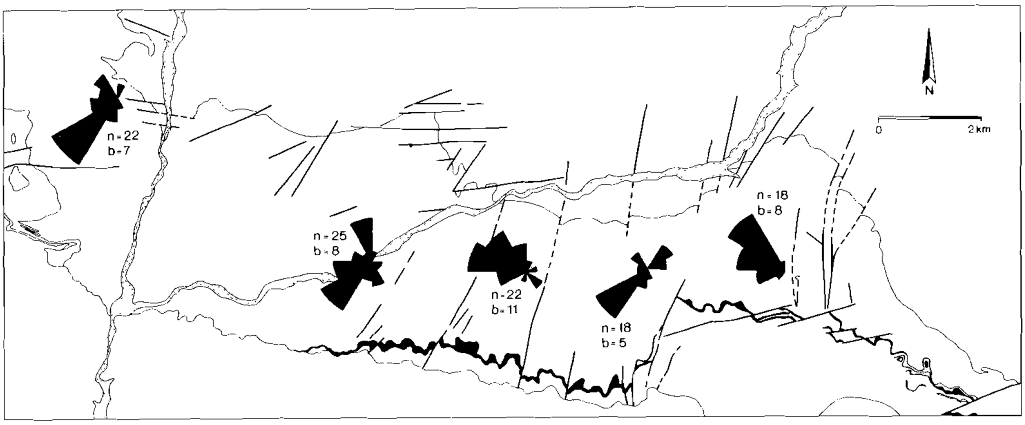


Fig. 3.10.- *Lateral distribution of paleocurrent directions within Megasequence 1. In general, an eastern area dominated by SE-NW directions can be distinguished from a western area dominated by NE-SW directions. This distribution corresponds with the dominance of the axial system in the eastern area and of the northern system in the western area. Directions towards NE probably represent tidal-flood paleocurrent directions. n: number of paleocurrent measures; b: number of sediment bodies in which paleocurrent directions were measured; in general, up to 3 or 4 measurements were considered from each sediment body.*

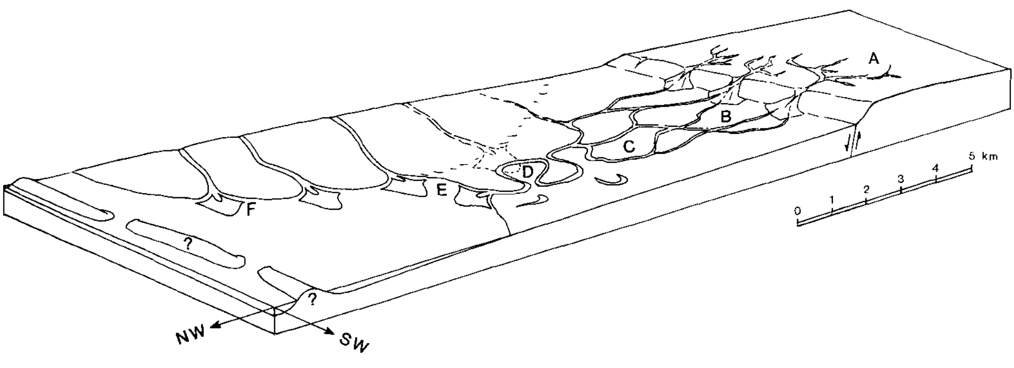


Fig. 3.11.- *Paleogeographic reconstruction during deposition of Megasequence 1. Letters indicate the positions of the different sections.*

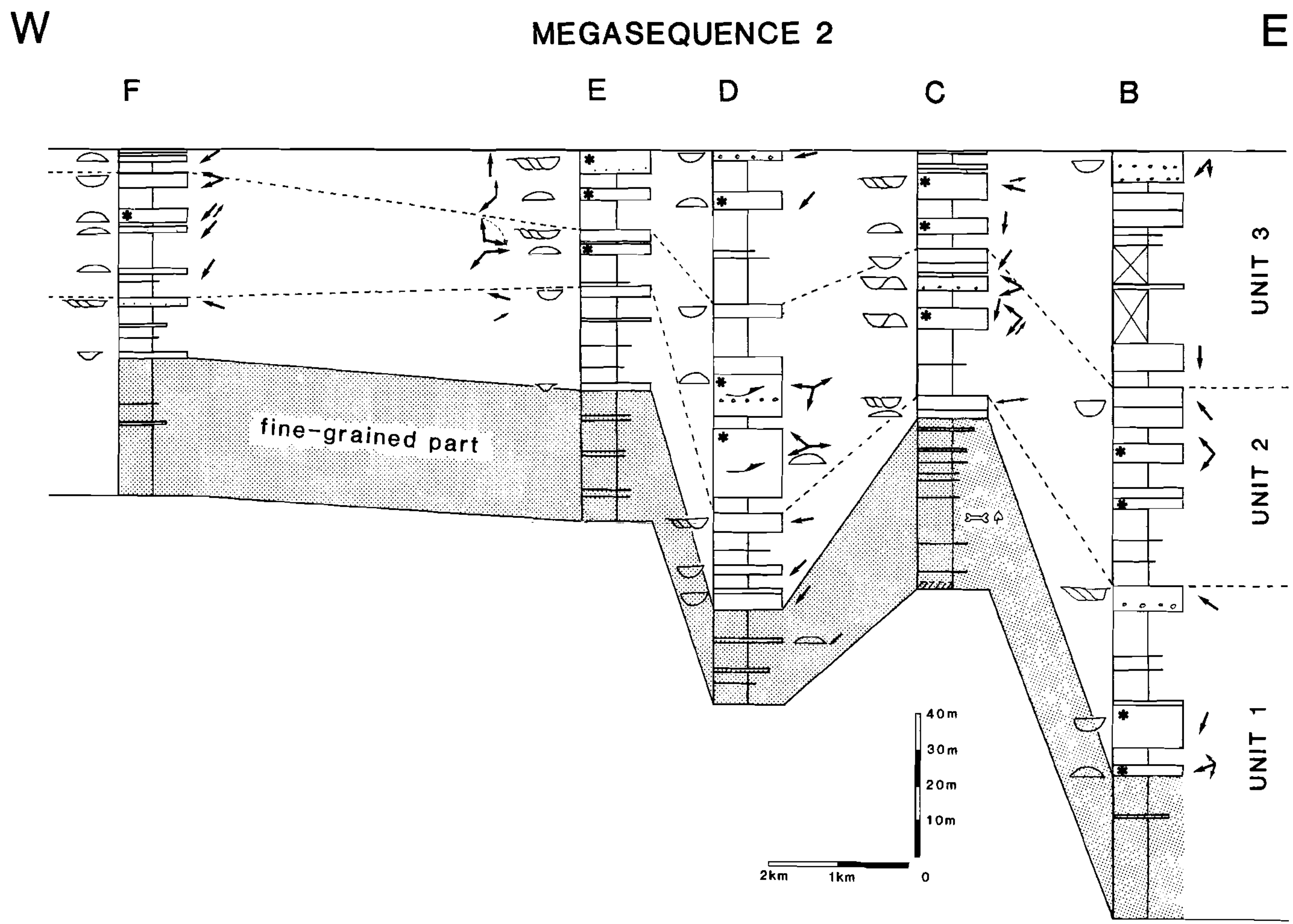
Discussion and interpretation of Megasequence 1 (see table 2)

After a period of deposition in a low-slope coastal plain (unit 1) the reactivation of the marginal reliefs and the deepening of the basin resulted in the progradation of the depositional systems. The sedimentation of pebble-lagged depositional lobes and meandering channels with tidal influence (base of unit 2) reveals that the rate of relative sea-level rise associated with the regional tectonic activity was higher than the rate of sedimentation (see table 2).

Higher in the sequence (upper part of unit 2) a minor regressive succession resulted from the progradation of channel deposits of both the northern and the axial systems over depositional lobes (fig. 3.8 and table 2). A progressive decrease of the slope of the axial system in particular, and of the coastal plain in general, is suggested by the fining upward succession of channel deposits in the upper part of unit 2. This decrease of slope probably resulted from the infilling of the basin.

A new slight deepening of the basin associated with the reactivation of the basin margins resulted in the formation of depositional lobes at the base of unit 3 with slighter coarser facies than the underlying channel deposits (sections E and F) and in the recurrence of pebble-lagged braided channels in the axial system (sections B and C, fig. 3.8). This reactivation of the axial system was ephemeral. After a short period the axial system was abandoned, leaving a coastal plain dominated by sheet deposits and fines, and dissected by some small channels (fig. 3.8). Later, larger tide-influenced meandering channels developed. Although the main WNW paleocurrent direction of these channels is consistent with that of the axial system, a purely tidal origin is not excluded for them. In the Recent tide-influenced Niger delta the distributary channels have a low sinuosity pattern, while the laterally associated tidal creeks meander widely (Allen, 1965 d).

The occurrence of tide-influenced channels and of depositional lobes at the top of unit 3 suggessts that the rate of relative sea-level rise exceeded the rate of sedimentation (see table 2).


W
MEGASEQUENCE 2
E
F
E
D
C
B
UNIT 3
UNIT 2
UNIT 1
fine-grained part
40m
30m
20m
10m
0
2km
1km

Megasequence 2 (figs. 3.12 - 3.16)

Megasequence 2 has a thickness between 100 m (section F) and 220 m (section B); the lower fine-grained part is 30-50 m thick (fig. 3.12). The transition between the fine-grained part and the coarse-grained part is very gradational. The general sequence is coarsening- and thickening-upwards; local variations occur, especially in the areas were subsidence was greater (sections B and D).

A difference with the other megasequences is that sediment bodies show a characteristic vertical distribution which is repeated in all sections (fig. 3.13). Each of the three units into which the megasequence is subdivided consists of depositional forms in its lower part and of channel deposits in its upper part (fig. 3.14).

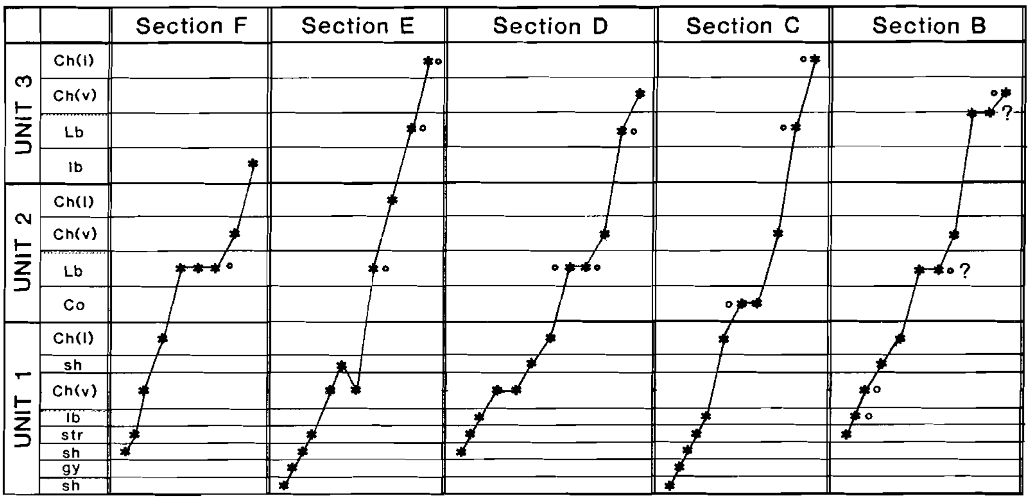


Fig. 3.13.- *Sequential distribution of sediment bodies in the different sections of Megasequence 2. Main types of sediment bodies are indicated in the second column from the left. Each asterisk indicates a sediment body (or several thin ones) of the type indicated in the second column. Lines connecting sediment bodies indicate the vertical sequence, from the base (left) to the top (right) of the megasequence. Megasequence 2 is subdivided into three units which are characterized by a particular succession of sediment bodies. sh: sheet deposits; gy: red mudstones with gypsum veins; str: stream-flood deposits; lb: thin depositional lobes; Ch(v): vertical aggradation channel deposits; Ch(l): lateral accretion channel deposits; Co: composite form; Lb: thick depositional lobe;* **o**: *indication of tidal influence in the adjacent sediment body.*

Fig. 3.12.- *(left page) Distribution of sediment bodies within Megasequence 2. See legend in fig. 3.9. Further information in the text.*

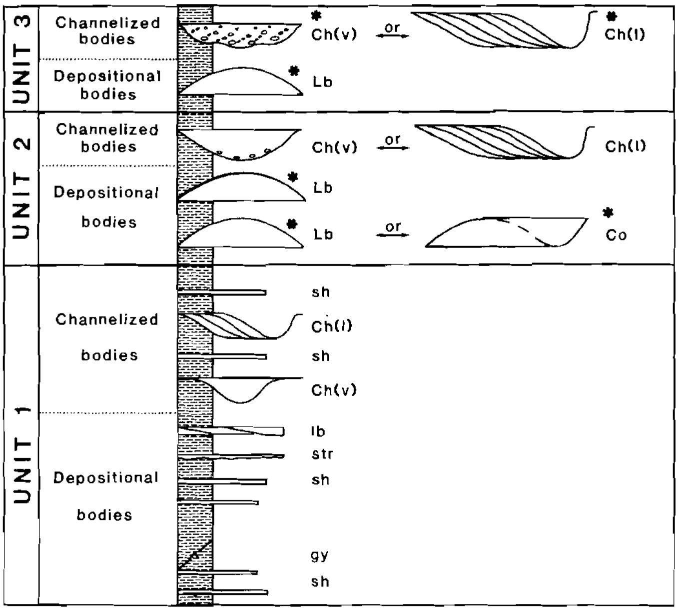


Fig. 3.14.- *Idealized sequence of Megasequence 2. Discrimination of the units is based on the occurrence of depositional bodies and channelized bodies. *: indication of tidal influence in the adjacent sediment body.*

Unit 1

Several types of thin depositional bodies are recognized in the lower part of the unit. They form a characteristic sequence (figs. 3.13 and 3.14). The lowermost part of the unit is dominated by sheet deposits. Here the sheet deposits are not laterally related to channel deposits; therefore they are interpreted as distal deposits of distributary channels (terminal fans of Friend, 1979) and not as lateral overbank deposits of channels. The most common sequence formed by the sheet deposits is coarsening- and thickening-upwards. Moreover, in some cases this typical sequence is overlain by a fining and thinning upwards succession of sheet deposits (sections C and E, figs. 3.13 and 3.14). The turning point between these two sequences is characterized by the occurrence of red or brownish mudstones with veins of secondary gypsum and paleosols.

