Excursion Guidebock

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Published by Servei Geològic de Catalunya

4TH INTERNATIONAL CONFERENCE ON FLUVIAL SEDIMENTOLOGY

October 2nd, 3rd and 4th, 1989. Barcelona-Sitges

FIELD TRIP 7

GEOLOGICAL SETTING

Introduction

Fluvial facies make up the Late Oligocene to Early Miocene deposits of the Loranca Basin. Greatly different examples of low and high sinuosity channel rivers are excellently exposed. All accumulate in two coalescing fluvial fans, in close association with a major compressional tectonic phase.

The selected outcrops correspond to: 1) braided palaeochannels, with preservation of composite bedforms that build up alluvial islands and sand flats; 2) longitudinal and transverse sections of point bar bodies, three dimensional outcrops of meander loops, chute channels; 3) the evolution of the Tórtola fan as the base level was rising, culminating in extensive gypsum deposits; 4) gypsum deposits with chert, in order to observe important bioturbation structures in particular.

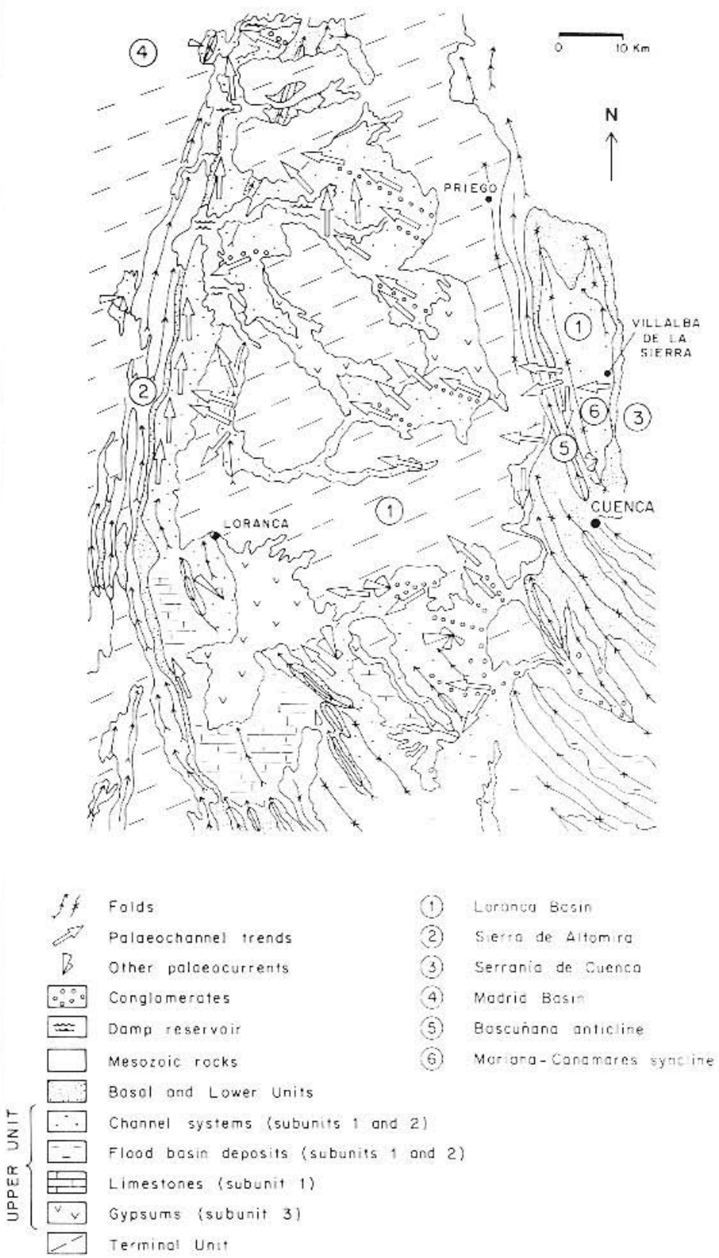
The Loranca Basin

The Loranca Basin is located in the central part of Spain (fig. 1). This basin is a marginal molasse-filled depression, originating from folding developed between the Iberian Range and the cratonic block of the Castilian Meseta (Díaz-Molina et al., 1985).

The Loranca Basin has an oval outcrop pattern (figs. 1 and 2), and is elongated along a north-south axis. The western basin margin is defined by the thrust belt of the Sierra de Altomira (fig. 2), where the Mesozoic cover has been thrust over the basement and its immediate covering (Alvaro et al., 1979). Décollement generally occurred within the «Triassic Keuper facies», but drilling shows that stratigraphic units are repeated by bedding faults which young successively towards the east (Díaz-Molina et al., 1985). Geophysical exploration reveals a large depression in the Mesozoic strata and basement, which lies between the thrust-belt and the folded Iberian Chain which forms the eastern limit of the Loranca Basin.

Figure 1
Location map of the Loranca basin.

Figure 2
Geological map and palaeocurrents for the Upper Unit.



The Loranca Basin depression formed during the Eocene (*p.p.*) to the Late Oligocene (*p.p.*). Within the basin, both pre- and syn- tectonic sediments are seen to be affected by folding when they are revealed at the outcrop beneath the younger Tertiary deposits. Along the western flanks these folds may be faulted. Furthermore, the folds have a NNE-SSW strike alignment, in the north, passing into a NNW-SSE orientation in the south (Alvaro et al., 1979).

The tectonic structures of the Mesozoic carbonate formations originated from shear stresses or pressure-dissolution. Structural analysis allows the recognition of three stages of compression that ocurred during tertiary sedimentation (Capote, 1983).

Stratigraphy

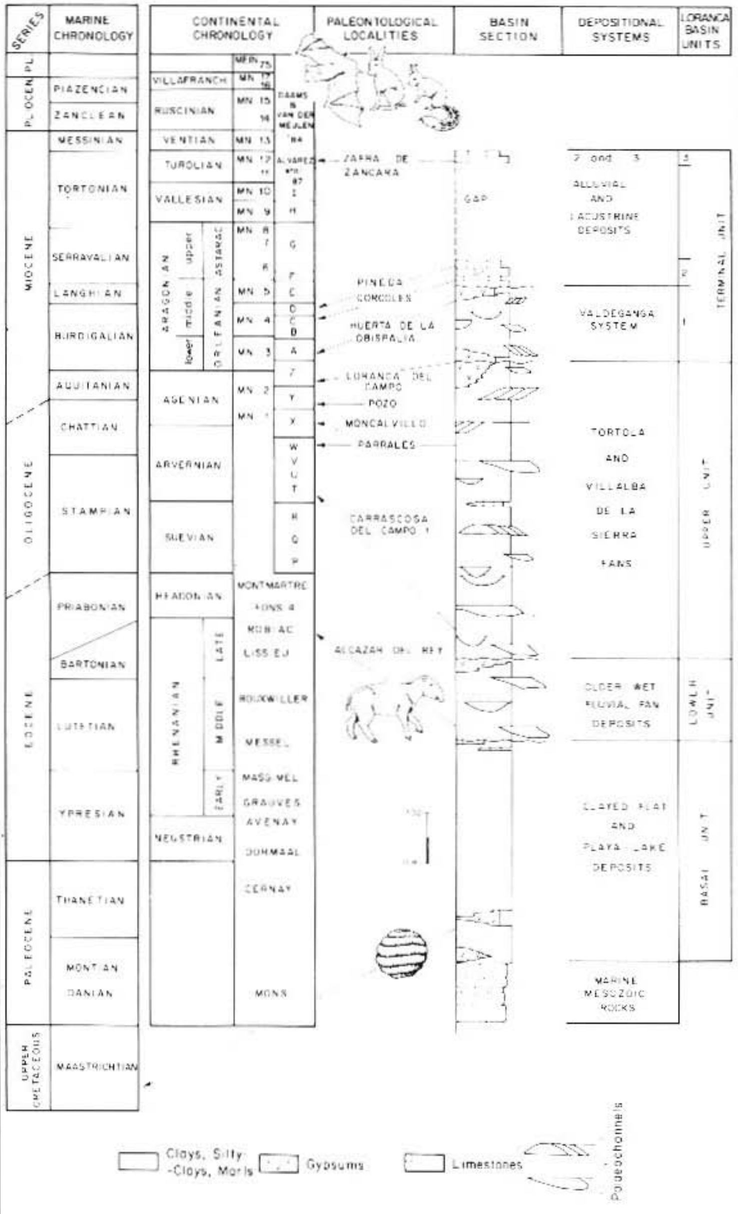
Three major stratigraphic units have been distinguished in the Tertiary stratigraphic succession of this area: the Lower Unit, the Upper Unit and the Terminal Unit (Díaz-Molina, 1974; García-Abbad, 1975; Díaz-Molina & López-Martínez, 1979; Díaz-Molina y al., 1985). The general stratigraphy of the Loranca Basin had previously been discussed by Vilas-Minondo & Pérez-González (1971); Meléndez-Hevia (1971) and Viallard (1973).

The Lower unit (fig. 3) accumulated in a basin that was more extensive than the present Loranca Basin. Its deposition was synchronous with the deformational event that formed the Sierra de Altomira and the Loranca Basin, during Eocene (p.p.) to Late Oligocene time.

The first stratigraphic unit of the Loranca Basin was the Upper Unit (fig. 3). In a previous work (Díaz-Molina et al., 1985), the Tórtola depositional system had been considered the only supply system filling the Loranca Basin from the Late Oligocene to the Early Miocene. New data indicate that another fan apex was located to the east of the Mariana-Cañamares syncline (fig. 2). This fluvial system has been called the Villalba de la Sierra fan. During the Late Oligocene to the Early Miocene (fig. 3) the Loranca Basin was connected with the Madrid Basin. In the Early Miocene, at the top of the Upper Unit, gypsum deposits extended into the basin indicating an endorreic situation.

In the Terminal Unit the facies distribution shows that in the Early Miocene (fig. 3) the supply systems flowed into an enclosed basin. The Terminal Unit has been divided into three subunits (Díaz-Molina et al., 1985). Subunit 1 corresponds to the sedimentation of the Valdeganga fan (fig. 3). Subunits 2 and 3 consist of alluvial and lacustrine sediments.

Figure 3
Simplified sedimentary log through the Loranca Basin. Paleontological data have been obtained by Daams et al. (1986) and Álvarez Sierra et al. (1987).



The Upper Unit

The Tórtola and Villalba de la Sierra depositional systems include the deposits of the fans as well as the deposits of their associated environments. These depositional systems overlie the previous sediments of the basin with onlap. Tectonic deformation also occurred during the deposition of this unit, as shown by progressive discordance on the flanks of some anticlinal folds and along the basin margins (fig. 4). The stratigraphic succession is flat lying in the centre of the basin, where only part of subunit 1 is exposed (fig. 2).

The greatest exposed thickness of the Upper Unit is 600 m. Its stratigraphic succession may be divided into three subunits (fig. 4), each with distinctive lithological characteristics which reflect the rate of diastrophism and changes in the base level of the systems.

Subunit 1 corresponds to the most active time interval of the fans. The stratigraphy contains mainly channelized bodies composed of conglomerates and sandstones, along with silts, silty clays, thin sandstones or siltstones sheets and limestone layers. To the south of the basin silty clays, limestones and gypsum are the dominant lithology.

Figure 4
Simplified sedimentary log through the Upper Unit.