Higher in the sequence streamflood deposits occur as relative coarse and scouring sheet deposits. They suggest a location closer to the distributary channel. Overlying this succession thin depositional lobes occur (types A2 and B2), which are interpreted as small terminal lobes.

These thin depositional forms are overlain by straight channels of the northern system (channel deposits of types 2, 3 and 5; **Ch.v** in figs. 3.13 and 3.14). In general, no indication of tidal influence is observed in them. The thickest channel deposits appear in sections B and D (fig. 3.12). Also the underlying depositional forms are thicker in these sections, suggesting that these areas were occupied by the major distributary channels of the northern system. By contrast, in section F channel deposits are thin and sheet deposits are very thin and mostly silty, suggesting lateral/distal areas in the fan delta system.

The straight channels of the northern system are in turn overlain by meandering channels of the axial system (type 1; **Ch.l** in figs. 3.13 and 3.14) or locally by small straight channels with similar paleocurrent directions (type 2, section E). No direct indication of tidal influence is recognized in any of the types.

Sheet deposits occur interbedded between the channels of the northern system and those of the axial system. It is suggested these are either overbank deposits of relative large (thick) channels (sections B and D) or terminal lobes of small (thin) channels (sections E and F) (figs. 3.12 and 3.13).

Unit 2

The lower part of the unit is characterized by relatively thick depositional lobes (types B3 to B6) and composite forms (respectively Lb and Co in figs. 3.12 - 3.14). Very thick depositional lobes are observed to be formed under the influence of synsedimentary faults (section D, fig. 3.12). Tidal influence is common, especially in the thick depositional lobes (fig. 3.12 - 3.14).

The upper part of the unit consists of channelized bodies, either of the vertical aggradation type (Ch.v) or of the lateral accretion type (Ch.l), without clear evidence of tidal-influence. Both the ocurrence of two types of channel deposits and two main paleocurrent directions suggest an interference between channels of both depositional systems.

Unit 3

The lower part of the unit consists in general of a single depositional lobe (types B4, B5 or B6). Interpretation from sedimentary structures indicates a tidal influence, with the exception of section F.

In the upper part of the unit both pebbly-sandstone channel deposits (Ch.v) of the northern system and thick lateral accretion channels deposits with pebbly lag (Ch.l) are recognized. Opposite paleocurrent directions, reactivation surfaces and very large bottomsets are found in the lateral accretion bodies as indications of tidal action. In the more conglomeratic channel deposits mud drapes occur in the foreset laminae, also suggesting a tidal influence.

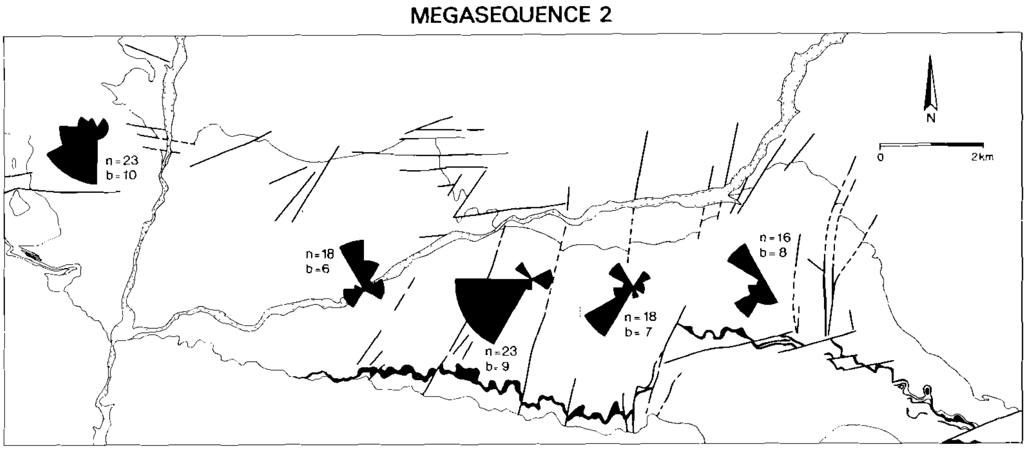


Fig. 3.15.- Lateral distribution of paleocurrent directions within Megasequence 2. NE-SW directions are dominant and associated to SE-NW directions in all sections, as a result of the dominance of the northern system and the episodic progradation of the axial depositional system. Directions towards NE and E probably represent tidal-flood paleocurrent directions. n: number of paleocurrent measures; b: number of sediment bodies in which paleocurrent directions were measured; in general, up to 3 measurements were considered from each sediment body.

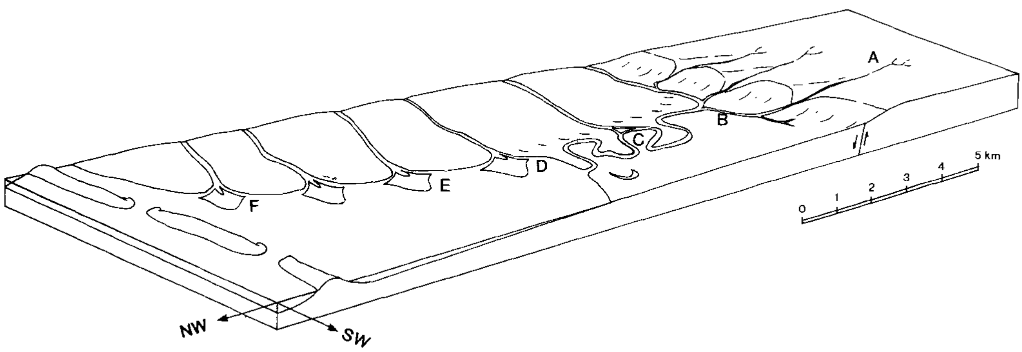


Fig. 3.16.- Paleogeographic reconstruction during deposition of Megasequence 2. Letters indicate the position of the different sections.

Discussion and interpretation of Megasequence 2 (see table 2)

In some cases a small shallowing sequence is recognized in the transition from Megasequence 1 to Megasequence 2. It is suggested that this sequence, which is very obvious in section C, resulted from the infilling of the basin. Here, the deeper facies are represented by the depositional lobes in the highest part of Megasequence 1 (fig. 3.8). These are followed by grey mudstones (lowest part of unit 1, fig. 3.12), which pass upward into a succession of sheet deposits interbedded in ochre mudstones. The end term of the sequence is marked by the occurrence of red mudstones with gypsum veins and pseudogley paleosol in red and yellow mudstones. Laterally to the paleosols, grey mudstones with plant fragments, fresh-water gasteropoda (planorbids) and rodent teeth indicate sedimentation in a small coastal pond (Bco. Estarán fossil locality, figs. 1.1, 2.7 and 3.12). This shallowing sequence resulted when the rate of relative sea-level rise was exceeded by the rate of sedimentation. In this lower fine-grained part of the megasequence the activity of the depositional systems and consequently, the rate of sedimentation were minimal, as can be inferred from the overall occurrence of fine deposits. A minor rate of relative sea-level rise must have accompanied the limited rate of sedimentation in order to produce such a shallowing sequence. It is suggested that both low rates were the result of a period of low tectonic activity, with only slight basin subsidence and little uplifting of the basin margins (see table 2).

The coarsening- and thickening-upward sequence from sheet deposits to straight channels which characterizes the lower part of unit 1, indicates the gradual progradation of the northern system within a model of a low-slope fan delta. The upper part of unit 1 represents the progradation of the axial system over a coastal plain which had been previously built up by the northern distributary system.

The transition between unit 1 and unit 2, i.e., the abrupt change from channelized forms to depositional forms, indicates a deepening of the basin. The deepening did not resulted from a shortage of sediment supply, as is evident from the general coarsening upwards sequence. Instead, it resulted from a tectonically induced deepening of the basin. The synsedimentary activity of faults during the deposition of the lower part of unit 2 (fig. 3.12, section D) supports the idea of a tectonic control on the deepening of the basin. Later, channels of both depositional systems prograded again over the depositional lobes.

Unit 3 represents a new complex episode of deepening of the basin and the subsequent progradation of channels of both depositional systems. The difference with the underlying unit 2 is that in unit 3 channel deposits of the northern system are coarser, lateral accretion channel deposits are larger (thicker) and both, especially the lateral accretion channel deposits, present evidence of tidal action. A tidal origin may be considered for the lateral accretion channel deposits, as in the underlying Megasequence 1. The coarser channel deposits indicate the further progradation of the depositional systems. However, the evidence of tidal influence in the channel deposits indicates that the rate of relative sea-level rise exceeded the rate of sedimentation.

Megasequence 3 (figs. 3.17 - 3.19)

Megasequence 3 has a variable thickness between 136 m (section E) and 223 m (section C). Ochre mudstones are the dominant lithofacies; red mudstone levels occur at the base, while grey mudstones are common in the western part. Sandstones and conglomerates are common (fig. 3.17).

The lower fine-grained part is very thick (26-60 m); the transition between the fine-grained part and the coarse-grained part is gradational (fig. 3.17). The general sequence is coarsening- and thickening-upwards; nevertheless, a fining and thinning upward trend which becomes more accentuated westwards, occurs in the upper part of the megasequence (fig. 3.17).

In Megasequence 3 the axial system presents a much coarser facies. Thick conglomeratic braided channel deposits (type 4) are common; smaller, sand-dominated channel deposits also occur. The thick conglomeratic channel deposits represent the major distributary channels of a wet fan delta system while the small sandy channels with conglomeratic lag deposits represent the minor distributaries.

Unit 1

Unit 1 is characterized at its base by red mudstones overlain by thin sheet deposits interbedded in ochre mudstones; higher in the sequence thicker sheet deposits and streamflood deposits occur (fig. 3.17). Atkinson (1987) described some pseudogley paleosols within the Capella Formation (his locality C1), which correspond with the red mudstone levels at the base of Megasequence 3.

The upper part of unit 1 consists of straight channel deposits of the northern system (type 2; fig. 3.17); channel deposits consisting of ripple-laminated fine sand are common; coarser facies with similar geometry occur locally. South and southeast oriented channels appear in the eastern sections (A and B), suggesting the radial morphology of the northern fan.

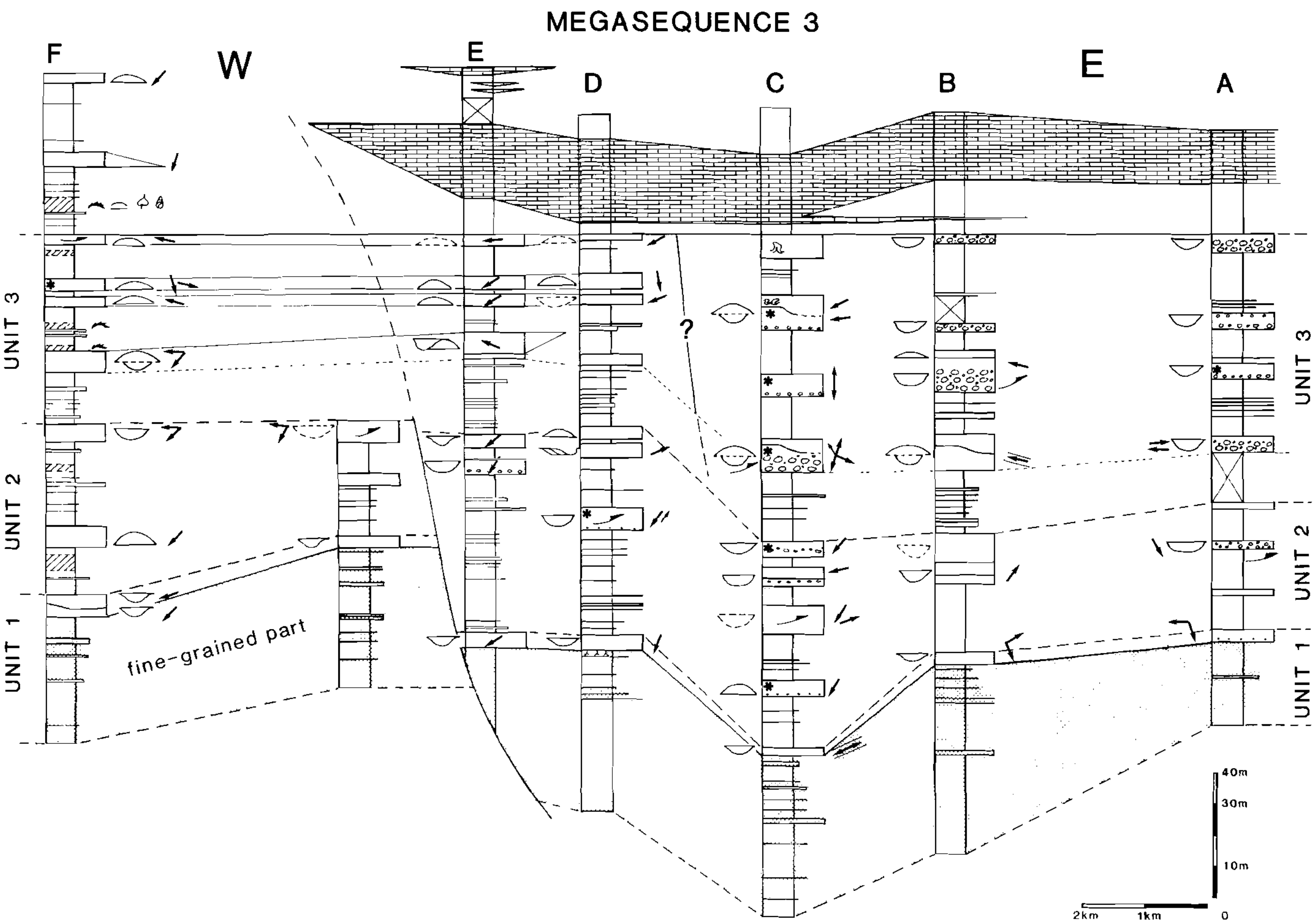
Unit 2

Unit 2 presents lateral variations in thickness and facies, which resulted from synsedimentary faulting (fig. 3.17).

The lower part of unit 2 consists of an interval of fines where some thick depositional lobes of the northern system are interbedded (types B5 and B3 respectively in sections C and F). In the western part (sections E and F) numerous thin, ripple-laminated sheet deposits are interbedded in the mudstones; in section F grey mudstones occur.

Higher in the sequence straight channels of the northern system prograded (subarkosic channel deposits, mainly of type 3); the first conglomeratic channel deposits occur at section A. Influence of tidal reworking is suggested for some of the channel deposits by opposite paleocurrent directions (section D).

Fig. 3.17.- *(next page) Distribution of sediment bodies within Megasequence 3. See legend in fig. 3.9. Further information in the text.*


MEGASEQUENCE 3
F
W
E
D
C
B
E
A
UNIT 3
UNIT 2
UNIT 1
fine-grained part
?
40m
30m
10m
2km
1km
0

An episode of tectonic instability affected the deposits; synsedimentary faults are observed in the channel deposits of sections A, C, D and E. At section E, the eastern block was downthrown along a NE-SW trending fault (fig. 3.17). As a result of the tectonic instability coarse subarkosic material was supplied to the basin; pebbly braided channels of the northern system prograded into the central area taking advantage of the tectonically subsiding block (channel deposits of type 4 in sections C and D).

Within the succession of channel deposits that infilled the more rapidly subsiding central block, a general evolution to finer deposits and smaller channels occurs (thin channel deposits of type 1 in section D and of type 2 in section E); a conglomeratic channel with indication of tidal reworking occurs only in section C.

Unit 3

The lowermost part of unit 3 consists of numerous sheet deposits interbedded in generally ochre mudstones. Some thin reddish mudstone levels occur in sections C, D and E. Sheet deposits exhibit one or several small coarsening- and thickening-upward sequences. Small depositional lobes occur occasionally above the sheet deposits (section E).

The middle-upper part of unit 3 is characterized by conglomeratic deposits in the eastern and central area (sections A, B and C), while sand deposits are the coarser sediments in the western part of the basin (sections D, E and F) (fig. 3.17). This lithofacies distribution is related to one or other depositional system prevailing. The eastern area was dominated by conglomeratic braided channel deposits of the axial system, while the western area was dominated by sandy depositional lobes of the northern system (figs. 3.18 and 3.19). An area of convergence of both depositional systems formed around section C, where the conglomerates of the axial system were deviated along an area of relatively rapid subsidence trending NE-SW (fig. 3.19).

Each section consists of four thick sediment bodies. Detailed description follows:

Section A is characterized by braided conglomeratic channels (type 4), as well as smaller sandy channels with conglomeratic lags. Opposite paleocurrent directions can be observed in some of the sandy deposits.

Section B is dominated by braided conglomeratic channels (type 4). Synsedimentary faulting of some of the sediment bodies resulted in local extreme thicknesses of channel deposits (11 m for the second sediment body). Some of the channels exhibit at the top a structureless very fine sandstone unit, up to 6 m thick. Outcrop conditions along this section do not allow an interpretation of these features; however, similar forms are better preserved and described for section C.

Section C is characterized by a fining- and thinning-upward sequence from conglomeratic deposits to sandy bodies with conglomeratic lags and to sandy deposits. Sediment bodies in section C are the thickest in the whole formation; some of these are characterized by a channelized lower member and an upper member with some features of a depositional lobe. The channelized member consists generally of cross-stratified conglomerates and sandstones or cross-stratified sandstones with a pebbly lag, in both cases with characteristics of a braided channel (channel deposits of type 4). Opposite paleocurrent directions, reactivation surfaces and mud drapes in the sandstones indicate tidal reworking. The channelized lower member is overlain and laterally replaced by a depositional lobe. The depositional lobe consists of low-angle cross-stratified to parallel stratified fine sandstone (lobe of type B4). The head of the lobes lies on a slip-face dipping about 18° southwest and a few meters high, on the underlying cross-stratified channel deposits. In one case the proximal part of the lobe is connected to hydroplastically deformed cross-stratified sandstones. Also the uppermost sand body in section C exhibits water-scape channel structures. A rapid rise of base level and a high sediment supply is inferred for these complex sediment bodies. As a result of the rise of base level, backfilling of the channel commenced. Channel avulsion was probably restricted by tectonic controls and the channel remained in a more or less fixed position. Once the channel was infilled by its own deposits, and as a consequence of the continuous rise of base level, sediments were deposited in the form of a lobe which directly overlaid the channel deposits; i.e., a transgression of distal facies over more proximal facies occurred.

The rapid rise of base level (or relative sea-level rise) in this area could have resulted either from a local increase of tectonic subsidence or from more regional controls. The tectonic instability of this area is evidenced by the presence of synsedimentary faults (sections B and C, fig. 3.17) and when compared with the western area the larger sediment thicknes indicates a higher subsidence.

Both the rapid deposition and the tectonic instability in this area could be responsible for the hydroplastic deformation of the deposits and the slip-faces in the channel deposits.

In the western area (sections D, E and F) an upward trend to finer facies and thinner bodies is observed for the middle-upper part of unit 3; coarse facies are quite similar in these sections. Thick sediment bodies begin in section E with a composite form (type b) and in section F with a complex body consisting of a lower channel-member and an upper lobe-member. Higher in the sequence three characteristic grey sand bodies of large lateral extent are recognized both in sections D and E. The bodies are interpreted as depositional lobes of type B4; sometimes a scouring base suggests a more channelized flow. In section F three sandy cross-stratified depositional lobes interbed with thin sand layers and grey mudstones; oysters are occasionally found in the mudstones. The three lobes at the top of sections D and E can be directly correlated with the three upper lobes of section F on the basis of lithofacies, thickness and sequential position. Evidence of opposite paleocurrent directions in the depositional lobes is rare. Nevertheless, from the presence of lagoon/bay facies in section F it can be inferred that the sedimentation rate was exceeded by the rate of relative sea-level rise when compared to the underlying deposits.

An interval of grey mudstones overlies unit 3, below the first limestone level of the Escanilla Formation.

MEGASEQUENCE 3

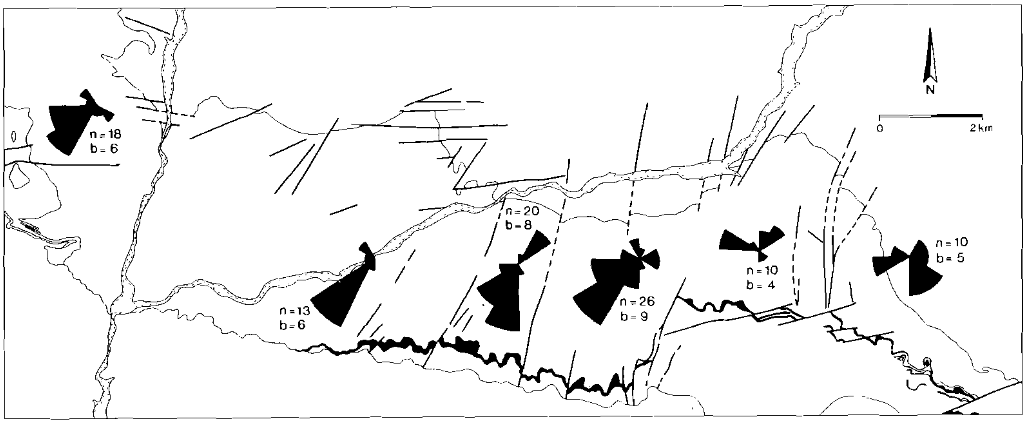


Fig. 3.18.- Lateral distribution of paleocurrent directions within Megasequence 3. In general, a small eastern area dominated by E-W directions can be distinguished from a western area dominated by NE-SW directions. This distribution corresponds with the dominance of the axial system in the more eastern area and of the northern system in the western area. Directions towards NE and E probably represent tidal-flood paleocurrent directions. n: number of paleocurrent measurements; b: number of sediment bodies in which paleocurrent directions were measured; in general, up to 3 measurements were considered from each sediment body.

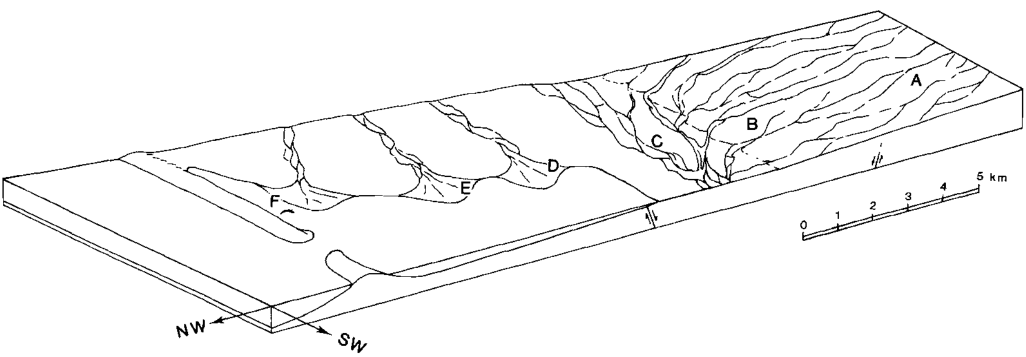


Fig. 3.19.- Paleogeographic reconstruction during deposition of Megasequence 3. Letters indicate the positions of the different sections.

Discussion and interpretation of Megasequence 3 (see table 2)

Small distributary channels of the northern system prograded over a low-slope coastal plain in which hydromorphic soil processes were common (unit 1).

A deepening of the basin resulted in the sedimentation of depositional lobes in the more areas of more subsidence (base of unit 2). The high frequency of weakly-bioturbated sheet deposits and the occurrence of grey mudstones and a depositional lobe of the avalanching foresetting type (type B3) at the base of unit 2 in section F suggest that the western area was less frequently submitted to subaerial exposure and that subtidal conditions were common. The beginning of a general transgressive sequence is evidenced here by the occurrence of subtidal facies.

After this transgressive episode, the progradation of larger channels of the northern system marked the beginning of a regressive episode (upper part of unit 2). Tectonic movements in the basin and in the basin marginal reliefs resulted both in the formation of an are of relatively more subsidence in the central part of the basin and in the progradation of the northern fan system into this area. The point of maximum progradation is indicated by the occurrence of pebbly braided channel deposits of the northern system (upper part of unit 2, fig. 3.17).

The local abandonment of the northern distributary system is suggested at the top of unit 2. Transport and sedimentation of conglomeratic deposits were concentrated into a major distributary channel at section C; tidal reworking of the channel deposits took place. The rest of the fan surface was drained by small channels with finer facies, some of which could have had a tidal origin (lateral accretion channel of section D).

A general small retrogradation of the northern depositional system is suggested by the occurrence of nearly ubiquitous sheet deposits over all the studied area at the base of unit 3 (fig. 3.17). That the retrogradation of the depositional system did not result in a transgression indicates that the rate of relative sea-level rise decreased parallel to the rate of sedimentation; i.e., the retrogradation was related to a phase of low tectonic activity (see table 2). The small thickening-upwards sequences observed for the sheet deposits indicate the further progradation of the depositional systems.

The occurrence of depositional lobes and conglomeratic channels in the middle part of unit 3 indicates, when compared with the lowermost part of unit 3, both the reactivation of the depositional systems and a transgression. As for the underlying megasequences, it is considered that this combination of depositional forms and coarser facies has a tectonic origin.

Transgressive conditions prevailed within the middle-upper part of unit 3 as manifested by:

a) the occurrence of lagoon/bay facies (section F),
b) the formation of depositional lobes directly overlying channel deposits (sections B, C and F), and
c) the sedimentation continuing with a sequence of depositional lobes (sections D, E and F).

The decrease of activity of the northern system in the western area (sections D, E and F) promoted the transgression in this area. This decrease of activity (or local abandonment) was probably associated to a relatively low subsidence rate, as can be inferred from the smaller thickness of unit 3 in the western area. Despite the relatively low subsidence rate, the rate of relative sea-level rise exceeded the sedimentation rate.

In the eastern area (sections B and D) the model is more complicated. The subsidence rate was relatively high, possibly in relation to local tectonic structures (fig. 3.17), and it was accompanied by an important supply of sediments from the axial alluvial fan system. As a result of the weak balance between tectonics, sediment supply and relative sea-level rise, several transgressive sequences were repeated on a small scale; these are indicated by the repeated occurrence of depositional lobes over channel deposits. The total rate of relative sea-level rise exceeded the total rate of sedimentation, producing a transgression. The transgression is manifested by the appearance of grey mudstones and brackish-lacustrine limestones above unit 3.