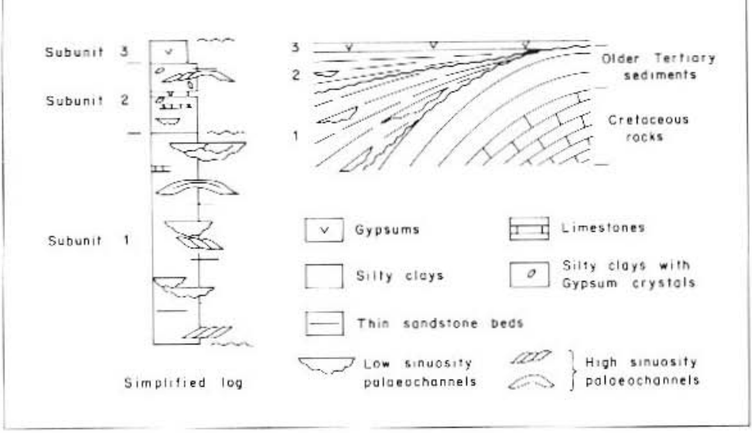
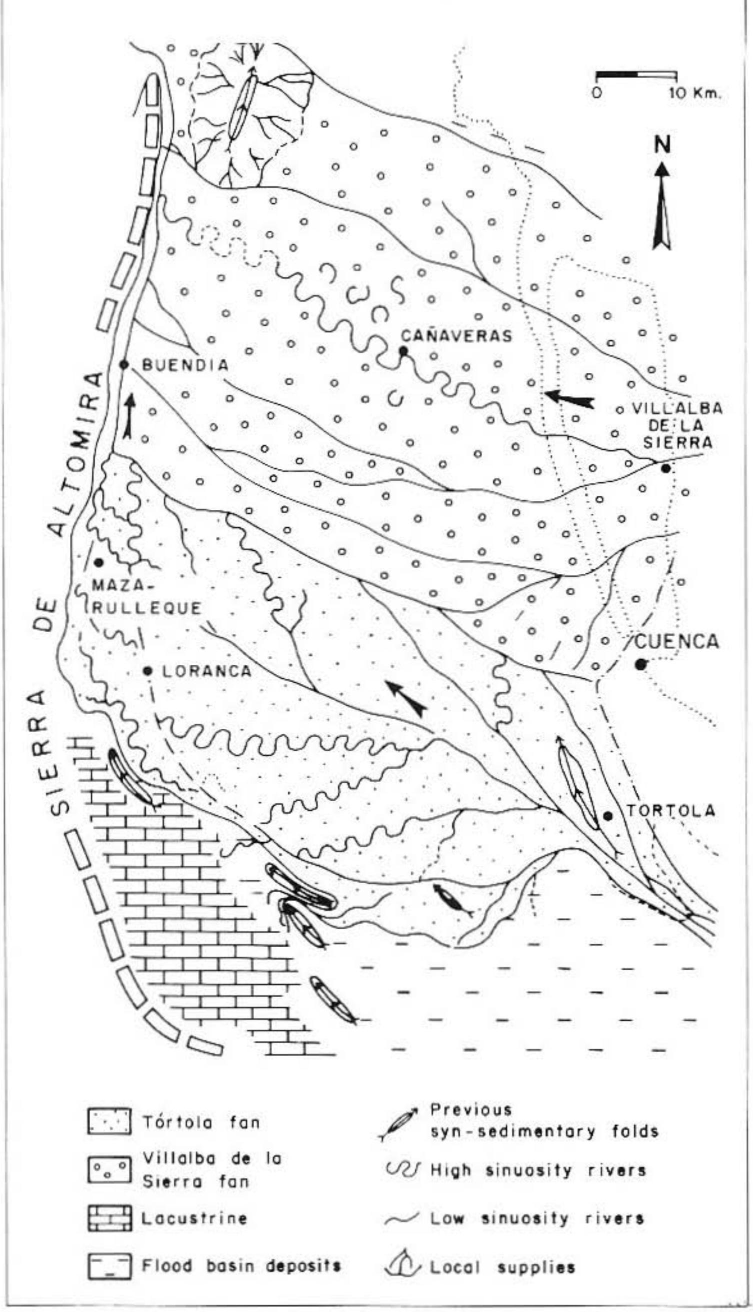


Figure 5
Reconstruction of the Tórtola and Villalba de la Sierra fans.



Subunit 2 differs from the previously described subunit in several aspects. The differences can be related to a decrease in tectonic activity, and subsequently, to a decrease in the rate of subsidence. The basin was filling up and local base levels were being established. In addition, during subunit 1 times, syn-sedimentary folding started to plug part of the area connecting the Loranca Basin and the Madrid Basin (fig. 5). Between subunits 1 and 2 there is a slight upwards decrease in palaeochannel dimensions and/or frequency. Gypsum and limestone layers interstratify with finer grained sediments containing gypsum crystals. Another characteristic of this unit is the abundance of sediments of local provenance fringing the basin, and indicating mass transport and ephemeral streams.

Subunit 3 is characterized by the dominance of gypsum deposits. The gypsum was clearly supplied from solution weathering of Late Cretaceous and Triassic rocks, which are very rich in gypsum deposits. Late Cretaceous gypsums are exposed along the Loranca Basin margins, whereas Triassic sediments appear in more internal areas of the Iberian Range. Triassic Keuper facies contain a type of authigenic quartz crystal (Jacinto de Compostela) which is found among the coarse grained sediments filling the channels.

The bulk of the distal facies of the fans are located in the Madrid Basin covered by younger Tertiary deposits. In the southwest margin of the Sierra de Altomira the stratigraphic equivalent is a thick succession of gypsum deposits.

The Upper Unit is a tecto-sedimentary unit, formed during the major compressional tectonic phase which affected the Iberian Range. This tecto-sedimentary unit reflects an upwards sedimentary polarity due to tectonic reactivation followed by a subsequent decrease in the diastrophism.

The depositional systems

A distinction between dry, relatively small fans formed by e-phemeral streams, and wet fans formed by perennial stream flow was proposed by Schumm (1977). Recent examples of the wet type are the Kosi River fan and the Riverine Plain (Shumm, o.c.). The Kosi River fan has built the largest fan described in the literature, covering 16,500 km^2 (Wells & Dorr, 1987). The Kosi River fan has been described by Gole & Chitale in 1966, and recently by Wells & Dorr. The Kosi River fan and the Tórtola and Villalba de la Sierra fluvial fans are similar in the dominance of individual channels which are part of a multiple channel system. In contrast to the Kosi River system the stratigraphic successions of the Tórtola and Villalba fans provide good exposures.

The catchment basins of the fans are not preserved at all, because tectonic deformation in Neogene time strongly modified the Late Oligocene to Early Miocene landscape.

The Tórtola fluvial fan is up to 94 km long, with a maximun width of 40 km, covering an area of 2,500 km^2 (fig. 5). Mapping reveals that it does not have the form of a cone segment. The Tórtola fan coalesces with the Villalba fan to the north, and its outline was tectonically controled to the south-west. In spite of its lacking a cone-like form it radiated downslope from the apex area, developing a multiple channel system. The apex of the Tórtola fan has been partially eroded, and it was also covered in part by the Valdeganga fan apex (Díaz-Molina et al., 1985). The Tórtola fan head filled a contemporaneous tectonic structure that was later folded into the synclinorium, trending south-east to north-west (fig. 5).

The Villalba de la Sierra fan covers an area of more than 4,200 km^2. This fan is up to 90 km long, with a maximum width of 60 km. The Villalba de la Sierra fan coalesces with the Tórtola fan to the south, and with another fan to the north. The apex of the Villalba de la Sierra fan was located within the Serranía de Cuenca. Later tectonic deformation and erosion have altered the original continuity of the fan sediments.

A paleoclimatic model was inferred on the basis of the Mammal succession recorded in the Loranca Basin (Lacomba, 1988). The procedure and method of inference is described in Weerd & Daams (1978); Daams & Meulen (1984). During the Late Oligocene to the Early Miocene, very important changes took place in the mammalian fauna including differences in diversity as well as in composition.

The mammalian faunal succession recorded in the Loranca Basin includes two important macromammal assemblages, found in Carrascosa and Loranca (see fig. 3). These two events reflect a relativeby arid landscape (Díaz-Molina et al., 1985; Lacomba, 1988; Lacomba & Morales, 1987). The faunas between both Carrascosa and Loranca events may indicate an episode of relatively more humid environment (Lacomba, 1988). A relatively cold climate was proposed for the arid events, and a warmer climate during the humid phases (Daams & Meulen, 1984).

Distributary and tributary areas

Distributary and tributary areas are distinguished using the analysis of palaeocurrents and variations in river type and depth. In figure 2 the palaeocurrent data from subunits 1 and 2 are shown. In most of the basin, palaeocurrent data were obtained selecting low sinuosity channels and measuring channel orientation where possible. In the centre of the basin, the data on figure 2 only represent the upper stratigraphic interval of the subunit 1. Only the predominant palaeochannel types present in each locality have been represented in figure 5.

Depending on the location in the fan area the predominant river type changes. In the Tórtola fan there is a general downstream change, from low (braided) to high sinuosity river types, in the areas with a distributary pattern, though in most of the sections low and high sinuosity channels are present. The gradual changes in channel types are also acompanied by a decrease in channel depth which is explained by channel division from the apex (fig. 5).

Other distributary areas have been recognized fringing the south-west lateral areas of the Tórtola fan where internal anticlines were present (figs. 2 and 5). These anticlines were topographic thresholds that occasionally were surmounted by rivers at their north-ward plugging ends. Because of this structural control, the Tórtola fan did not reach part of the south-west margin of the basin, where the multiple channel system was replaced laterally by flood basin, lacustrine or evaporitic sediments.

The Villalba de la Sierra fan carried coarser material than the Tórtola fan, and braided river type predominate. In the Cañaveras area (fig. 5) extensive meander loop deposits are present, but there is an upward change to braided channel deposits.

Towards the western margin of the basin, a tributary situation has been deduced, and at the Buendia locality to the north, low sinuosity channels dominate (Díaz-Molina et al., 1985). In contrast, thin point bar bodies are the most frequent type of river deposit, from Mazarullegue to Buendía. With time, a tributary pattern was also established, between Mazarullegue and Buendía, and low sinuosity palaeochannels can be followed between both localities, palaeochannels being parallel to the Sierra de Altomira.

The tributary zone to the west of the Loranca Basin is characterized by low sinuosity channels and a noticeable growth of channel density, which could have been produced by the lateral concentration of the channels of the fluvial systems. The channels of this area flowed to the north and passed to the Madrid Basin. On entry to the Madrid Basin they received new supplies from the north-east (Díaz-Molina et al., 1985).

Another major tributary area existed where the fans coalesced. The coalescence can be analysed along the Mayor River Valley, where channels belonging to both of these fans are present in the exposures.

An interpretation of the multiple fluvial systems is shown in figure 5. At times, runoff may have been perennial and limited to restricted areas of the fans; the surfaces of the fans were probably active during high flood events.

The western tributary zone seems to have been located along a topographic low, parallel to the structural strike of the Sierra de Altomira.

Later evolution of the fans

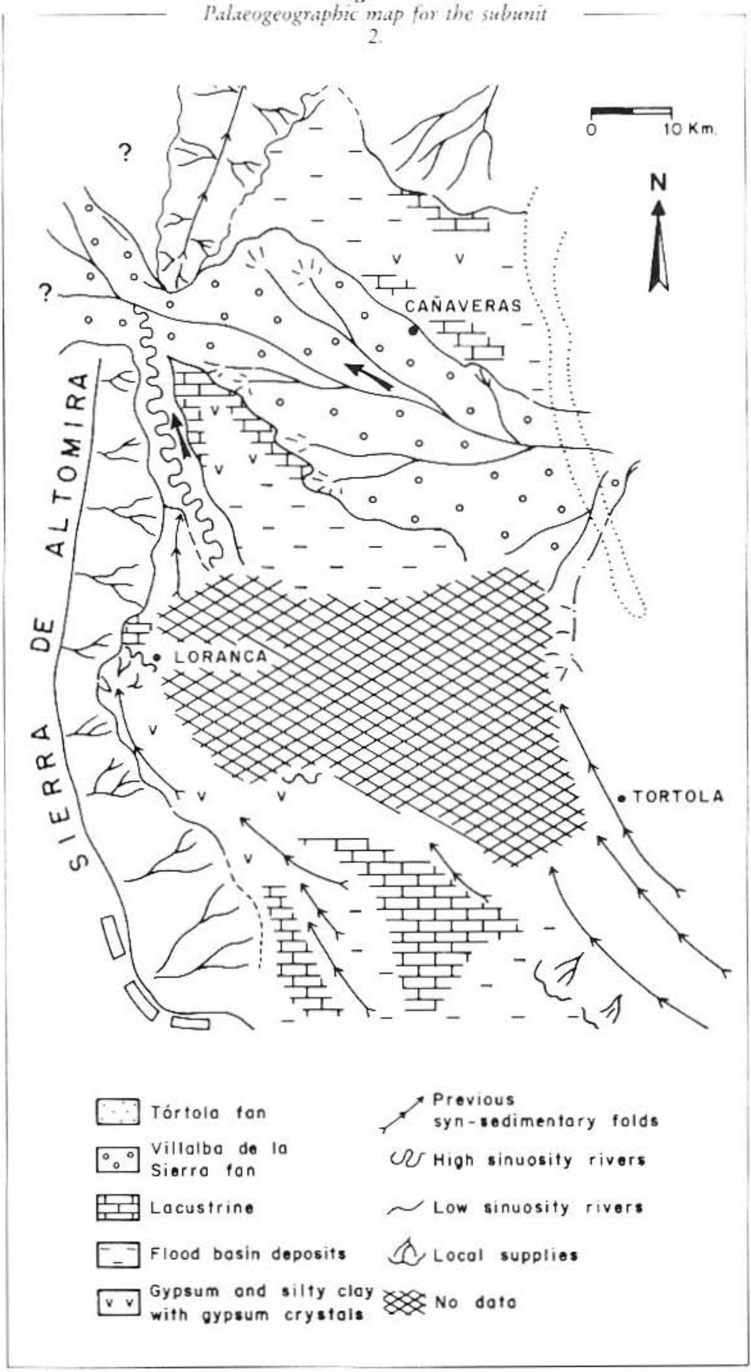
The fluvial network deduced from subunit 1 became restricted with time, its channels occupying a less extensive area than before. Exposures in the southern and northern aeras of the basin indicate that the fluvial systems did not flow over the whole basin. Near the Buendía locality a tributary pattern was maintained and rivers probably flowed to the west into the Madrid Basin (fig. 6).