Summary of the architectural analysis of the megasequences

Three sequential units are recognized within each megasequence. Each unit comprises an episode of progradation of the depositional systems; these were separated by incidental deepenings of the basin. The most complete succession of events within a megasequence is:

1.a) Sedimentation of mainly fine-grained deposits on a coastal plain of low slope as a result of the retrogradation of the depositional systems in relation to a phase of low tectonic activity.
   b) Sedimentation at the distal part of a low-slope fan delta. Thin depositional lobes formed in front of the major distributary channels of the northern system.
   c) Progradation of the channels of the northern system over its depositional forms.
   d) Progradation of the channels of the axial system.

2.a) Tectonically controlled deepening of the basin and reactivation of the northern basin margins. The northern system changed from a low-slope fan delta to a high-slope fan delta model.
   b) Formation of thick depositional lobes.
   c) Progradation of channels of both the northern and the axial system.

3a) Tectonically controlled deepening of the basin and reactivation of the northern basin margins. The northern system maintained a high-slope fan delta model.
   b) Formation of thick depositional lobes.
   c) Restricted progradation of channels of both the northern and the axial system. Increase of the rate of relative sea-level rise over a coastal plain with a slope built up by infilling.

In the two lower units the rate of sedimentation either compensated or exceeded the rate of relative sea-level rise; in the upper unit the rate of sedimentation was exceeded by the rate of relative sea-level rise (see table 2). The final result was a slight transgression at the top of each megasequence; an accumulative transgression is evidenced in the upper part of the Capella Formation by the appearance of bay/lagoon facies.

2.4.5.- Subunit III.3

In this section the basal limestones of the Escanilla formation are discused in terms of their significance in the basin evolution. From this point of view, the limestones are laterally related to the uppermost deposits of the Capella Formation in the Esera valley.

The fining- and thinning-upward trend of the upper part of Megasequence 3 in most of the sections suggests that the lower limestones of the Escanilla Formation were deposited when transport energy in the basin was minimal and consequently deposition of terrigenous material was restricted (fig. 3.17). Similar low-energy conditions also prevailed during the deposition of a thick sediment interval of section F, although no lacustrine limestones were deposited there. Instead a thick interval of grey and ochre mudstones with interbedded thin layers of very fine sandstone were deposited above the upper depositional lobes of Megasequence 3. A layer of carbonaceous shales, rich in oysters, gasteropoda and bivalves is found within this interval (fig. 3.17). This thick fine-grained interval is inferred to be the western lateral equivalent of the mudstones and lower limestones of the Escanilla Formation in the central and eastern part of the studied area.

The gradational abandonment of the northern distributary system together with the relative sea-level rise resulted in the transgression of lagoonal facies over intertidal facies in the western area, and in the deposition of lacustrine limestones in the central and eastern areas. Fine terrigenous deposits and lignites interbedded with the limestones in section E, suggest this locality was close to the lake edge. The lateral distribution of facies from lagoonal to lacustrine may have been influenced by the presence of a threshold separating both subenvironments. The threshold, a low-subsiding area west of section E (between Capella and the Esera River) could correspond with the upthrown block of a reactivated fault like those which are known to have affected Megasequence 3 (unit 2, section E, fig. 3.17).

In the eastern area (section A) the vertical transition from fan delta deposits to lacustrine facies is not as gradational as in the central area; no fining- and thinning-upwards sequence is observed on the scale of the axial depositional system (fig. 3.17). However, a fining-upwards sequence is recognized for the uppermost channelized body of the axial system, indicating its progressive abandonment and upfilling. In this area the lacustrine facies consists of several thin white limestone beds interbedded with grey mudstones; palustrine facies (in situ breccias and other pedogenetic-associated limestones) are observed, indicating the edge of the lake. The uppermost channel deposits of the axial system are onlapped eastwards by the thin limestone beds. The onlapping is not interpreted as an unconformity but as the contact between two different depositional systems. As the rate of relative sea-level rise exceeded the rate of sedimentation, a rise of the phreatic level took place and a lake was formed. The area previously occupied by the axial system was drowned; the onlapping of marginal lake deposits over older fan delta deposits, which had a higher depositional slope, resulted from the continued relative rise of sea-level and consequent expansion of the lake.

The lower limestones of the Escanilla Formation are comparable to the lower fine-grained parts of the Megasequences in their sequential position in the architectural framework and in their significance in the basin evolution (see table 2). Both lower fine-grained parts and limestones formed during episodes of low sediment influx, related to the maximum retrogradation of the northern depositional system during a phase of low tectonic activity. In the case of the lower fine-grained parts of the megasequences the basin was filled-up by fine terrigenous deposits until subaerial conditions prevailed and hydromorphic paleosols were developed. The formation of paleosols indicates minimum rates of both relative sea-level rise and sedimentation. For the Escanilla limestones the infilling of the basin was dominantly by lacustrine deposits after the retrogradation of the northern system and the abandonment of the axial system. A period of minimum rates of both relative sea-level rise and sediment supply is indicated by the occurrence of thick paleosol levels overlying the lacustrine limestones.

Both the lower fine-grained parts of the Megasequences and the lower limestones of the Escanilla Formation represent cyclic stages of relatively low sediment supply, low tectonic uplifting and low subsidence. These stages followed periods of tectonic instability, high sediment supply and progradation of the terrigenous depositional systems. From this point of view the lower limestones of the Escanilla Formation, together with the underlying and the laterally equivalent grey mudstones, must be considered as the lower part of a new megasequence (as used in this work) in the sequential development of the basin.

CHAPTER IV: BASIN ANALYSIS

1.- INTRODUCTION

The sediment filling of the Tremp-Graus Basin during the Eocene was strongly influenced by tectonic movements. During the deposition of the Capella Formation, the basin was subdivided into different compartments separated by N-S trending faults. Synsedimentary tectonic movements along these faults resulted in a rapid lateral change of sediment facies and thickness. In an E-W profile two fault-controlled depressions are recognized: the Isábena Depression and the Viacamp Depression (fig. 4.1). A detailed description of the Capella Formation in the Isábena Depression has been given in previous chapters. Some aspects of the sediment fill of the Viacamp Depression, in relation to the Capella Formation, are considered next.

2.- CORRELATION OF TECTONOSEDIMENTARY UNITS WITH OTHER AREAS IN THE TREMP-GRAUS BASIN.

2.1.- Tectonosedimentary units in the Viacamp Depression

The Viacamp Depression is located several kilometers east of the Isábena Depression. As in the Isábena Depression, the relatively thick succession was due to the synsedimentary tectonic activity of the Lascuarre and the Chiro fault systems, respectively (fig. 4.1).

As mentioned in chapter II, in the Isábena Depression, two facies-related lithostratigraphic units were recognized betweeen the Castisent-La Roca correlation level and the Escanilla lacustrine limestone: the shallow-marine Perarrúa Formation and the more continental Capella Formation (Nijman and Nio, 1975, fig. 2.3). However, the Perarrúa Formation grades eastwards into a more continental facies. Therefore, in the Viacamp Depression only a lithostratigraphic formation, the Capella Formation, was distinguished (Nijman and Nio, 1975).

The limestones and overlying detritic facies which appear in the Viacamp Depression above the Capella Formation were considered as Escanilla Formation (Garrido Megías, 1968); in the present study the name of Viacamp Formation assigned to these facies by Donselaar and Geel (1985) is used.

Further observations in the Viacamp Depression, between the Chiro fault system and the Ribagorzana River, indicate that the sedimentary succesion formed by the Capella Formation plus the Viacamp Formation can be subdivided into two tectonosedimentary units separated by an unconformity (figs. 4.1 and 4.2):

1) A lower tectonosedimentary unit (L.V.TSU), consisting of mudstone, sandstones and conglomerates, and characterized by:

a) occasional presence of grey mudstones with oyster shells (Werver, open file report, 1987) (fig. 4.1),
b) tidal influence in the cross-bedding of the sandstone bodies,
c) little variation in the composition of the coarse detritic components (lower Capella monomictic conglomerates of Werver, 1987, open file report). No granitic, nor doleritic pebbles are found (Kunst in Mijnssen, 1986; Werver, 1987; open file reports).
d) concentration of ochre-coloured fines and ruiniform calcitic soil horizons (cp. Freytet, 1971) at the top of the unit,
e) frequent small scale faulting (E-W and NE-SW trending faults are common) results in chaotic outcrop conditions and difficult internal correlations.

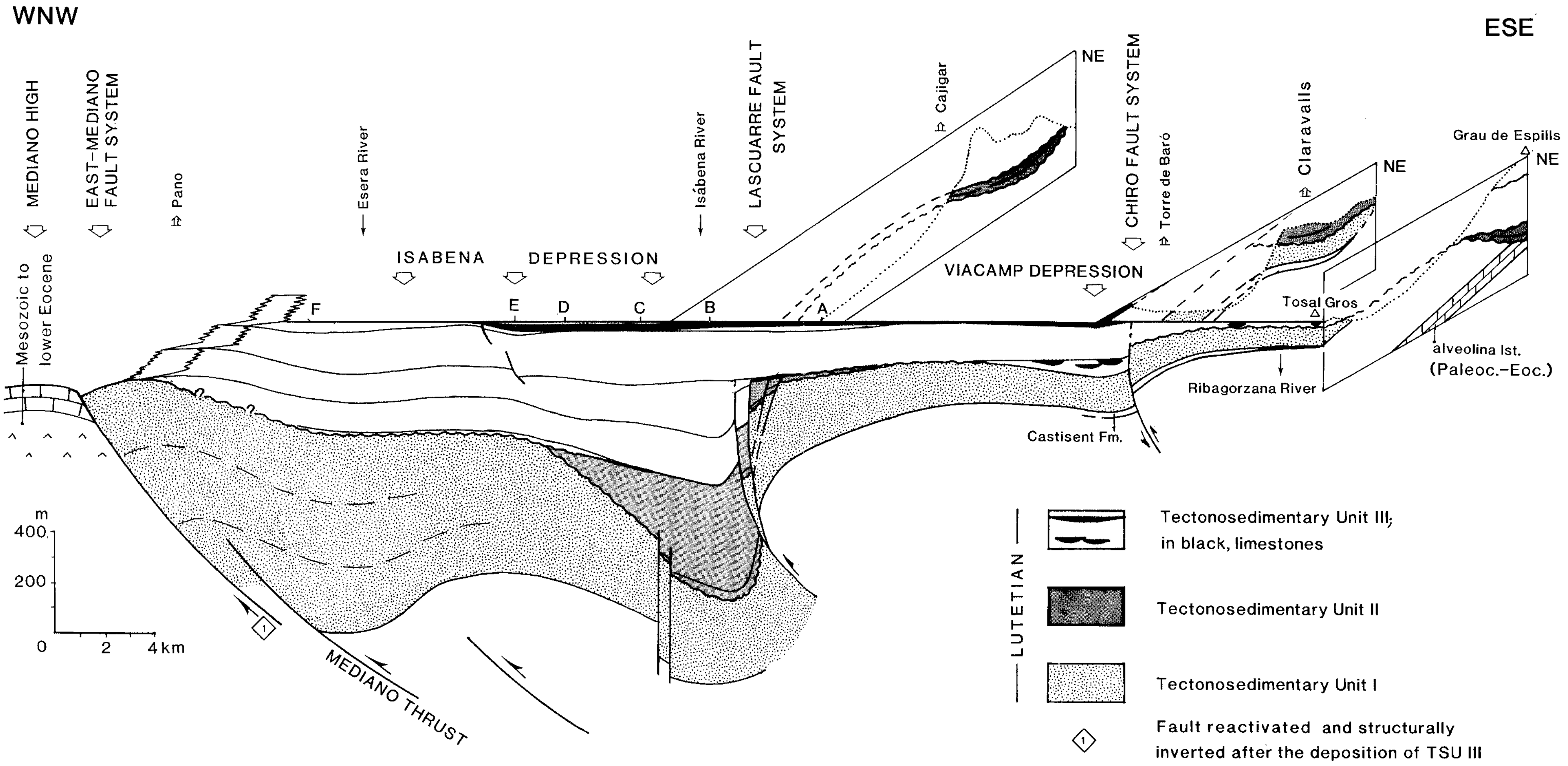
The lower tectonosedimentary unit consists of the lower part of the Capella Formation.

2) An upper tectonosedimentary unit (U.V.TSU) consisting of conglomerates, sandstones, mudstones and some limestone levels. It is characterized by a greater variation of the composition of the coarse detritic components: fragments of dolerite, metamorphic rocks and occasionally granite, as well as sedimentary rocks, are found (Kunst in Mijnssen, 1986; polimictic conglomerates of Werver, 1987; open file reports). This unit consists of the upper part of the Capella Formation and the Viacamp Formation.

The contact between the tectonosedimentary units is an angular, erosive, paleovalley-shaped unconformity as can be observed near Torre de Baró (fig. 4.1). The lower unit is tilted southwards and fractured into blocks which are erosively overlain by the basal conglomerates of the upper tectonosedimentary unit (fig. 4.1).

Moreover, at the eastern margin of the Ribagorzana valley, in the Tosal Gros hill, Mijnssen (open file report, 1986) recognized two lithostratigraphic units within the facies of the Capella Formation (fig. 4.2): the "Tosal Gros Mudstones" and the "Tosal Gros Conglomerates". Based on the lithology of the conglomeratic clasts, he correlated these units with the "lower" and the "upper" Capella Formation, respectively, at the western margin of the valley. A subtle angular unconformity is recognized between these two units at the Tosal Gros hill. This unconformity is correlated with the unconformity at Torre de Baró; the Tosal Gros Mudstones unit is included in the lower tectonosedimentary unit and the Tosal Gros Conglomerates unit in the upper tectonosedimentary unit (figs. 4.1 and 4.2).