The Tórtola system was dominated by meandering rivers in the exposures (fig. 6), though low sinuosity palaeochannels are also found. These palaeochannels dissected silty clay deposits with gypsum crystals which indicate proximity to salt pan environments. From the Huete locality to the north a white colour characterizes the palaeochannel infills, and is due to gypsum cement. These palaeochannels also carried detrital gypsum and other clast types of intrabasinal origin. The gypsum cement probably precipitated during an early burial stage.

In contrast, the Villalba de la Sierra fan was dominated by low sinuosity palaeochannels (fig. 6). Palaeochannels with gypsum cement are relatively scarce. The more active channels flowed into the Madrid Basin, but channels of smaller dimensions died out in internal wet areas leaving silty and sandy sheets. In the internal wet areas gypsum and carbonate deposits precipitated.

Eventually, the downstream ends of both systems retreated and were replaced by the deposits of wet-land environments. The exact pattern of these changes varied in the different areas. Powder gypsum is present as detrital deposits or continuous compact bioturbated layers. Coarsely crystalline gypsum facies are also present as continuous layers. Carbonate sedimentation occurred in association with the gypsum facies: homogeneous micritic limestones with gypsum crystals, algal laminated micritic limestones with tepees and mud cracks structures. Carbonaceous marls are present locally, associated with the carbonate sediments. All these facies reveal a more permanent high water table with the development of ephemeral swamps and salt pan environments with sporadic anoxic stages. In addition, the gypsum and carbonate facies are often interbedded with thick mudstone and silstone layers with

Figure 6
Palaeogeographic map for the subunit 2.



gypsum crystals. These represent crevasse-splay and distal deposits of the secondary channels of the fluvial systems.

The Tórtola fan became abandoned before the Villalba de la Sierra fan. In its place a permanent wet area developed of facies characteristics similar to the contemporaneous lacustrine deposits that existed between the two fluvial systems in the centre of the basin.

Subunit 3 covered the last detrital deposits and the Loranca Basin became endorreic. The transition between subunits 2 and 3 is gradual as the fluvial systems died. During subunit 3 deposition a supply of terrigenous sediment continued, though it was restricted to the fan apex areas.

Palaeochannel type recognition

Braided channels

Braiding is identified in these outcrops on the basis of three criteria: (1) the presence of minor imbricate channels, (2) channel incision on sandy bars and (3) the preservation of bed forms that are characteristic of the building of alluvial islands (Díaz-Molina et al., 1985).

The presence of minor imbricate channels is the result of aggradation in a composite-stream river (Allen, 1985; William & Rust, 1969). Inside the minor channel, composite bars may also be found also.

Sandy bars are the most conspicuous bed form preserved in the distributary area adjacent to the Tórtola fan apex. Their internal structure is foreset cross-stratification. Sandy bar thicknesses oscillate between 0.25 m and 4 m, and their climbing and descending components (Friend, 1982) have usually usually been preserved. The climbing component is formed by sedimentary structures ranging from ripples to minor bars. Reactivation surfaces formed by superimposed bed forms (Mc Cabe & Jones, 1977) may be found in the descending component. Sandy bars are not limited to the braided model, though they have been described frequently in braided rivers (Collinson, 1970; Smith, 1971; Cant 1978; Cant & Walker, 1978). However the characteristic feature of braided rivers is the dissection of bars by smaller channel during the falling and low stages (Ore, 1964; Smith, 1971; Cant & Walker, 1978; Blodgett & Stanley, 1981). When the bars are dissected, parts of them become emergent and the rivers show a braided pattern. In some of these ancient examples this process can be inferred.

Composite bed forms have also been preserved, and are of a larger scale than the sand bars. Preservation includes not only sequences but also the topographic build up of the islands. Among these composite bars sand flats (Cant & Walker, 1978) are relatively frequent. Composite bars have been found isolated, and this implies the avulsion of the braided channels.

Meandering channel deposits

High sinuosity rivers left point bar and meander loop complexes (Díaz-Molina, 1978, 1979 a; 1984; Díaz-Molina et al., 1985). The general characteristics of the point bar bodies are those of the classic type which were summarized by Moody-Stuart in 1966.

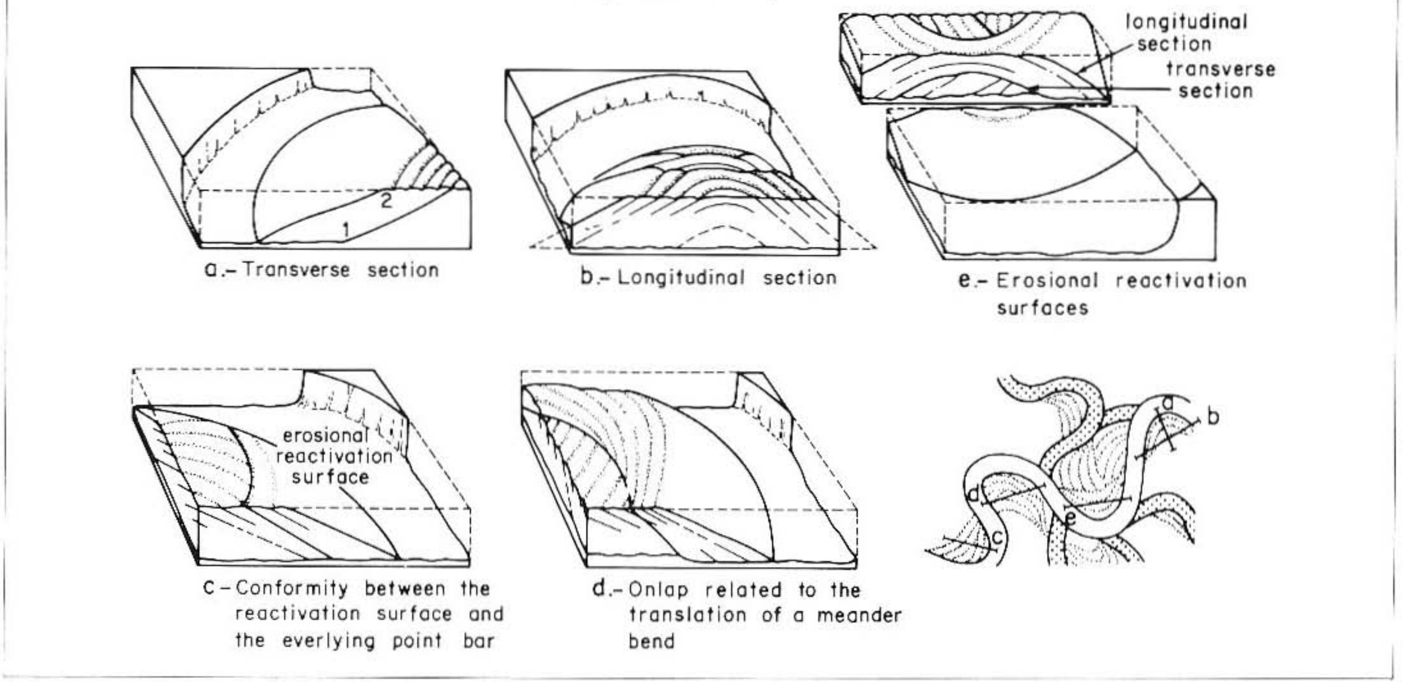
In the point bar bodies two sub-units can be identified. The lower sub-unit consists of gravel size to medium grained sand and the upper one of fine sand to silt. Lateral accretion surfaces are always present in the upper sub-unit and are less frequent in the lower sub-unit. In contrast sedimentary structures are well preserved in the lower sub-unit but not always in the upper one, where some pedogenic processes can be distinguished (Díaz-Molina et al., 1985). Within individual lateral accretion episodes a fining upward sequence usually developed, and a log through the whole point bar body also tends to show fining upwards. However, some point bars show departures from the fining upward pattern which reflect different flow conditions from one episode to the next.

In most of the examples, point bar sedimentary structures were made by bed forms that moved up the point bar, as could be expected if a curvature-related helicoidal flow were active. In one example (stop 5) where sedimentary structures indicate parallel-to-the-bend and down-the-point-bar movement, the estimated sinusity is 1.6, while in the other example, in which the helicoidal flow pattern was well developed the estimated sinuosity is 2 (stop 7). No more estimations of palaeochannel morphology have been possible because it is difficult to find sections clearly coincident with a meander loop radius, and point bar dimensions vary significantly in oblique or longitudinal sections to the meander bend.

Point bar geometries in section transverse or longitudinal to the meander loop have been discussed in previous works (Díaz-Molina, 1978, 1984; Díaz-Molina et al., 1985). The different geometries seen in different types of section are shown in figure 7.

Meander loop deposits in the Loranca Basin frequently display laterally stacked sequences deposited by adjacent point bars, separated by reactivation surfaces. Some of the relations between these surfaces, bed set geometry and changes in the sense of

Figure 7
Point bar body geometries and meander loop reactivation surfaces.



migration of the adjacent point bar bodies have been represented in figure 7.

Low sinuosity palaeochannels with vertical accretion

When the orientation of a short stretch of these palaeochannels can be observed, they are straight or slightly sinuous. However they cannot be considered straight channels in the Leopold and Wolman (1957) sense of the term. They lack either a sequence of sedimentary structures or any geometric characteristics in which side bars of riffle and pool bed-from patterns can be recognized. These examples differ from braided river deposits in lacking minor channel incisions on the river bed. Some of them can be compared with the ribbon type geometry defined by Friend et al. (1979), though lateral wings are not present. Like ribbon type channels they can be simple or multi-storey, and where there are several superimposed units, they are more or less tabular bodies separated by flat surfaces. These aggradational units are considered to be backfilling units (Díaz-Molina, 1979 b), and simple ribbon examples represent a process of continuous backfilling.

Water escape structures

Water escape structures normally appear in fine to medium grained sand, though occasionally they have also affected coarse grained sand (Díaz-Molina, 1978). They are of the three kinds distinguished by Lowe (1975): hydroplastic deformation, liquefaction and fluidization. Frequently hydroplastic deformation and liquefaction are associated. Some palaeochannels show water escape structures affecting the whole channel fill. In spite of this difficulty, a distinction between low or high sinuosity channels is possible in many cases using body geometry alone.

Sandstone petrology

The petrology of sandstones from the Loranca Basin has been described by García Palacios (1974), who analysed sandstone textures and the mineralogy of detrital framework grains and interstitial material. In the present study, samples of channel sandstones from stratigraphic subunits 1 and 2 have been examined. Further work to analyse the sandstone composition of the different fans is still in progress. The petrology of each subunit is summarised below.

Subunit 1

Coarse to medium grained sandstones occur at the base of channels and these fine upwards into fine grained deposits. They are well sorted with angular to subrounded detrital quartz grains (Powers, 1953).

The main detrital components are extrabasinal non-carbonate grains (NCE of Zuffa, 1980) including monocrystalline quartz featuring minor quartz overgrowths, K-feldspar and microcline. Quartz with anhydrite inclusions is also present. A small number of extrabasinal carbonate grains occur (e.g., micritic and sparry limestones and dolostones).

The mean modal composition of extrabasinal detrital grains is Q_{60} F_{10} RF_{30} (fig. 8), reflecting a high proportion of recycled sedimentary rocks derived from the Mesozoic succession of the Iberian Range.

Variable amounts of intrabasinal carbonate grains (e.g. micritic limestone grains) are also present (fig. 8c). These limeclasts are derived from pedogenetic carbonates and/or palustrine-lacustrine limestones.

The mean modal sandstone composition is similar throughout the unit and it is therefore possible to infer that the lithology of the source area remained constant during deposition of the sequence.

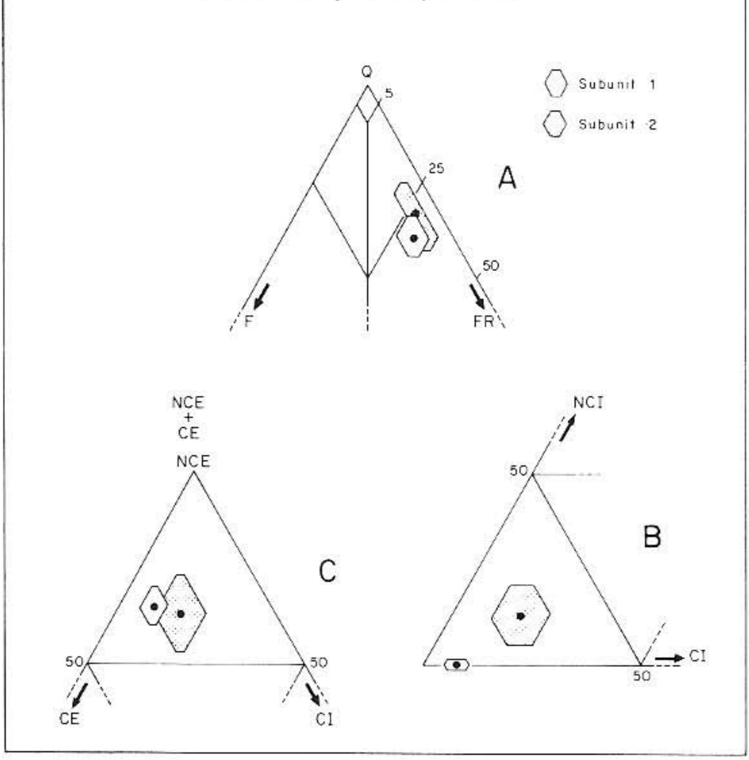
Sandstone diagenesis is limited to minor compaction (e.g., deformed labile grains) and patchy carbonate cementation. The deposits have retained a high primary porosity which is locally reduced by calcite cement. This forms a porosity occluding mosaic and/or rims around monocrystalline carbonate grains.

Subunit 2

Channel sandstone grainsize and sorting is similar to that described in subunit 1. However fine grained channel-fill sandstones occur more frequently in subunit 2.

The modal composition of the extrabasinal components is similar to that of the underlying deposits indicating erosion from a hinterland of similar geology.

Figure 8
Framework composition of sandstones.



However, differences are noted in the composition of intraclasts within the sandstones of subunit 2. Firstly, a higher proportion of micritic limestone intraclasts are present thus indicating expansion of calcareous palaeosol and/or palustrine-lacustrine depositional environments during sedimentation of the channel sandstones. Secondly, detrital gypsum grains are preserved in the sandstones (NCI in fig. b). In the absence of petrographic evidence the origin of the detrital gypsum remains problematic. The grains may be intrabasinal or extrabasinal. However, the increase in limestone intraclasts which accompanies the incoming of gypsum would suggest that grains were eroded from contemporaneous gypsum deposits (e.g., crusts and primary gypsum layers) which precipitated within the basin.

The higher proportion of intrabasinal clasts within subunit 2 indicates a decrease in tectonic activity and elevation of the base level of the fluvial system. This resulted in a decrease in the erosion and supply of extrabasinal detritus to the Loranca Basin, combined with more active fluvial reworking of the floodplain of the fans.

The diagenesis of subunit 2 sandstones differs markedly from that previously described for subunit 1. Two types of cement exist: 1. *Dolomite*; this coats all detrinal grains with a thin envelope of small crystals. Some gravitational or meninsus cements are also preserved. 2. *Poikilotopic gypsum*; this cement forms large crystals (up to 15 cm long). In areas of high abundance of gypsum grains, syntaxial overgrowths have developed around these grains. Both gypsum and dolomite cements corrode and replace detrital framework grains.

Cementation occurred during and soon after the deposition (eodiagenetic) of subunit 2 sandstones from highly saline vadose and phreatic pore waters. The increase in groundwater salinity during deposition of subunit 2 correlates with the increase in base level of the fluvial system and lake level highstand.

Primary porosities within subunit 2 sandstones are low due to extensive dolomite and gypsum cementation. However, in some channel sandstones, telodiagenetic dissolution of gypsum has taken place resulting in high secondary porosity values. Dissolution has mainly affected detrital gypsum grains, primary gypsum cements and replasive gypsum cements after carbonate grains.

Acknowledgements

Financial support for this work has been provided by the C.I.C. Y.T., research grant no. PB85-0022. We thank Dr. Peter Friend for his critical comments on the draft manuscript, and Jacqueline Bond and Donna Balin for their help with the preparation of the text. Fernando Pérez drafted most of the figures.

References

ALLEN, J.R.L. (1965).
A review of the origin and characteristics of recent alluvial sediments. SEDIMENTOLOGY 5: 89-191.

ALLEN, J.R.L. (1982).
SEDIMENTARY STRUCTURES. THEIR CHARACTER AND PHUSICAL BASIS. VII, Elsevier, 663 pp.

ÁLVAREZ-SIERRA, M.A.; DAAMS, R.; LACOMBA, J.I.; LÓPEZ-MARTÍNEZ, N. & SACRISTÁN-MARTÍN, M.A. (1987).
Succession of micromammal faunas in the Oligocene of Spain. MÜNCHNER GEOWISS. ABH. 10: 43-48.

ALVARO, M.; CAPOTE, R. & VEGAS, R. (1979).
Un modelo de evolución geotectónica para la Cadena Celtibérica. ACT. GEOL. HISPÁNICA 14: 172-177.

BLODGETT, R.H. & STANLEY, K.O. (1980).
Stratification, bedforms and discharge relations of the Platte braided River system, Nebraska. J. SEDIMENT. PETROL. 50: 139-148.

BREWER, R. (1964).
FABRIC AND MINERAL ANALYSIS OF SOILS. John Wiley & Sons Inc., London, 470 pp.

BRICE, J.C. (1974).
Evolution of meander loops. GEOLOG. SOC. AM. BULL. 85: 581-586.

CANT, D.J. (1978).
Bedforms and bar types in the South Saskatchewan River. J. SEDIMENT. PETROL. 48: 1321-1330.

CANT, D.J. & WALKER, R.G. (1978).
Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. SEDIMENTOLOGY 25: 625-648.

CAPOTE, R. (1983).
La tectónica de la Cordillera Ibérica, in I.G.M.E., ed., LIBRO JUBILAR J.M. RÍOS – GEOLOGÍA DE ESPAÑA, *t.II: 109-120.*

COLLINSON, J.D. (1970).
Bedforms of the Tana River, Norway. GEOGRAF. ANN. 52-A: 31-56.

DAAMS, R.; LACOMBA, J. & LÓPEZ-MARTÍNEZ, N. (1986).
Nuevas faunas de micromamíferos del Terciario continental de la Depresión Intermedia (Provincia de Cuenca, España Centro-Oriental). ESTUDIOS GEOL. 42: 181-196.

DAAMS, R. & MEULEN, A.J. van der (1984).
Paleoenvironmental and paleoclimatic interpretation in the Upper Oligocene and Miocene of north central Spain. PALÉOBIOLOGIE CONTINENTALE, *Montpellier, 14: 241-257.*

DÍAZ-MOLINA, M. (1974).
Síntesis estratigráfica preliminar de la serie terciaria de los alrededores de Carrascosa del Campo (Cuenca). ESTUDIOS GEOL. 30: 63-67.

DÍAZ-MOLINA, M. (1978).
BIOESTRATIGRAFÍA Y PALEOGEOGRAFÍA DEL TERCIARIO AL ESTE DE LA SIERRA DE ALTOMIRA. *Tesis doctoral. Facultad de Ciencias Geológicas. Universidad Complutense de Madrid. 370 pp.*

DÍAZ-MOLINA, M. (1979a).
Características sedimentológicas de los paleocanales de la Unidad Detrítica Superior al N. de Huete (Cuenca). ESTUDIOS GEOL. 35: 241-251.

DÍAZ-MOLINA, M. (1979b).
Descripción del frente distal de un abanico fluvial húmedo en el Terciario continental situado al Este de la Sierra de Altomira. ESTUDIOS GEOL. 35: 119-129.

DÍAZ-MOLINA, M. (1983).
Un ejemplo del sistema fluvial «braided» con preservación de «sand flats». Unidad Detrítica Superior. Terciario de la provincia de Cuenca. ESTUDIOS GEOL. 39: 345-357.

DÍAZ-MOLINA, M. (1984).
Geometry of sandy point bar deposits, examples of the Lower Miocene, Tajo Basin, Spain. ABSTRACTS I.A.S. 5TH EUROPEAN REGIONAL MEETING, MARSEILLE, FRANCE. 140-141.

DÍAZ-MOLINA, M.; BUSTILLO-REVUELTA, A.; CAPOTE, R. & LÓPEZ-MARTÍNEZ, N. (1985).
Wet fluvial fans of the Loranca Basin (central Spain). Channel models and distal bioturbated gypsum with chert. EXC. GUIDE-BOOK I.A.S. 6TH EUROPEAN REGIONAL MEETING, LÉRIDA, Spain. 149-185.

DÍAZ-MOLINA,M & LÓPEZ MARTÍNEZ, N. (1979).
El Terciario continental de la Depresión Intermedia (Cuenca). Bioestratigrafía y Paleogeografía. ESTUDIOS GEOL. 35: 149-167.

ETHRIDGE, F.G. & SCHUMM, S.A. (1978).
Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations and assessment, in Miall, A.D. ed., FLUVIAL SEDIMENTOLOGY. CAN. SOC. PETROL. GEOL. MEM. 5: 703-722.

FREYTET P. & PLAZIAT, J.C. (1982).
Continental carbonate sedimentation and pedogenesis: Late Cretaceous and Early Tertiary of Southern France. CONT. SEDIMENTOLOGY 12: 213 pp.

FRIEND, P. (1982).
Towards the field classification of alluvial architecture or sequence. SPEC. PUBL. INT. ASS. SEDIMENT. 6: 345-354.

FRIEND, P.F.; SLATER, M.J. & WILLIAMS, A.C. (1979).
Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain. J. GEOL. SOC. LOND. 136: 39-46.

GARCIA PALACIOS, C. (1974).
Estudio petrogenético del Terciario entre la Sierra de Altomira y la Sierra de Bascuñana (C. Ibérica). ESTUDIOS GEOL. 30: 109-144.

GARCIA-ABBAD, F. (1975).
ESTUDIO GEOLOGICO DE LA REGION DEL PANTANO DE ALARCÓN (CUENCA). *Tesis Doctoral. Facultad de Ciencias Geológicas. Universidad Complutense de Madrid, 175 pp.*

GOLE, C.V. & CHITALE, S.V. (1966).
Inland delta building activity of Kosi River. AM. SOC. CIVI. ENG. J. HYDRAUL. DIV. HY-2, 111-126.

HICKIN, E.J. (1974).
Development of meanders in natural river channels. AM. J. SCI. 274: 414-442.

JONES, C.M. (1977).
Effects of varying discharge regimes on bedform sedimentary structures in modern rivers. GEOLOGY 5: 567-570.

KLAPPA, C.F. (1980).
Rhizolits in terrestrial carbonates classification, recognition, genesis and significance. SEDIMENTOLOGY 27: 613-631.