Fig. 4.1.- *(next page) Correlation profile of Lutetian stratigraphic units within the Tremp-Graus Basin. The Escanilla basal limestones and the Viacamp upper limestones are used as horizontal reference levels. For the sake of the correlations, the deposits and the tectonic structures younger than these limestones are, in general, not represented. Only east of the Mediano High are some younger structural features depicted.*



WNW
ESE
MEDIANO HIGH
EAST-MEDIANO FAULT SYSTEM
Pano
Esera River
Isábena River
LASCUARRE FAULT SYSTEM
Cajigar
NE
CHIRO FAULT SYSTEM
Torre de Baró
Claravalls
NE
Grau de Espills
NE
ISABENA DEPRESSION
VIACAMP DEPRESSION
Mesozoic to lower Eocene
F
E
D
C
B
A
Tosal Gros
alveolina lst. (Paleoc.-Eoc.)
Ribagorzana River
Castisent Fm.
m
400
200
0
2
4 km
MEDIANO THRUST
1
LUTETIAN
Tectonosedimentary Unit III, in black, limestones
Tectonosedimentary Unit II
Tectonosedimentary Unit I
1
Fault reactivated and structurally inverted after the deposition of TSU III

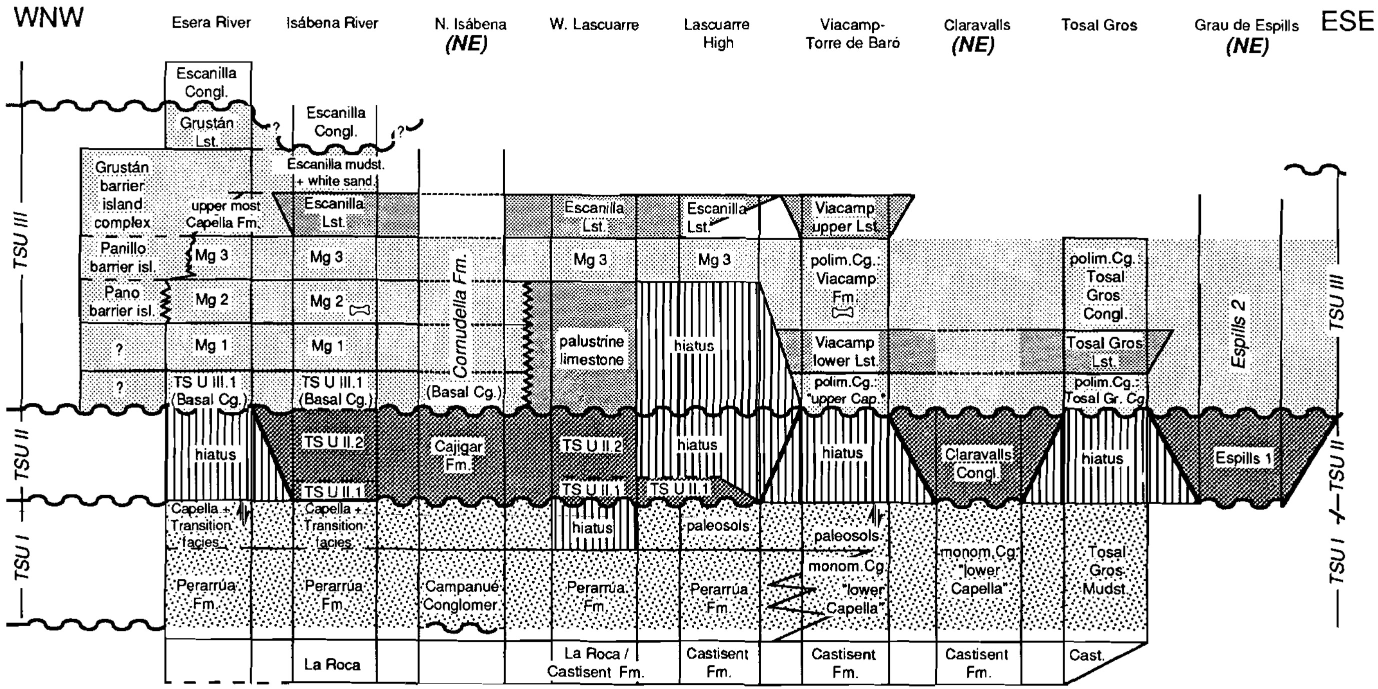


Fig. 4.2.- *Correlation diagram of Cuisian to Lutetian stratigraphic units within the Tremp-Graus Basin.*

2.2.- Correlation of tectonosedimentary units between the Viacamp Depression and the Isábena Depression.

The two tectonosedimentary units observed in the Ribagorzana valley (Viacamp Depression) are compared and correlated with those observed in the Esera valley (TSU I and TSU III; Isábena Depression).

The lower tectonosedimentary unit of the Viacamp Depression (L.V.TSU) is correlated with TSU I. The upper tectonosedimentary unit of the Viacamp Depression (U.V.TSU) is correlated with TSU III (figs. 4.1 and 4.2). The common features between these tectonosedimentary units are summarized as follows:

a) Both L.V.TSU / U.V.TSU and TSU I / TSU II are separated by an angular and erosive unconformity.
b) Both L.V.TSU and TSU I have a more marine character than the overlying U.V.TSU and TSU III, respectively.
c) Both L.V.TSU and TSU I show a regressive sequence, with a vertical evolution to finer and shallower facies. Ochre mudstones, either with isolated carbonate nodules or ruiniform calcitic soil horizons form the top of the sedimentary succession in both cases.
d) Both L.V.TSU and TSU I were tectonically deformed and fractured intoblocks prior to the deposition of the overlying U.V.TSU and TSU III, respectively.
e) The E-W to NE-SW trending fault systems which fractured L.V.TSU and TSU I were also active during the deposition of the basal layers of the respectively overlying U.V.TSU and TSU III.

If we accept this correlation, TSU I is recognized as a basin-wide tectonosedimentary unit which, in the Isábena Depression, comprises the Campanué-Perarrúa Formation and the lowermost deposits of the Capella Formation, and in the Viacamp Depression, the lower Capella Formation and the Tosal Gros Mudstones unit.

Tectonosedimentary Unit I is characterized by a westwards transition from continental to shallow marine facies and by a general shallowing- and finning-upward sequence at the top of the unit. The calcitic paleosols formed at the top of the sequence are characterized by ruiniform horizons in the more continental areas of relatively low subsidence (Lascuarre High and east of the Chiro fault system) and by levels with disorthic nodules in the more seaward areas of relatively high subsidence (Isábena Depression).

The upper boundary of TSU I is defined by the unconformity of the Esera River-Bco. Lereu-Torre de Baró-Tosal Gros. TSU I corresponds with the lower part of the Santa Liestra depositional sequence of Mutti et al. (1985).

TSU III is also a basin-wide tectonosedimentary unit which, in the Isábena Depression, comprises the upper part of the Capella Formation and the lower part of the Escanilla Formation. In the Viacamp Depression, the upper Capella Formation and the Viacamp Formation, and in the eastern Ribagorzana valley, the Tosal Gros Conglomerates unit (fig. 4.1).

The fact that TSU II does not occur either in the Esera valley or in this part of the Ribagorzana valley is probably due to the laterally restricted, paleo-valley-shaped geometry of this tectonosedimentary unit. However, TSU II is recognized in more northern parts of the Ribagorzana valley, see section 2.3.

Paleontological evidence for the correlation of the upper tectonosedimentary unit of the Viacamp Depression with TSU III of the Isábena Depression is supplied by the rodent fauna of Torre de Baró, which indicates the same age as the rodent fauna of Barranco Estarán (López, Daams and Van der Meulen, pers. com.). The Torre de Baró vertebrate locality occurs in the Viacamp Depression, above the Viacamp basal limestones (U.V.TSU). The Barranco Estarán vertebrate locality appears in the Isábena Depression, at the base of Megasequence 2, within TSU III.

The paleontological correlation indicates that the basal part of the Viacamp Formation is the time equivalent of the upper part of the Capella Formation in the Isábena Depression (fig. 4.2).

Correlation of lithofacies within Tectonosedimentary Unit III

Above the Torre de Baró vertebrates locality the sedimentary sequence continues with a thick pile of coarse facies (Viacamp polimictic conglomerates of Werver, 1987, open file report) overlain by a new interval of lacustrine limestones (Viacamp upper limestones). Above the Barranco Estarán vertebrate locality the sequence is characterized by a sandstone-rich unit (Megasequence 2), a more conglomeratic unit (Megasequence 3) and an interval of lacustrine limestones (Escanilla limestone of Garrido Megías, 1968). If a correlation between the lacustrine limestones at the top of both sequences is assumed, the Megasequences 2 and 3 in the Isábena Depression are the lateral equivalent of the detritic facies of the Viacamp Formation. The paleogeographic picture during the sedimentation of the limestones would be that of two lakes separated by an emergent area, the Lascuarre High, where palustrine facies were deposited.

2.3.- Tectonosedimentary units in the northern margin of the basin.

The proximal facies of the northern depositional systems which infilled the Isábena and Viacamp Depressions are represented in the northern margins of the basin by thick intervals of conglomeratic deposits. Within these conglomeratic facies several lithostratigraphic units have already been described (Cuevas et al., 1985); generally, the boundaries between the formations are unconformities. With one exception, field correlation of proximal facies with distal facies in the central part of the basin is difficult. Correlations can only be carried out by recognizing unconformity-bounded tectonosedimentary units (fig. 4.1).

The Campanué conglomerates outcrop at the western margin of the Isábena River. They grade into and can be correlated in the field with the Perarrúa Formation (Garrido Megías, 1968; Nijman and Nio 1975; De Boer, open file report, 1976) (fig. 2.3). The Campanué conglomerates form part of TSU I.

Further north and at the eastern margin of the Isábena River, two Eocene conglomeratic units unconformably overlie older Tertiary facies: the Cajigar Formation and the younger Cornudella Formation (figs. 4.1 and 4.2). The coarse detritic components of the Cajigar conglomerates consist of several types of limestone fragments. Transported nummulites are common in the matrix. The Cornudella conglomerates consist mainly of dark grey lime-mudstone and granite fragments (Koops and Van Rossem, 1985). The Cajigar Formation, characterized by reddish conglomerates, is deformed by the Cajigar syncline. The northern limb of the syncline is uncomformably overlain by the Cornudella Formation. The angular unconformity grades southwards to a parallel unconformity (Koops and Van Rossem, 1985) (fig. 4.1).

Further east, in the area of Claravalls, a reddish conglomeratic unit lies unconformably above older Tertiary deposits (including the Castisent Formation and the lower Capella Formation). Several types of limestones are recognized among the coarse components of the Claravalls conglomerates; nummulites are recognized in the conglomerate matrix (Van der Bilt, pers. com.). This conglomeratic unit is deformed by the syncline of Soliva (fig. 4.1).

At the eastern margin of the Ribagorzana River, three Eocene conglomeratic units are observed to unconformably overlie older Tertiary deposits: the Espills-1, the Espills-2 and the Espills-3 units (Cuevas et al., 1985; Mijnsen, open file report, 1986). The Espills-1 unit is characterized by reddish conglomerates consisting of sedimentary rock fragments; no granite fragments are found. Transported nummulites occur in the matrix (Van der Bilt, pers. com.). The Espills-2 conglomerates contain dolerite and occasionally granite fragments, as well as sedimentary rock components. The Espills-1 conglomerates are preserved in a synform below the unconformable Espills-2 conglomerates (fig. 4.1).

To summarize, isolated, extensive outcrops of conglomerates are observed in the northern margin of the Tertiary basin. The outcrops are characterized by a lower reddish conglomeratic unit that was deformed in a smooth synclinal structure prior to the deposition of an upper "polimictic" greyish conglomeratic unit. The contact between the reddish and the greyish conglomeratic units is a clear angular unconformity in the northern part; southwards the unconformity tends to disappear.

Based on the facies and on the stratigraphic and structural features, two tectonosedimentary units are recognized:

1) A lower tectonosedimentary unit, which is characterized by reddish conglomerates dominated by calcareous components. It consists of the Cajigar Formation, the Claravalls conglomerates and the Espills-1 conglomerates.
2) An upper tectonosedimentary unit, which is characterized by greyish conglomerates comprising components of sedimentary rocks and others (granite, dolerite, etc). It consists of the Cornudella Formation and the Espills-2 conglomerates.

Supply systems within the tectonosedimentary units of the northern margins.

A direct petrological correlation can be established within the lower tectonosedimentary unit between the Espills-1 conglomerates and the Claravalls conglomerates. However, a slight difference in facies is observed with the Cajigar conglomerates. The Cajigar conglomerates present a smaller particle size and a less variation in the composition of the carbonatic clasts (Van der Bilt, pers. com.). It is suggested that at least two alluvial fans supplied material to the basin from the north: the Espills-1 fan in the Ribagorzana valley and the Cajigar fan in the Isábena valley.

A similar situation is observed for the upper tectonosedimentary unit: the Espills-2 conglomerates contain abundant bioclastic-limestone fragments, some metamorphic and dolerite fragments and only occasionally granite fragments (Mijnssen, open file report, 1986). The Cornudella conglomerates contain bioclastic (alveolina)-limestone fragments only at the base; the rest is dominated by grey-micrite fragments and granite fragments (Koops and Van Rossem, 1985). It is suggested that in this period two systems supplied material to the basin from the north: the Espills-2 alluvial fan in the Ribagorzana valley and the Cornudella alluvial system in the Isábena valley. The proximal parts of the supply systems were separated from each other by a structural high. The high was situated between the Ribagorzana valley and the Isábena valley. Part of this structural high is preserved and forms the eastern wall of the Cornudella paleovalley (Koops and Van Rossem, 1985). In Claravalls no evidence of the Cornudella-Espills-2 tectonosedimentary unit is observed, probably because this was an intervalley high.