LACOMBA, J.I. (1988).
ESTUDIO DE LAS FAUNAS DE MICROMAMIFEROS DEL OLIGOCENO SUPERIOR Y MIOCENO INFERIOR EN LAS CUENCAS DE LORANCA, EBRO RIOJANO Y EBRO ARAGONES. ASPECTOS PALEOECOLOGICOS. Tesis Doctoral. Universidad Complutense de Madrid, 389 pp.

LACOMBA, J.I. & MORALES, J. (1987).
Los mamíferos del Oligoceno superior de Carrascosa del Campo (prov. de Cuenca, España). MUNCHNER GEOWISS. ABH. 10: 389-300.

LEOPOLD, L.B. & WOLMAN, M.G. (1957).
River channel patterns: straight, meandering and braided. U.S., GEOL. SUV., PROFESS. PAPERS 282-B: 39-85.

LEVEY, R.A. (1978).
Bedforms distributions and internal stratification of coarse-grained point bars, upper Congaree River, s.c., in Miall, A.D., ed., FLUVIAL SEDIMENTOLOGY. CAN. SOC. PETROL. GEOL., MEM. 5: 105-128.

LOWE, D. (1975).
Water escape structures in coarse grained sediments. SEDIMENTOLOGY 22: 157-204.

McCABE, P.J. & JONES, C.M. (1977).
Formation of reactivation surfaces within superimposed deltas and bed forms. J. SEDIMENT. PETROL. 47: 707-715.

MEIN, P. (1975).
Biozonation du Neogène Mediterranéen à partir des Mammifères. IUGS RCMNS. REPORT ON ACTIVITY OF THE RCMNS WORKING GROUPS BRATISLAVA: 78-81.

MELENDEZ-HEVIA, F. (1971).
Estudio geológico de la Serranía de Cuenca en relación a sus posibilidades petrolíferas. PUBL. DE LA FAC. DE CIENC. DE LA UNIVERSIDAD COMPLUTENSE DE MADRID, SERIE A, 153, 245 pp. and 154, 80 fig. and 15 pl.

MOODY-STUART, M (1966).
High- and low-sinuosity deposits, with examples from the Devoniana of Spitsbergen. J. SEDIMENT. PETROL. 36: 1102-1117.

ORE, T.H. (1964).
Some criteria for recognition of braided stream deposits. WYOMING CONTR. GEOL. 3: 1-14.

PETTIJOHN, F.J.; POTTER, P.E. & SIEVER, R. (1973).
SAND AND SANDSTONES. Springer-Verlag, New York-Heidelberg-Berlin, 618 pp.

POWERS, M.C. (1953).
A new roundness scale for sedimentary particles. JOUR. SED. PETROLOGY 23: 117-119.

SCHUMM, S.A. (1977).
THE FLUVIAL SYSTEM. John Wiley & Sons. New York, London, 338 pp.

SMITH, N.S. (1971).
Transverse bars and braiding in the Lower Platte River, Nebraska. GEOL. SOC. AM. BULL. 82: 3407-3420.

STOOPS, G. & ILAWI, M. (1981).
Gypsum in arid soils. Morphology and genesis. PROCEEDINGS THIRD INTERNATIONAL SOIL CLASSIFICATION WORKSHOP, ACSAD: 175-185.

VILAS-MINONDO, L. & PÉREZ-GONZÁLEZ, A. (1971).
Contribución al conocimiento de las series continentales de la mesa manchega (Cuenca). BOL. R. SOC. ESPAÑOLA HIST. NAT. (GEOL.) 69: 103-114.

VIALLARD, P. (1973).
RECHERCHES SUR LE CYCLE ALPIN DANS LA CHAINE IBÉRIQUE SUDOCCIDENTALE. Thèse Université Paul Sabatier, Toulouse, 445 pp.

WEERD, A. van der & DAAMS, R. (1978).
Quantitative composition of rodent faunas in the Spanish Neogene and Paleoecological implications. PROCEED. KONINKL. NEDERL. AKAD. WETENS. SER. B, 81(4): 448-473.

WELLS, N.A. & DORR, J.A. Jr. (1987).
A reconnaissance of sedimentation on the Kosi alluvial fan of India. S.E.P.M., SPEC. PUBL., 39: 51-61.

WILLIAMS, P.F. & RUST, B.R. (1969).
The sedimentology of a braided river. J. SEDIMENT. PETROL. 39: 649-679.

ZUFFA, G.G. (1980).
Hybrid arenites: their composition and classification. JOUR. SED. PETROLOGY 50: 2129-2137.

FIELD TRIP

Figure 9
Itinerary and location of stops.

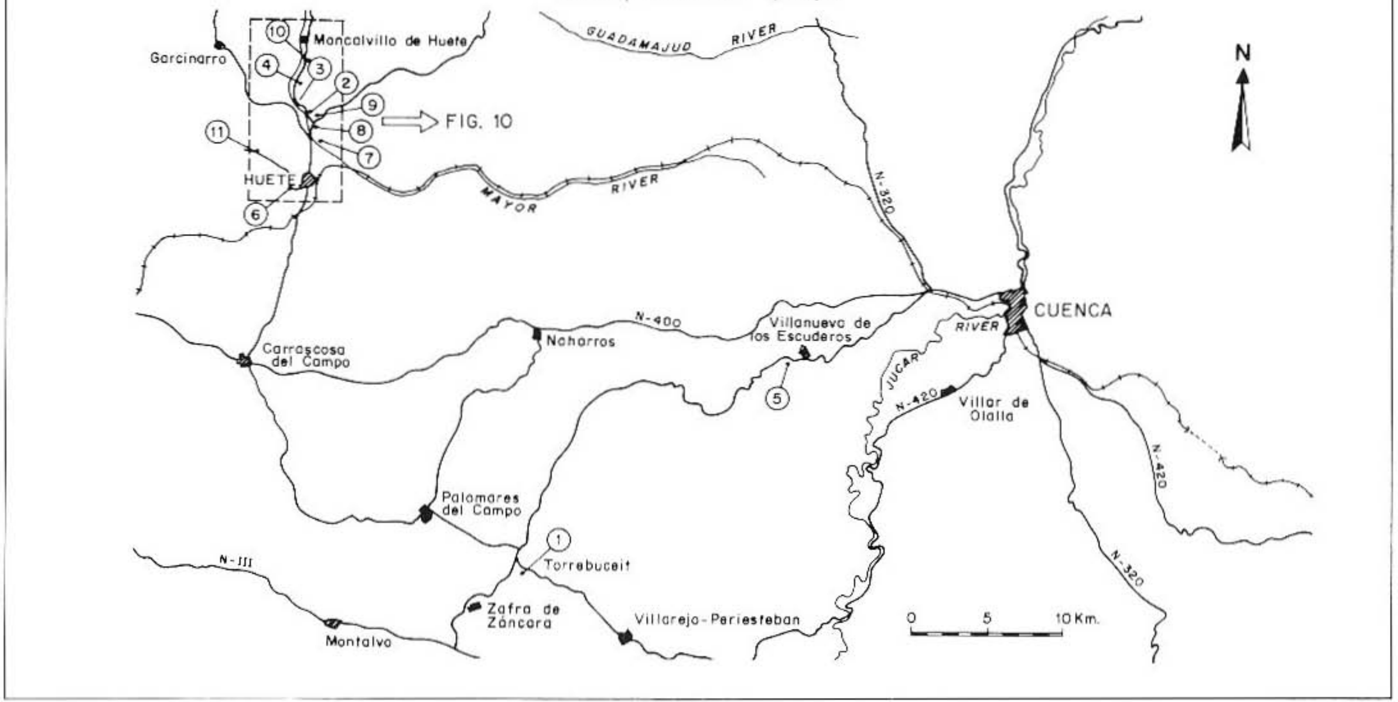
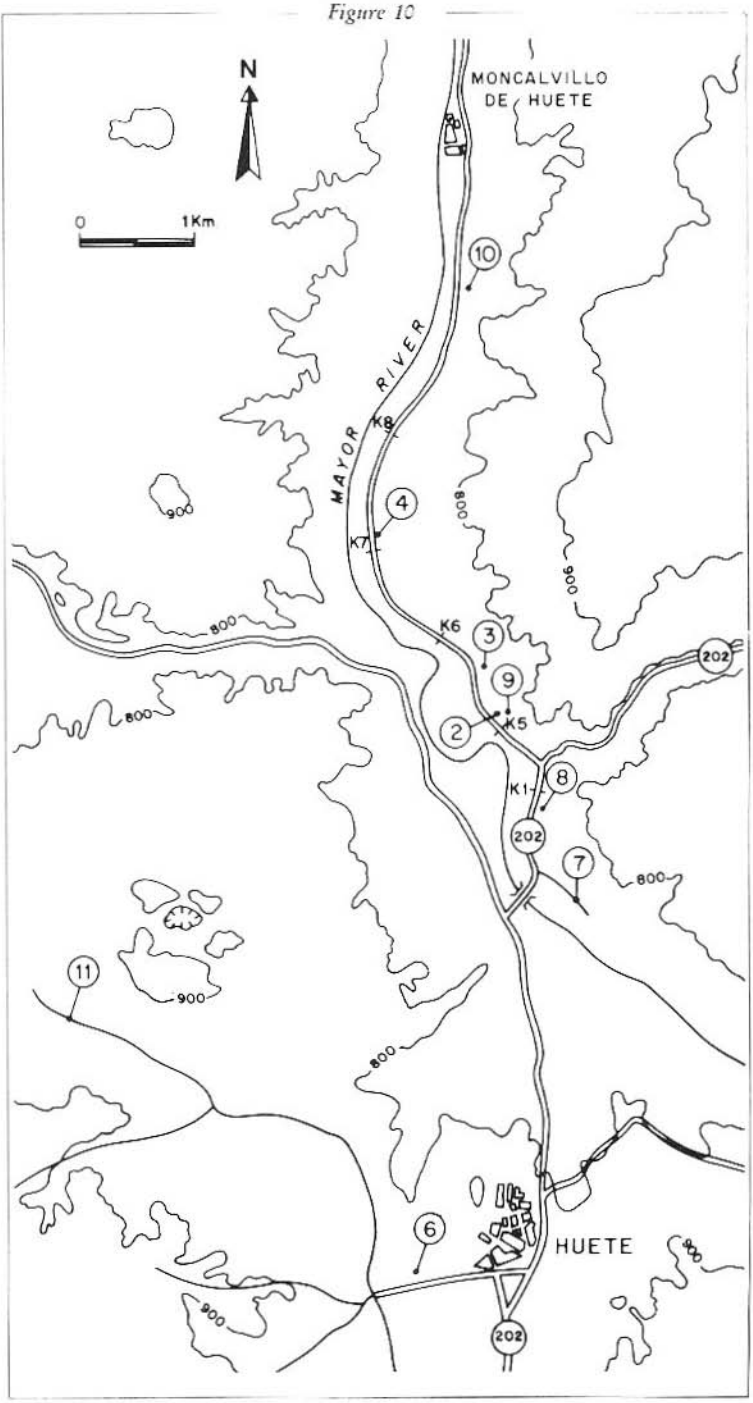
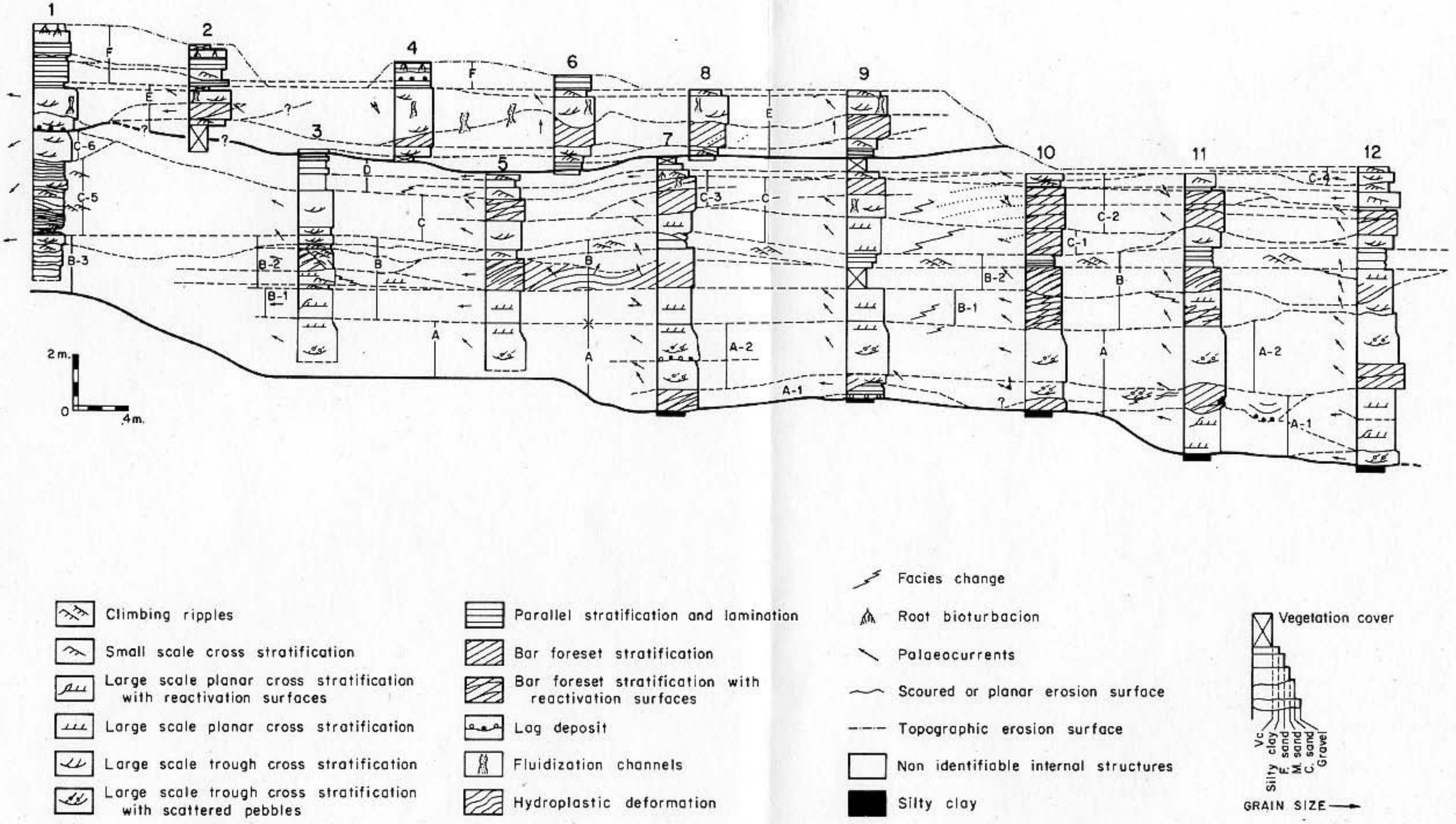


Figure 10



N
0
1Km
MONCALVILLO
DE HUETE
10
MAYOR RIVER
K8
4
K7
800
900
900
800
K6
3
9
2
K5
800
202
8
K1
202
7
800
11
900
800
6
HUETE
900
800
202

Figure 11
Braided palaeochannels. Preservation of sand flat sequences and geometries. Channel incision on the bars.



STOP 1

Example of braided channel with sand flat preservation

Km 43.8 between Torrebuceit and Villarejo-Periesteban (fig. 9). Parking site by the road. Exposure 50 m to the south.

The outcrop has a lateral extent of 170 m and is formed of two superimposed palaeochannel fills (fig. 11). Both palaeochannels were filled by braided rivers. At the top of the lower channel fill an abandonment of part of its active bed can be recognized; in contrast, the upper channel sequence of structures indicates a progressive incorporation to the active bed of the system. This deduction implies the connection of the upper channel with at least one active channel located laterally in a lower topographic level.

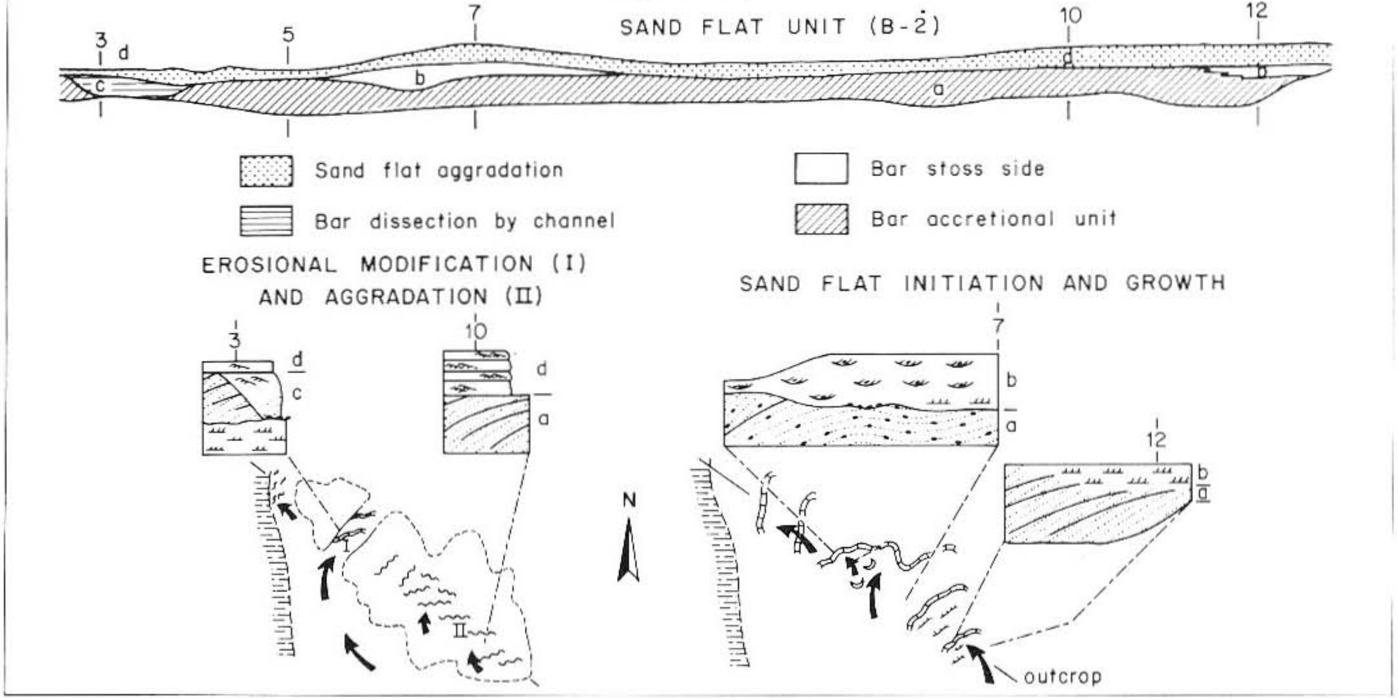
The interpretation of braided river action (Díaz-Molina, 1983) has been based on: the presence of smaller imbricate channels (unit A, fig. 11), the preservation of sand flat sequences and geometries (units B-2, C-2 and C-3, fig. 11) and lastly, minor channel incision on the bars (units B-2, C-2 and E).

The sand flat sequences begin with foreset cross stratification which is the internal structure of transverse or linguoid bars. The foreset cross stratification overlies scoured surfaces (figs. 11 and 12), and this suggests that the bars have formed in areas of flow expansion (Cant, 1978).

Within the bar foreset cross stratification, there are multiple convex upwards erosion surfaces, which have clearly been formed by superimposed bed forms (Mc Cabe & Jones, 1977; Levey, 1978). This type of reactivation surface and the channel incision on bars are considered to be modifications formed at moderate to low rates of falling stage. In this exposure, reactivation surfaces cannot be considered a modification since they appear from the bar initiation and throughout its development. Occasionally reactivation surfaces are present for only over a short interval, perhaps indicating stage changes during bar migration as Cant & Walker (1978) have suggested.

The horizontal upper surface of the foreset cross stratification

Figure 12
Sand flat development.



unit in B-2 and C-2 (fig. 11) has involved a levelling by erosion. In the Platte River this kind of erosion on bars originated by shallow and high velocity water during the falling of the stage (Blodgett & Stanley, 1980).

In one of the examples of bar accretion, unit B-2, changes in the crest line have been deduced (figs. 11 and 12). These changes could have happened during continuous falling stage, when bars develop irregular or asymmetrical patterns (Cant & Walker, 1978). The migration change to the river margin deduced in unit B-2, could also have originated as a result of decrease in channel depth, as has been described in the Platte River (Smith, 1971).

Aggradation on the bars was associated with ripple cross-bedding (B-2, figs. 11 and 12). The aggradation ocurred during higher stages, as in the Saskatchwan River (Cant & Walker, 1978).

The bars and sand flat development was contemporaneous with the channel infill, since downcurrent a simultaneous aggradation was produced by dune bed form (units B-1 and C, fig. 11).

STOP 2

Braided channel. Preservation of the convex up morphology of composite bars. Upwards evolution of the channel filling architecture.

Km 5.1 between Huete and Moncalvillo de Huete (figs. 9 and 10). Parking site on a track to the left of the road.

The channel fill has been divided into 6 units, corresponding to successive aggradational stages (fig. 13). The body is 4.30 m thick and has a lateral extent of 57 m.

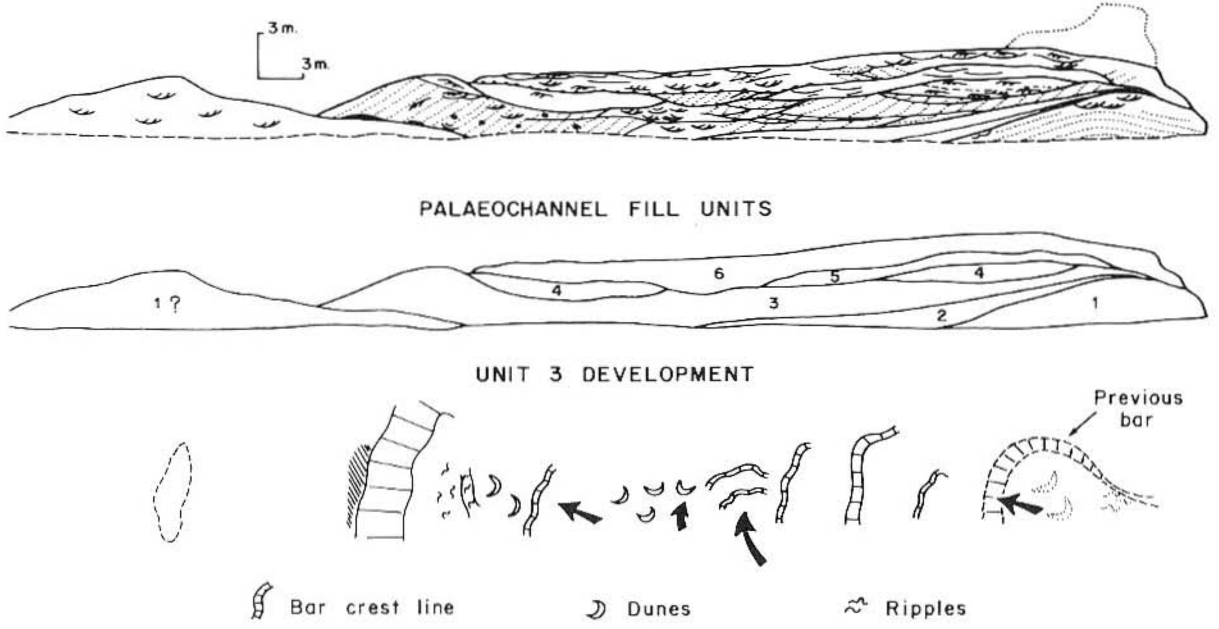
Unit 1 consists of a composite bar interpreted to have had a liguoid form. The bar body external geometry is convex in a section perpendicular to the flow, and its internal structure is formed by convex laminae overlain by large-scale cross-stratification at the top. The cross stratification was probably related to dune migration on the stoss side of the bar. The bed form surface has been only lightly eroded, and its preservation may be a consequence of a rapid fall of discharge, though very large bed forms may not adjust even with relatively slow changes in discharge (Jones, 1977). Discharge decrease was probably accompanied by the exposure of the macro-bed form as a braid bar in the river bed.

In unit 2 the subsequent discharge increase could have produced lateral accretion units on the bar.