The northwestwards trending fluvial system observed by Koops and Van Rossem (1985) to interfinger with the northern Cornudella alluvial system could represent either a lateral finer part of the Espills-2 system or an axial system perpendicular to the latter.

2.4.- Correlations between tectonosedimentary units of proximal (northern) areas and tectonosedimentary units of distal (southern) areas.

The northern tectonosedimentary units are correlated with those of the Viacamp and Isábena Depressions (figs. 4.1 and 4.2):

1) The Tosal Gros Conglomerates unit is correlated with the Espills-2 conglomerates based on petrological criteria (Mijnssen, open file report, 1986). This correlation suggests that the the upper tectonosedimentary unit of the northern margins form part of TSU III, as the Tosal Gros Conglomerates unit does; i.e., the Cornudella and the Espills-2 conglomerates form the proximal part of TSU III. The Cornudella system supplied material to the Isábena Depression, while the Espills-2 fan mainly supplied the Viacamp Depression. Actually, the Capella Formation, above the unconformity of the Esera River-Bco. Lereu (TSU III, Isábena Depression), has many features in common with the northern Cornudella system as described by Koops and Van Rossem

(1985). These are summarized as follows:

a) Petrology:
   * Most frequent pebbles are dark grey micrite limestone.
   * Granite pebbles occur occasionally. Subarkosic sandstones are common.
   * Alveoline-limestone cobbles and boulders are found in the basal subarkoses, overlying the unconformity.

b) Sequence:
   In the Capella Formation three large sequences of coarsening an then slightly fining-upwards sediments are recognized overlying the basal conglomeratic subarkoses. In the northern Cornudella system three coarsening-fining sequences can also be recognized from the coarse/fine detritics ratio (A, B in figure 9.6 of Koops and Van Rossem, 1985).

c) Paleocurrents:
   The largest part of the Capella Formation which is included in TSU III was deposited by a fan delta system with NE-SW direction of progradation. Similar paleocurrent directions characterize the northern Cornudella system.
   However, in the upper part of TSU III (Megasequence 3) the axial fan system supplied material from the Viacamp Depression to the Isábena Depression as inferred from paleocurrent directions (ESE-WSW) and petrology (alveoline limestone and sandy limestone pebbles).

2) The lower tectonosedimentary unit of the basin's northern margins is situated stratigraphically above TSU I at Claravalls, and below TSU III in the northern Isábena and Ribagorzana valleys, and it can therefore be correlated with TSU II as defined in the Isábena Depression. The common features between the lower tectonosedimentary unit of the basin northern margins and TSU II in the Isábena Depression can be summarized as follows:

   a) the red colour of the facies,
   b) the coarser character of the unit when compared with the overlying tectonosedimentary unit,
   c) the petrological composition of the pebbles and cobbles, consisting basically of limestone fragments,
   d) the dominant paleocurrent directions towards the southwest, and
   e) the preservation in a structural synform.

2.5.- Proposed correlations

The recognition of tectonosedimentary units allows a detailed correlation of the Capella Formation in the Isábena Depression with other deposits in the Viacamp sub-basin and in the basin's northern margins. Some tectonosedimentary units of basinal extent are proposed; the units include several of the traditionally accepted lithostratigraphic formations.

Tectonosedimentary Unit I includes:
in the Isábena Depression:
 the Campanué conglomerates (or a part of it),
 the Perarrúa Formation (or a part of it) and
 a very small part of the Capella Formation;
in the Viacamp Depression:
 the lower part of the Capella Formation and
 the Tosal Gros Mudstones unit (Ribagorzana valley).

Tectonosedimentary Unit II includes:
in the Isábena Depression:
 the Cajigar Formation and
 the middle Capella Formation;
in the Viacamp Depression:
 the Espills-1 conglomerates (northern Ribagorzana valley).

Tectonosedimentary Unit III includes:
in the Isábena Depression:
 the Cornudella Formation,
 the upper Capella Formation,
 the basal part of the Escanilla Formation,
 the Pano Formation and probably the Grustán Formation as well;
in the Viacamp Depression:
 the Espills-2 conglomerates (northern Ribagorzana valley),
 the Tosal Gros Conglomerates (Ribagorzana valley),
 the upper part of the Capella Formation and
 the Viacamp Formation (former Escanilla Formation).

Besides these tectonosedimentary units of basinal extendt, in the Viacamp Depression it is possible to discriminate other lithostratigraphic units separated by unconformities (Werver, open file report, 1987); these unconformities are not observed in the Isábena Depression.

3.- BASIN DYNAMICS OF THE ISABENA DEPRESSION

3.1.- The Isábena Depression

The Isábena Depression extends about 19 km in an ESE-WNW direction, from Lascuarre in the east to Pano in the west (fig. 4.1). The eastern margin of the Isábena Depression is defined by the Lascuarre fault system. This fault system outcrops west and southwest of Lascuarre, with a N-S direction (fig. 2.1). The sedimentation of the Capella Formation was controlled for a large part by this faulting system, as will discussed later in more detail. It is assumed that the Lascuarre fault system is related to a lateral ramp of the Montsech thrust sheet as delineated by Cámara and Klimowitz (1985).

The western boundary of the Isábena Depression is a system of faults observed between the village of Pano, in the Isábena Depression, and the diapiric Mediano High in the west (fig. 4.1). Cámara and Klimowitz (1985) delineated a thrust tip-line eastwards of the Mediano High, coinciding with the western margin of the Isábena Depression. The present study suggests that the Pano Formation, in the most western part of the depression, was formed on a structural high. The structural high could be a ramp anticline related to the Mediano thrust line as delineated by Cámara and Klimowitz (1985) (fig. 4.1). In which case the configuration of the Isábena Depression, during the deposition of the Capella Formation, would have been comparable in E-W profile comparable to the model proposed by these authors for sedimentation areas between two ramp anticlines. In the Isábena Depression a depositional trough formed at the toe of the Lascuarre lateral ramp, coinciding with the present NNE-SSW section of the Isábena valley (figs. 1.1 and 4.1). From here westwards the basin graded into a smooth area of less subsidence on the rear slope of the Eastern-Mediano thrust-sheet.

Reactivation and structural inversion of both the Lascuarre fault system and the Eastern-Mediano fault system probably happened after the sedimentation of TSU III (fig. 4.1).

Besides the Eastern-Mediano and the Lascuarre fault systems, which greatly controlled the facies distribution and sediment thickness of the Capella Formation, other subordinate structures also affected the sedimentation. These subordinate structures are mentioned in relation with each evolutive stage.

3.2.- Evolutive stages of the Isábena Depression

Tectonosedimentary Unit I: Profile 1 (fig. 4.3)

Tectonosedimentary Unit I consists mainly of Perarrúa facies, and for a small proportion of transition and Capella facies. The shallowing sequence from Perarrúa to Capella facies resulted from a decrease of the rate of relative sea-level rise, which was probably related to tectonic movements in the basin (chapter III, part 2.2).

A relatively low sedimentation rate is indicated for the eastern part of the studied area by the smaller sediment thickness of TSU I and by the occurrence of calcitic ruiniform horizons in the upper part of TSU I. Further to the west only isolated carbonate nodules occur.

Tectonosedimentary Unit II: Profiles 2 & 3 (fig. 4.3)

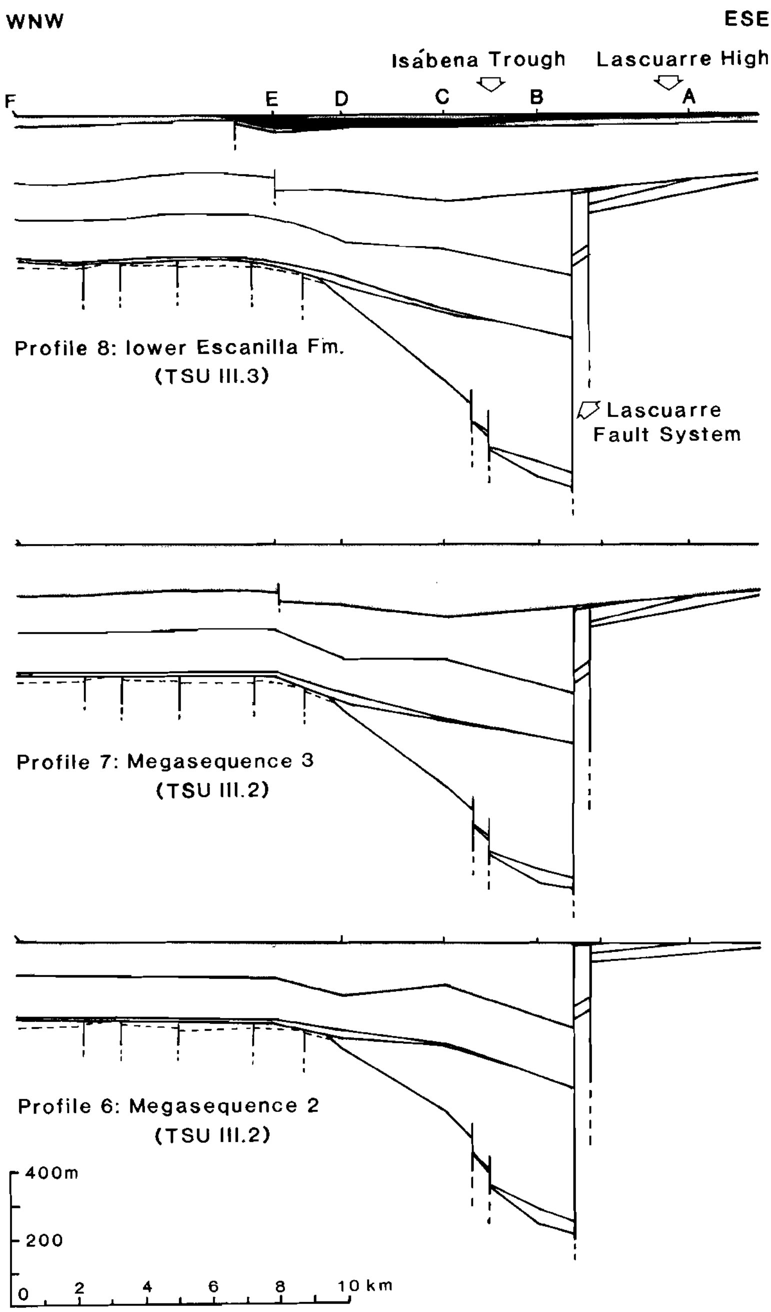
The difference in tectonic strike and dip between TSU I and TSU II indicates that TSU I was tectonically deformed prior to the deposition of TSU II. The ESE-WNW smooth folds which show the Perarrúa Formation in the Esera valley (fig. 1.1 and Langeraar, open file report, 1978; Van Dijk et al., 1985) are probably due to compressive tectonic deformation prior to the deposition of TSU III and probably also prior to TSU II.

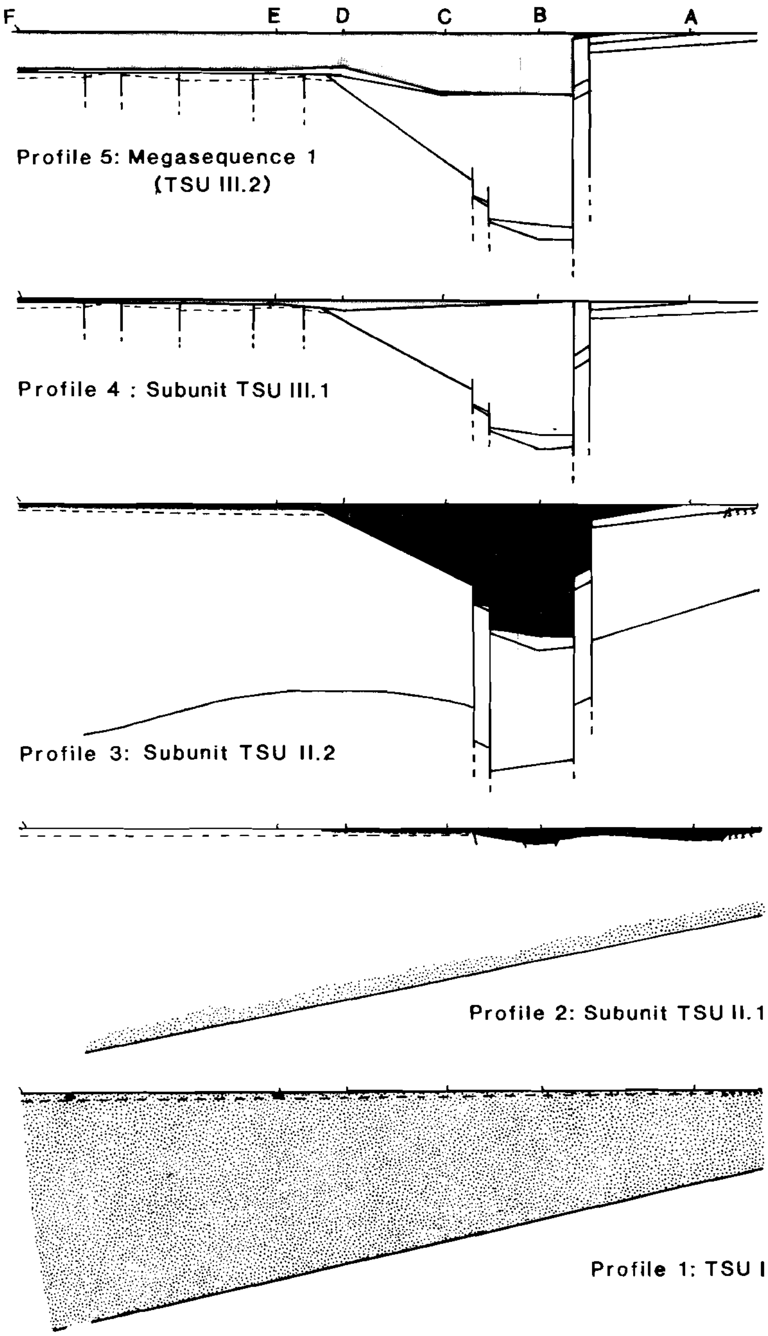
A change in the basin towards a more extensive situation is inferred between TSU I and TSU II from the displacement of the proximal facies (Campanué conglomerates in TSU I and Cajigar Formation in TSU II) away from the basin center. Both the shallowing sequence observed at the top of TSU I and the expansion of the basin observed between TSU I and TSU II suggest a change in the tectonic evolution of the basin (a decrease of the rate of tectonic tilting according to the model of Megías, 1982, for tectonic ruptures).