Unit 3 is mainly formed by foreset cross stratification which is the internal structure of bars. Palaeocurrents indicatethat the flow was transverse to the topographic high formed by unit 1, though some lightly scoured surfaces limiting the internal structures o the bars may have originated by a flow perpendicular to the outcrop (flowing to the east, see fig. 13). Laterally a large composite bar developed, growing to the margin of a channel. On the stoss side of the bar, ripple and dune bed-forms migrated and the bar overlay silty deposits which are a relic of a previous lower stage.

Units 4, 5 and 6 consist of imbricate minor channel fills. Though the braided pattern persisted, each successive stream chan-

Figure 13
Cross-section through braided channel deposits. Preservation of composite bars.



nel was formed by channels of smaller dimensions (Díaz-Molina, 1979 a). With the progressive aggradation and abandonment, the channel bed was probably excessively wide in relation to the available discharge, and flow was confined to small channels where ripple bed-forms predominated.

STOP 3

Stacked low sinuosity palaeochannels. Water escape structures.

Km 5.5 between Huete and Moncalvillo de Huete (fig. 10). Parking site on a track to the right.

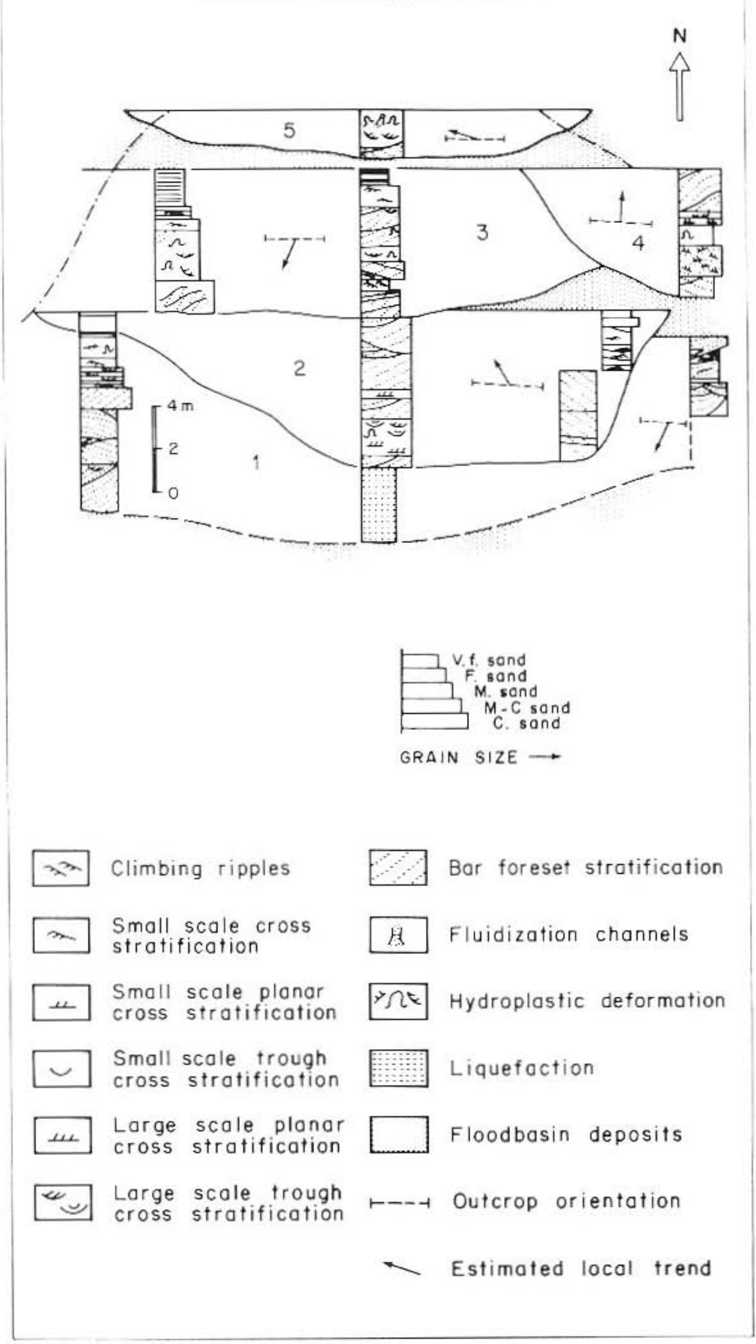
This outcrop has a lateral extent of 200 m and five stacked palaeochannels can be distinguished (fig. 14). Palaeochannel local trends are to the north, north-west and south-west. Each of these channel examples appears to have been of low sinuosity.

In most of these palaeochannel infills, foreset cross stratification related with bar bed form is the predominant sedimentary structure. Bed forms on the stoss side of bars have been occasionally preserved. Bar units are isolated or superimposed.In the former situation they may be delimited by erosional surfaces indicating stage changes between the aggradational bar units.

Channel aggradation with cross-channel bars seems to be the predominant characteristic in these examples. Accumulation from cross-channel bars was contemporaneous with aggradation at the channel margins (e.g., P 2 in fig. 14).

Water escape structures appear abundantly. Hydroplastic deformation and pipes are frequently associated. Liquefaction is also present in palaeochannel 1, where it has been eroded and buried by the thicker part of the next channelized body (fig. 14).

Figure 14
Stacked low sinuosity palaeochannels.



STOP 4

Successive aggradation stages of a low sinuosity channel

Km 7.1 between Huete and Moncalvillo de Huete (figs. 9 and 10). Parking site in km 6.9 to the right of the road.

Two low sinuosity palaeochannels can be identified. The lower channel body shows a section parallel to the channel local trend, while the upper one is perpendicular to the channel.

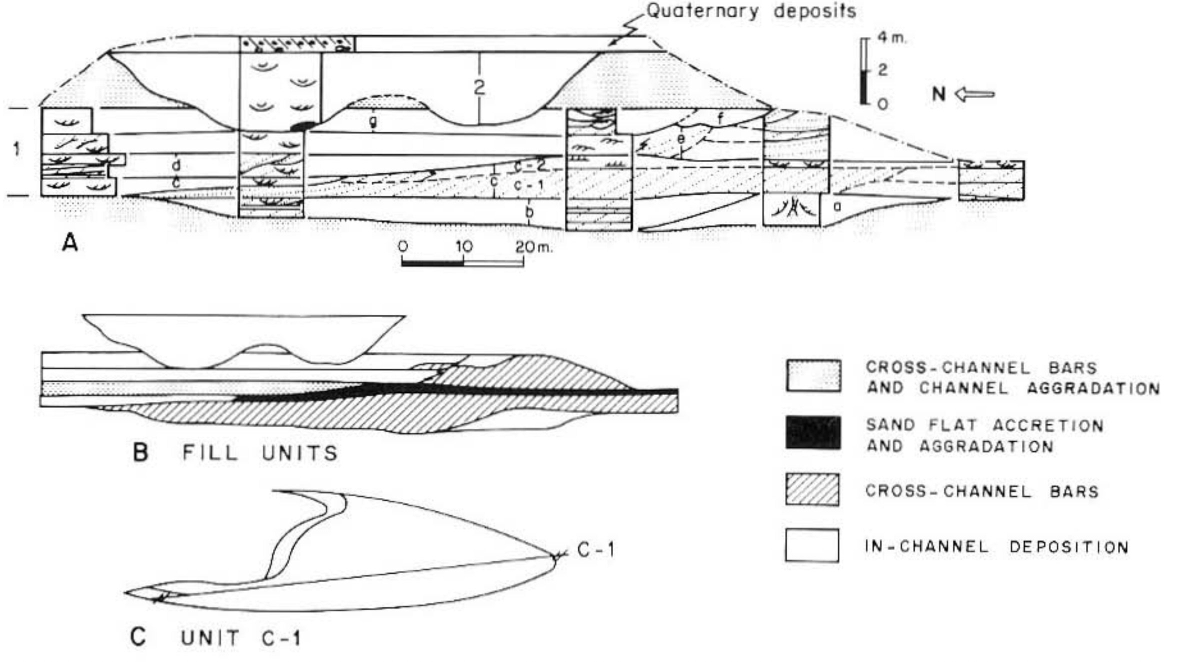
Cross channel bar structures are present in the lower channel. As in previous examples, their formation has been controlled by the position of pre-existing minor channels in the river bed.

The lower channel has been divided into several aggradational units (fig. 15). Four different types of facies and facies associations have been identified similar to those characterising the Saskatchewan River model (Cant & Walker, 1978). They have been named to indicate the process o aggradation that has been deduced: 1) in-channel deposition, 2) cross-channel bar, 3) sand flat aggradation, 4) cross-channel bar and chanel aggradation.

The sand flat in this example (units C1 and C2, fig. 15) contains one large planar tabular set, 130 m long and 2 m thick, with changes in the foreset lamination geometry, associated with the downstream lengthening of the bar (C, fig. 15). This phenomenon could have occurred during the lowering of the river stage, when expanding flow downstream of the nucleus results in the development of horns (Cant & Walker, 1978). At the next high stage, up to 0.75 m of large scale cross stratification was deposited on the bar (C-2, fig. 15), which downcurrent evolves to a new deposit of foreset cross stratification on the lee side of the previous bar.

Sand flat dimensions in this example are similar to those of the Saskatchewan River (Cant & Walker, 1978).

Figure 15
Aggradational units of a low sinuosity palaeochannel.



STOP 5
Example of an intermediate sinuosity palaeochannel

Km 13.8 between Abia de la Obispalia and Villanueva de los Escuderos (fig. 9). Parking site by the road.

The exposure has been formed in a three dimensional-largely arcuate body which is interpreted as a point bar deposit (Díaz-Molina et al., 1985). This interpretation has been based not only on its external geometry, but also because in a section parallel to a bend radius, low angle lateral accretion units are present. In contrast, along the channel bend, the depositional units only reveal vertical aggradation (fig. 16).

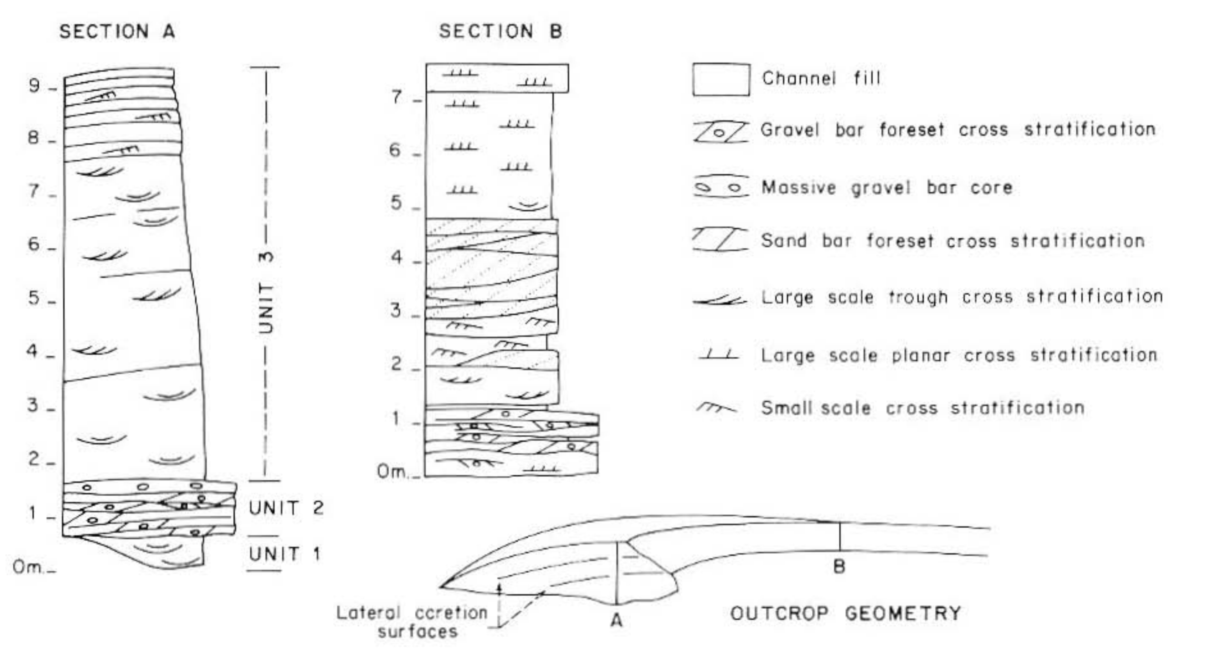
The geometries and facies relationships of this example differ from those of high sinuosity point bar deposits in the same basin. The geometry of the whole body is more elongate than the point crescent bar attached to the inner meander bend of high sinuosity river examples. In addition, the palaeocurrent observations along the whole palaeochannel reveal that the flow lines were nearly parallel to the bend, instead of being directed towards the inner bank.