Both the relative sea-level fall and the small-scale faulting that occurred at the end of TSU I resulted in the formation of an incipient paleovalley (profile 2, fig. 4.3). This paleovalley was infilled by the deposits of TSU II.1. During the deposition of TSU II.1, a relatively low subsidence rate characterized the Lascuarre High, as indicated by the occurrence of multi-storey channel bodies in this area.

TSU II.2 presents a pronounced depocenter westwards of the Lascuarre fault system (fig. 2.9). Both the NE-SW direction of progradation of the depositional system and the occurrence of the coarsest facies in the depocenter area suggest that the depocenter is not only the result of postsedimentary tectonics, differential erosion and preservation, but that it also has a synsedimentary origin. Probably the Lascuarre fault system as well as other faults at the western margin of the Isábena valley were already active, promoting the formation of a tectonically-controled paleovalley (profile 3, fig. 4.3).

Fig. 4.3.- *(next two pages): Dynamic evolution of the Isábena Depression during the deposition of the Capella Formation, in a series of stripped back profiles.*


WNW
ESE
Isábena Trough
Lascuarre High
F
E
D
C
B
A
Profile 8: lower Escanilla Fm.
(TSU III.3)
Lascuarre
Fault System
Profile 7: Megasequence 3
(TSU III.2)
Profile 6: Megasequence 2
(TSU III.2)
400m
200
0
2
4
6
8
10 km


F
E
D
C
B
A
Profile 5: Megasequence 1
(TSU III.2)
Profile 4 : Subunit TSU III.1
Profile 3: Subunit TSU II.2
Profile 2: Subunit TSU II.1
Profile 1: TSU I

Tectonosedimentary Unit III: Profiles 4 to 8 (fig. 4.3)

The deposits of TSU II, as well as those of the underlying TSU I, were affected by distensive tectonic deformation. This is manifested by:

1a) A system of diaclases aligned both in 110° and 45°-50° E. This association of diaclases occurs both in the Transition facies of TSU I (sections B and E) and in TSU II (section B) but it has not been observed in the overlying TSU III.

1b) A system of normal faults and monoclinal flexures, aligned about E-W, affected the deposits of TSU I from section C to section F (fig. 1.1 and profile 4 in fig. 4.3). A southern downthrow is most common in these structures. Some of the faults are fossilized by deposits of TSU III.
NE-SW trending faults also affected the deposits of TSU I and TSU II. However, it is not evident whether these faults were active prior to the deposition of TSU III.

2) Southwest of Lascuarre the deposits of TSU II are affected by small normal NW-SE faults with a southwestern downthrow. Faults present a distinct slurring, suggesting a low degree of consolidation of sandstones and conglomerates. The faults are fossilized by deposits of TSU III.

The Lascuarre fault system affected the deposits of TSU II. However, it is difficult to determine exactly when the activity of this fault system started and when it had its maximum intensity. The Lascuarre High, the eastern upthrown block of the Lascuarre fault system, is thought to have been an area of relatively low subsidence during TSU II. The Lascuarre fault system's influence on the sedimentation of a large part of TSU III, is discussed next.

Most of the subunits and megasequences which form part of TSU III are present from the most westward part of the studied area (section F) up to the Lascuarre fault system (figs. 2.4 and 2.9). Only the upper part of TSU III (Megasequence 3 and subunit TSU III.3) is present on the Lascuarre High (section A). Whether the subunit TSU III.1 and the Megasequences 1 and 2 were ever deposited on the Lascuarre High is not known. The only deposit found here between the fractured conglomerates of TSU II and the deposits of Megasequence 3 is a thin level of "palustrine" limestones (cp. Freytet, 1964). It is assumed that the limestones were deposited on the Lascuarre High while the detritic deposits forming TSU III.1 and/or the Megasequences 1 and 2 were deposited in the depression itself. It is not known which of these lithostratigraphic levels corresponds with the limestones.

Subunit III.1: Profile 4 (fig. 4.3)

The depocenter and the coarsest facies of this subunit are localized at sections D and E; the depositional system had a NE-SW direction of progradation.

This subunit overlies the unconformity at the base of TSU III and forms the "basal conglomerate" of this tectonosedimentary unit. Like its northern proximal eqivalent, the lowermost part of the Cornudella Formation, this subunit contains coarse fragments eroded from the underlying conglomeratic fan systems. These fragments are not characteristic of the new depositional system.

The subarkosic deposits of this subunit fossilize E-W trending faults affecting the underlying tectonosedimentary units. However, the subarkoses are also partially deformed by the same faulting system, indicating the continuity of these tectonic movements at the base of TSU III.

Subunit III.2

The Pano Formation forms the coastal and shallow-marine equivalent of the Capella Formation in a westwards direction (Nijman and Nio, 1975; Donselaar and Nio, 1982; Cuevas et al., 1985).

Within the Pano Formation three stages (Pano, Panillo and Grustán) are recognized, which correspond with three subsequent positions of barrier-island complexes during the transgression (Donselaar and Nio, 1982). During conditions of a constant rate of relative sea-level rise, barrier islands need an increasing rate of sediment supply to build-up and maintain them in place. If the relative sea-level rise is not accompanied by an increase of sandy sediment supply, the barrier island migrates landwards (Donselaar, 1989). The different barrier-island complexes observed in the Pano Formation can be correlated with the megasequences observed in the Capella Formation.

The building up of the barrier-island complexes took place during phases of relative sea-level rise which were accompanied by an increasing rate of sediment supply; i.e., the barrier islands grew up at the same time as the coarse-grained parts of the megasequences were being deposited. At the top of each megasequence the rate of relative sea-level rise exceeded the rate of sedimentation (chapter III, part 2.4.2), which resulted in the tidal reworking of the alluvial channels and the beginning of the barrier-island drowning. The transition to a new megasequence is determined by a reduction of sand supply to the basin, as a result of the retrogradation of the depositional systems (lower fine-grained part of the megasequences). It is suggested that under these circumstances the barrier island could not survive the drownning process which had already started and it was progressively destroyed by shoreface erosion. At the same time, a new barrier nucleus started to form in a more landward position. This barrier nucleus developed into a mature barrier-island complex during the following phase of relative sea-level rise associated with an increasing rate of sand supply (coarse-grained part of the megasequences).

From this point of view, and considering that Megasequence 3 is overlain by the uppermost barrier-island complex (Grustán), it is proposed that the lower barrier-island complex (Pano) corresponds with the coarse-grained part of Megasequence 2 and the middle barrier-island complex (Panillo) corresponds with the coarse-grained part of Megasequence 3 (fig. 2.7). It is also suggested that the Grustán barrier-island complex corresponds with sandstones of the Escanilla Formation (levels 2 and 3 of TSU III.2; chapter II, part 3.4). The sedimentation of the Escanilla lower limestones would correspond with the migration episode from the Panillo stage to the Grustán stage.

The sedimentation of Megasequence 1 might have been limited to the west either by a structural high or by a new barrier-island complex, corresponding to an earlier stage of the Pano Formation (fig. 4.2). None of these features have been described yet; fracturing and poor exposure are common in the area.

The main problem for the correlation between the stages of the Pano Formation and the megasequences of the Capella Formation is the difference in sediment thickness. The Pano Formation has a maximum thickness of about 300 m, while the Capella Formation at the western margin of the Esera River (section F) reaches a thickness of 470 m, of which about 260 m correspond to Megasequences 2 and 3. To explain the difference in thickness it is suggested that the position of the coastal area was structurally controlled. Probably, the Pano Formation records the deposition of sand in high-energy shallow marine conditions over the roughly N-S oriented crest of a ramp anticline. The Capella Formation, on the other hand, shows mud and sand deposition in a protected coastal plain on the rear slope of the ramp anticline. A comparable structural control of coastal facies is suggested by Noe-Nygaard and Surlyk (1988) for the Cretaceous Robbedale and Jydegard Formations of Bornholm (Denmark).

Megasequences 1 and 2: profiles 5 and 6 (fig. 4.3)

Both Megasequence 1 and Megasequence 2 present a depocenter westwards of the Lascuarre fault system. Sediment thickness decreases gradually from here to the west and is steady from section E to section F, suggesting a platform of less subsidence in this area. Local thickness variations are related to the activity of synsedimentary faults and to the lateral shifting of the depositional systems. The area around section D stands out as an unstable area, with frequent synsedimentary faulting and anomalous sediment thicknesses (figs. 3.15 and 3.18). Synsedimentary faults on the scale of the sediment bodies trend ENE-WSW with a northern downthrow. However, larger synsedimentary faults with more northerly trends are suggested by the abrupt sediment thickness variations in an E-W direction.

Megasequence 3: profile 7 (fig. 4.3)

Megasequence 3 is recognized throughout the studied area. This megasequence presents a depocenter at section C; i.e., the depocenter was displaced westwards in relation to the earlier megasequences. At the time of deposition of Megasequence 3 the Lascuarre fault system had little influence on the sedimentation and the Lascuarre High did not form a threshold. Conglomeratic deposits, probably supplied by the Espills-2 fan system in the Viacamp Depression, entered from the east into the Isábena sub-basin, were they mixed with sediments supplied by the Cornudella fan system from the northeast.

Synsedimentary tectonic activity is evidenced by the presence of synsedimentary faults in Megasequence 3. Faults are recognized at two scales:

a) On the scale of the sediment bodies: These are in general normal faults, with a vertical displacement of decimeters to meters. They influence mainly the thickness and facies distribution of sediment bodies. Some of the faulting trends are: 1) NW-SE trending faults in unit 2; 2) E-W to ESE-WNW trending faults in unit 3.

b) Larger NE-SW trending faults: These faults have vertical displacements of several tens of meters and affect a thick part of the sedimentary succession. Synsedimentary faults with this strike affected the sedimentation of unit 2 and probably also that of unit 3 (fig. 3.17). Faults with this orientation are common in the study area (fig. 2.1). However, most of them were active after the deposition of the Capella Formation and it is difficult to recognize whether they were also active at the time of deposition of Megasequence 3.

Escanilla lower limestones and uppermost Capella Formation: profile 8.

Overlying the deposits of Megasequence 3 and during a period of relative tectonic tranquillity and low sediment supply, grey marls and limestones were deposited in the central and eastern part of the basin. Towards the west, and possibly separated from the central part by a structural high, clastic lagoonal facies were deposited.

The maximum limestone thickness was deposited in the central area (sections C, D and E), while towards the east the limestone thickness decreased and interbedded fine clastic deposits in the limestones became more important. The central area was dominated by lacustrine limestones with the possible influence of brackish water (Nickel, 1982), while pedogenetic facies (palustrine limestones) occurred eastwards of section A and at the base of the limestone interval in section B. Thickness and facies distribution as well as the eastwards onlapping of the limestones (see chap. III, part 3.4) indicate that the lake had an eastern shallow to emergent area on the Lascuarre High.

Summary and interpretation

A) During the deposition of the largest part of the Capella Formation, the Lascuarre High was an elevated area, where little sedimentation took place and pedogenetic features developed well.

B) The Isábena trough was an area of relatively high-subsidence during the deposition of the Capella Formation. However, subsidence was rapidly compensated by sedimentation, resulting not in deeper, more marine facies, but in the largest sediment accumulation. Westwards sediment thickness decreased gradually.

C) Besides the large N-S trending structures, which conditioned the sedimentation of the Capella Formation and which are assumed to be directly related to thrust planes, other forms of tectonic deformation are also observed. The following possible relations between the deformation structures and the displacement of thrust sheets are suggested. Further structural work would need to be done to confirm them.

1- The ESE-WNW trending folds observed in the Perrarúa facies (TSU I) may be related to the southward displacement of the ESE-WNW trending western frontal ramp of the Montsech thrust as delineated by Cámara and Klimowitz (1985) to the north of the Campanué-Perrarúa Formation.

2- The small faults with general E-W trends and southern downthrow, which are observed from the top of TSU I to the base of TSU III and which are more common in the western half of the sub-basin, may represent a period of expansion after the compressive deformation generated in 1-.

3- The N-S oriented Lascuarre fault system was a compressive dextral strike-slip fault related to a similarly trending lateral ramp of the Montsech thrust-sheet. The predominance of N-S trending faults over E-W trending faults in the eastern half of the depression suggests that this area was more influenced by tangential movements.

4- Normal faults striking about E-W and with northern downthrow are characteristic of TSU III. These faults may represent distension structures formed on the rear slope of a southward displacing thrust sheet; i.e., they suggest the presence of a southern thrust-sheet, which according to Cámara and Klimowitz (1985) correlates with the Eastern-Mediano one. The NE-SW faults observed within TSU III might have resulted from differential movements in the southward advance of the thrust sheet.

This work supports Van Hoorn's (1970) suggestion that the facies distribution of other formations in the Tremp-Graus Basin could have been influenced by large N-S trending structures in the same way as the Capella Formation. The position of the thrust-related structures is not fixed in time. Therefore the lines of major lateral variations in facies and sediment thickness ("break lines") can move laterally through the stratigraphic record.

CHAPTER V: CONCLUSIONS

1.- BASIN GEOMETRY

The sedimentation of the Capella Formation was affected by synsedimentary faults, which influenced the distribution of sediments. The distribution of sediment thickness and facies, and the recognition in the field of several tectonic structures allows the discrimination within the Tremp-Graus Basin of several depressions controlled by N-S trending faults, i.e., the Isábena Depression in the western part and the Viacamp Depression in the eastern. These depressions existed at the time of sedimentation of the Capella Formation.

Within the western part of the Tremp-Graus Basin the most conspicuous elements are:

a) The Lascuarre fault system. This is a system of faults striking approximately N-S, which conditioned the distribution of sediment thickness and facies of a large part of the Capella Formation. The Lascuarre fault system is located between the Isábena valley and the village of Lascuarre.
b) The Lascuarre High was an elevated area, where little sedimentation took place and pedogenetic features developed well. The Lascuarre High formed on the eastern, upthrown block of the Lascuarre fault system, and separated the Isábena Depression from the Viacamp Depression.
c) The Isábena trough was an area of high subsidence characterized not by the deeper, more marine facies, but by the largest sediment accumulation. The Isábena trough formed on the western, downthrown block of the Lascuarre fault system.
d) The existence of a structural high is suggested at the western margin of the Isábena Depression, below the Pano Formation.
e) Besides these large structures, smaller faults with ESE-WNW or NE-SW strike influenced the sedimentation of the Capella Formation on a smaller scale.

2.- TECTONOSEDIMENTARY ANALYSIS

The recognition of unconformities and their correlative conformities has allowed the subdivision of the Capella Formation into three tectonosedimentary units (TSU I, TSU II and TSU III), according to the definition of Megías (1982).

3.- RELATIONS BETWEEN FORMATIONS

Perarrúa Formation-Capella Formation relation

The Capella Formation was considered to be, in its totality, the lateral equivalent of the shallow-marine Perarrúa Formation (Nijman and Nio, 1975). The present study agrees that the Perarrúa Formation had a more continental equivalent (Capella Formation) in eastern areas. However, the largest part of the Capella Formation in the study area is unconformable with the underlying Perarrúa Formation. The Perarrúa Formation and the lower part of the Capella Formation form part of TSU I, while the middle and the upper part of the Capella Formation form part of TSU II and TSU III, respectively.

The shallowing (regressive) sequence indicated at the top of TSU I by the transition from the Perarrúa Formation to the Capella Formation did not result from the progradation of the depositional systems as suggested by Nijman and Nio (1975). Instead, it was mainly due to a decrease of the rate of relative sea-level rise, followed by a relative sea-level fall. As a result part of the basin was subaerially exposed and subsequently submitted to erosion. The changes of relative sea-level probably resulted from the tectonic deformation and uplifting of the basin.

Capella Formation-Escanilla Formation relation

The Escanilla Formation was considered to unconformably overlie the Capella Formation (Garrido Megías, 1968). In the present study no unconformity has been observed between these formations. It is believed that the lower part of the Escanilla Formation belongs to the same tectonosedimentary unit as the upper part of the Capella Formation (TSU III). The lower limestones of the Escanilla Formation in the Isábena valley are considered to be the lateral equivalent of the uppermost facies of the Capella Formation in the Esera valley.

Capella Formation-Pano Formation relation

Based on sequential criteria the Pano Formation is considered to be the lateral equivalent of the upper part of the Capella Formation and of the lower part of the Escanilla Formation. This sedimentological interpretation is supported by new biostratigraphic criteria.

4.- DEPOSITIONAL SYSTEMS

The Campanué fan delta was considered to be the only Lutetian depositional system which supplied sediments to the Isábena Depression from the north (Nijman and Nio, 1975). However, within the Lutetian Capella Formation two other northern depositional systems are recognized; these depositional systems had their proximal facies in the Cajigar Formation and the Cornudella Formation:

1) The Cajigar fan delta, with some conglomeratic facies, had a NE-SW direction of progradation and was active at the time of deposition of Tectonosedimentary Unit II.

2) The Cornudella fan delta also had a NE-SW direction of progradation. Sandstones were more dominant than conglomerates. This system was active during the deposition of Tectonosedimentary Unit III.
Besides the northern Cornudella fan delta, an axial channel system with a ESE-WNW direction of progradation was responsible for the deposition of the Tectonosedimentary Unit III. The axial system formed a conglomeratic fan delta system in the upper part of the Capella Formation (TSU III).

The depositional systems were affected at their distal parts by tidal action. The tidal system eventually formed its own system of channels in the coastal plain.

5.- FACIES ANALYSIS

The sediment bodies of the Capella Formation are classified on the basis of the external geometry and the internal organization. Four major types are recognized: channel deposits, depositional lobes, sheet deposits and composite forms. A further differentiation of these groups is based on similar criteria.

Three main types of channel deposits are recognized:
1) meandering channels, 2) straight channels and 3) braided channels. Braided and straight channels characterize the northern system. Braided and meandering channels characterize the axial system. The meandering channels are often tidally-influenced. A tidal origin is not excluded for them.

Several types of depositional lobes are recognized:
- Types A1, A2 and B2 are thin, intensely bioturbated, structureless lobes either with horizontal or inclined bedding. They are interpreted either as overbank splay lobes or as the terminal lobes of channels.
- Type B1 is conglomeratic and matrix-supported. It is interpreted as a debris-flow deposit. This type occurs only locally.
- Type B3 consists of avalanche foresetting and is interpreted as a small delta deposited in a standing water body.
- Types B4, B5 and B6 are thick, cross-stratified depositional lobes with frequent indications of tidal reworking. They are interpreted as depositional lobes formed in an intertidal area at the mouths of distributary channels.

The sheet deposits are thin sediment bodies with a relatively large lateral extent; ripple-lamination is scarce but bioturbation intense. They are interpreted either as terminal lobes of small channels or as overbank deposits of channels, depending on their sequential and lateral association with other sediment bodies.

The composite forms comprise the association of channel deposits with a depositional lobe. They represent the interaction of both depositional systems.

The mud deposits, which form the 75 % of the Capella Formation are thought to have been deposited mostly on an intertidal coastal plain. The presence of grey mudstones, hydromorphic paleosols and calcitic paleosols reveals the existence of subtidal to supratidal conditions.

The architectural analysis indicates that a large part of the sheet deposits and the thin depositional lobes (types A1, A2 and B2) are sequentially related to small straight channels; they were formed in the distal part of a fan delta with low depositional slope. The largest depositional lobes (types B4 to B6) are related to larger channels; they formed in the intertidal, distal part of a high-slope fan delta. The thickness of depositional lobes with tidal reworking is frecuently around 5 m, which may indicate the predominant water depth.

6.- ARCHITECTURAL ANALYSIS

Tectonosedimentary Unit I

Tectonosedimentary Unit I exhibits a regressive sequence at the top related both to the retrogradation of the depositional systems and to the decrease in the rate of relative sea-level rise and/or the relative sea-level fall. A tectonic origin is inferred for this regression.

Tectonosedimentary Unit II

No characteristic sequence can be recognized for Tectonosedimentary Unit II because of the discontinuity of the outcrops and the limited lateral extent of the unit.

Tectonosedimentary Unit III

Three megasequences are recognized within Tectonosedimentary Unit III. The megasequences are the product of alternating stages of progradation and retrogradation of the depositional systems. These large-scale events interfered with small-scale changes in the depositional slope of the basin and with changes in the rate of relative sea-level rise.

As a result of the variations in the ratio of the rate of relative sea-level rise and the rate of sedimentation, repeated oscillations in the depth of the basin took place. The overall trend of these oscillations is transgressive, as inferred from the architectural analysis and as evidenced both by the appearance of lagoon/bay facies in the upper part of the Capella Formation and by the correlation with the Pano Formation.

Several types of episodes are observed:

S: Stationary episodes, when the rate of relative sea-level rise was compensated by the rate of sedimentation. These episodes were related to the retrogradation of the depositional systems in association with a phase of low tectonic activity, i.e., episodes with low rates of both sedimentation and subsidence. These episodes are characterized by thick intervals of mudstones with paleosols and thin sheet deposits.

R: Regressive episodes, when the rate of sedimentation exceeded the rate of relative sea-level rise. These episodes were, in general, related to the progradation of the depositional systems and are characterized by the occurrence of fluvial channel deposits.

D: Deepening episodes, resulting from an abrupt increase of tectonic subsidence. The tectonic movements also produced a reactivation of the marginal reliefs; the deepening episode was rapidly compensated by a high rate of sediment supply. These deepening episodes are recognized by the occurrence of thick depositional lobes with frequent indications of tidal influence.

T: Transgressive episodes produced when the rate of relative sea-level rise exceeded the rate of sedimentation on a coastal plain with a slope built up by infilling. These transgressive episodes are gradational and difficult to compensate by the sediment supplied to the basin. These episodes are recognized by the occurrence of tidally influenced meandering channels, by the sedimentation of progressively thinner depositional lobes, and by the occurrence of lagoon/bay grey mudstones and lacustrine limestones (table 2).

The stationary, regressive, deepening and transgressive episodes are all associated, forming sequential units. Three types of units are observed, which repeat in a sequential order within each megasequence:

Unit 1 consists of a stationary episode *S*, followed by a regressive episode *R*.

Unit 2 consists of a tectonically-induced deepening *D*, followed by a regressive episode *R*.

Unit 3 consists of a tectonically-induced deepening *D*, followed by a transgressive episode *T*.

Each megasequence consists of these three units and includes the following succession of episodes: $S+R$ - $D+R$ - $D+T$.

The transition from one megasequence to the overlying one is characterized by a shallowing episode, which resulted from a decrease in the rate of sea-level rise and/or a relative sea-level fall, accompanied by a decrease in the rate of sedimentation. The sediment record of such an episode is very thin and can not always be easily recognized or may even be missing.

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ACKNOWLEDGEMENTS - AGRADECIMIENTOS

At the conclusion of this thesis I wish to express my gratitude to all those who in one way or another have contributed to its realization.

I am indebted to prof. dr. S.D. Nio who directed this thesis. I am very grateful for the hospitable reception he gave me in the Comparative Sedimentology Division and for introducing me in the sedimentology of the tidal environment. Thanks for his constructive criticism, for his inspiring suggestions and the numerous revisions of the manuscript.

To my co-director dr. M. Díaz Molina of the Complutense University of Madrid I am indebted for all she taught me about the fluvial environment, for the manner in which she taught me to observe in the field, and especially for the pleasant way in which we discussed.

Many thanks to prof. dr. N. López of the Complutense University of Madrid and her biostratigraphy group for the very valuable information on the biostratigraphy of the area. It was a pleasure for me as a member of the Comparative Sedimentology Division, to be able to cooperate with this group in the joint project on the sedimentology and biostratigraphy of the Tremp-Graus Basin.

During the field campaigns I have collected fossil remains that are not mentioned in this thesis. I wish to thank my dear colleague dr. J. Morales Romero of the C.S.I.C. in Madrid for the study of the remains of *Lophiodon*. I regret that dr. G. Gutierrez of the University of Valencia can not hear my words of gratitude for his classification and datation of the charofites. Also many thanks to L. Granados from ENADIMSA, Madrid.

I wish to thank all members and ex-members of the Comparative Sedimentology Division of the University of Utrecht for their scientific and practical support during these past years. Thank you Sha for your geographical insight and discussions; thanks Jan-Rik, I learned a lot in Philipsdam. Also thanks to Kees Geel, Marjan Reith, Tom Zalm and Ben Driever for always lending me a helping hand when needed.

I wish to mention the students who have supplied field data of the adjacent areas: Gert, Jan, Sandra, Jos, Onno, Greg, Ronald, Chris: many thanks for your cooperation. I have tried to synthesize your data and mine into a concise basin model. It surely is not perfect, but it is one step ahead in the right direction.

J. Bergenhenegouwen, F. Quint, I. Santoe, H. Schiet and F. Trappenburg of the audiovisual department of the Institute of Earth Sciences of the University of Utrecht produced part of the figures as well as the reproduction of the photographs.

J. Senior-Rogers corrected the english manuscript. Thank you, Jackie.

Parts of the research were carried out during the tenure of a postgraduate scholarship of the Spanish Ministry of Education, and with financial assistance of the Spanish Foreign Office and the Ministry of Industry and Energy. The author held various short term contracts with the University of Utrecht during the fulfillment of the thesis. Their help is gratefully acknowledged.

A mis padres Celia y Gustavo: gracias por haberme dejado siempre seguir mi vocación y por el esfuerzo realizado para que pudiera estudiar.

Querido Rick, gracias por toda la energía que has puesto en que yo acabara mi tesis, desde Sterrenberg hasta Alt-F6.

Mis queridas "paraninfas" Amparo y Margreet, gracias por vuestra amistad y vuestro apoyo moral a qualquier hora del día. Por lo mismo, gracias a mis otras ninfas María y Evelyne.

Mis queridos amigos de los Pirineos: gracias por vuestra hospitalidad. El estar en vuestra compañía ha sido una de las partes mas agradables de este trabajo.

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