At the base of the palaeochannel filling, isolated conglomeratic bodies are present. These conglomeratic bodies have been formed mainly by gravel bar aggradation and they are interpreted as composite bars. The discontinuous arrangement of the composite bars, and their location at the base of a point bar deposit suggest formation by riffle and pool bed migration.

In fig. 16 two logs through the arcuate body are presented. The section A is perpendicular to the channel bend, coinciding with a meander bend radius. In this section, unit 3 shows a fining upwards sequence. However, in section B there is no sequential pattern of structures. The foreset cross stratification that appears in section B (fig. 16) is interpreted as the internal structure of bars. These bed forms could have originated from flow expansion on entry into the channel bend.

According to Ethridge and Schumm (1978) there is a simple

Figure 16
Intermediate sinuosity palaeochannel.



and direct method for reconstructing palaeochannel morphology and flow characteristics. The values for this example obtained using this method indicate an intermediate sinuosity.

$W^* = 76$ m	$W = 114$ m
$D^* = 10$ m	$D = 6.5$ m
	$F = 17.5$
	$P = 1.61$
	$Qm = 110\ m^3/s$
	$Qma = 706\ m^3/s$
	$S = 2.8$ m/km

In a previous study (Díaz-Molina et al., 1985) the local exposure was considered as part of the Valdeganga depositional system. However new mapping has revealed that this stratigraphic succession belongs to the Tórtola depositional system.

STOP 6

Stacked and adjacent point bar bodies. Longitudinal section geometry.

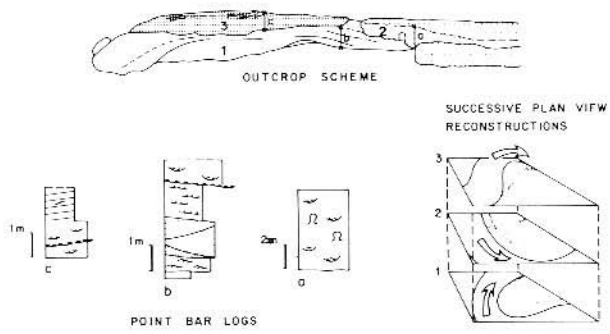
Road to Mazarulleque, at 500 m from Huete (figs. 9 and 10). Parking site by the road.

Three distinctive sandy bodies form this exposure (fig. 17), and are interpreted as point bar deposits. The numbers 1, 2 and 3 in figure 17 indicate the chronologic order of deposition.

The older deposit has a three-dimensional exposure. In spite of lacking a fining upward sequence of sedimentary structures (fig. 17), helicoidal flow conditions can be deduced from the fact that the bed forms moved upwards around the whole bar body. The superimposed lateral accretion units indicate changes in the flow conditions, associated with each lateral accretion unit. The bar has a convex geometry corresponding to a section longitudinal to the meander bend (Díaz-Molina, 1978 and 1984).

Sand body 2 formed around and adjacent the point bar 1. The palaeocurrents indicate an opposite channel direction. Though lar-

Figure 17
Stacked and adjacent point bar bodies.



ge scale cross stratification is the only sedimentary structure in log a (fig. 17), a fining upward tendency can be observed laterally to the south.

Point bar 3 overlies the earlier meander loop topography with an erosional surface. It presents a fining upward sequence which is the most common situation in the point bar deposits of the Loranca Basin.

STOP 7

Meander loop reactivation surfaces. Meander loop reconstruction.

Track to the right, after the bridge over the Mayor River (figs. 9 and 10). Parking site by the row of houses.

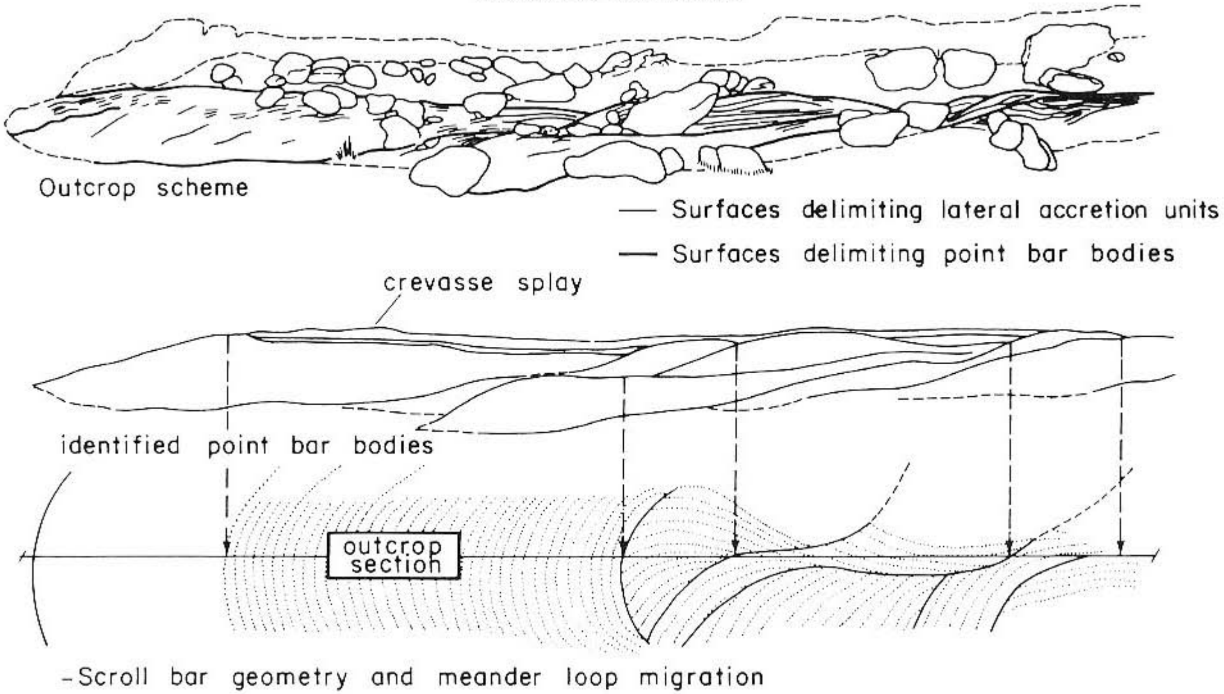
This exposure has been interpreted as a meander loop complex, formed by adjoining point bar bodies (fig. 18). A single point bar consists of a set of conformable lateral accretion units that fine upwards. Each set of lateral accretion units shows a different type o section, transverse or approximately longitudinal; the former are characterized by a convex arrangement of the sets. The whole meander loop section has an undulating upper surface.

Between two adjacent point bars, a well defined surface exists which can be considered a reactivation surface of the meander loop (Díaz-Molina, 1978, 1979a and 1984). Two types of reactivation surfaces are present in this example: erosional and conformable. In either case, the overlying point bar may show onlap over the previous one. The erosional type is distinguished when the underlying point bar has been eroded, and this is analogous with the erosional and slightly discordant lateral accretion surfaces described by Allen (1982). The reactivation surface is conformable when it coincides with the depositional topography of the underlying point bar body.

The changes in point bar migrations of this outcrop have been drawn in figure 18. Estimations of the morphologic and hydrologic characteristics, using the method of Ethridge & Schumm (1978), were made from the transverse section through the last point bar body (Díaz-Molina et al., 1985):

$W^* = 19.4$ m	$F = 7.17$ m
$D^* = 6.3$ m	$Qm = 11$ m^3/s
$W = 29.1$ m	$Qma = 151$ m^3/s
$D = 4.1$ m	$P = 2.06$
	$S = 0.42$ m/km

Figure 18
Meander loop reconstruction.



STOP 8

Meander loop. Preservation of lateral accretion surfaces. Chute channel fill.

Km 0.8 (N–202). Parking site to the right by the junction between the Villalba del Rey road ant the N–202 road (figs. 9 and 10).

Two adjacent point bars are present here (A and B, fig. 19), showing sections nearly transverse to their meander loops. The point bar bodies can be distinguished because of the differences in their vertical thicknesses.

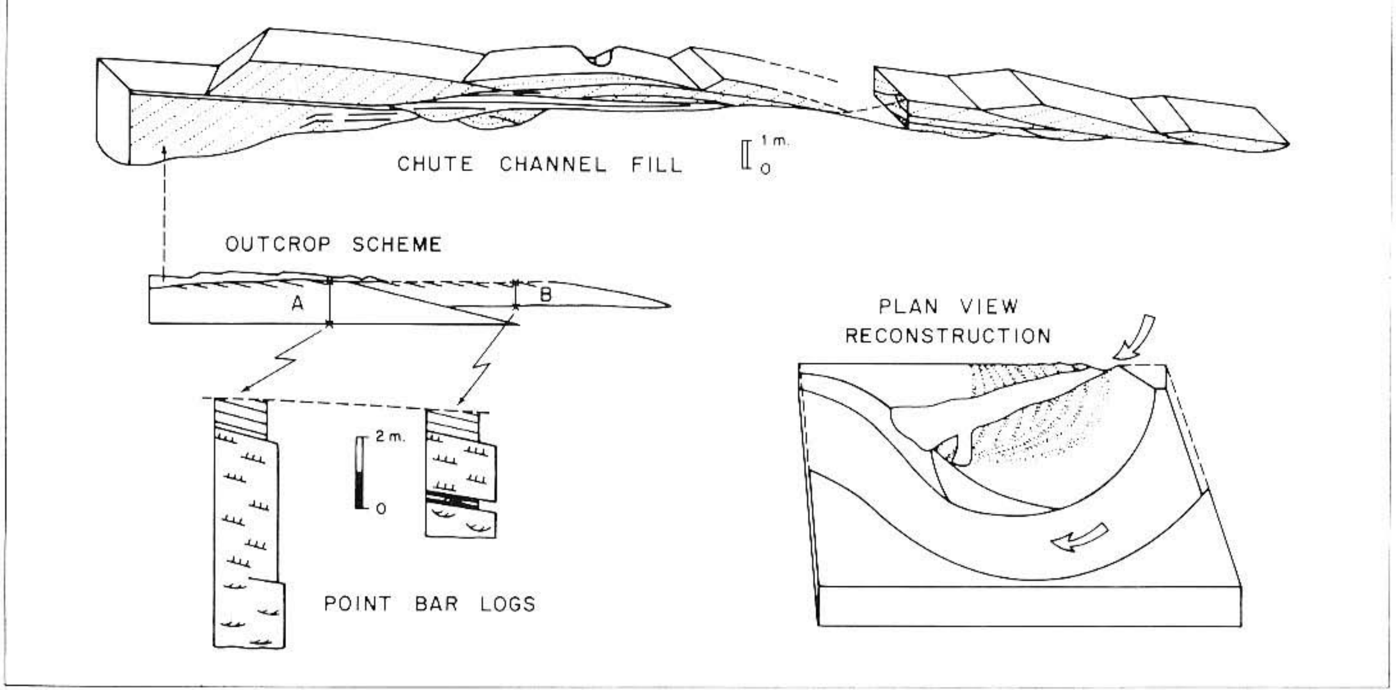
Point bar A shows a fining upward sequence including large scale trough cross stratification, large scale planar cross stratification and small scale cross stratification. In the large scale planar cross stratification, reactivation surfaces, which relate to bed form superposition, are abundant.

The adjacent point bar B, in fig. 19, does not show a fining upward sequence. In the lower sub-unit of the point bar deposit, and interfingering between large scale structures, finer grained deposits with ripple cross stratification have been preserved.

In both point bar deposits the sedimentary structures were made by bed forms that moved up the bar body.

The meander loop formed by the adjacent point bars has been eroded by a chute channel (fig. 18). The chute channel infill consists mainly of foreset cross stratification, and the topography of the bars has been partially preserved.

Figure 19
Meander and chute channel fill.



STOP 9

Development of secondary arcs on the perimeter of a meander loop. Geometry of entire meander loop deposits. Upper transition to gypsum deposits.

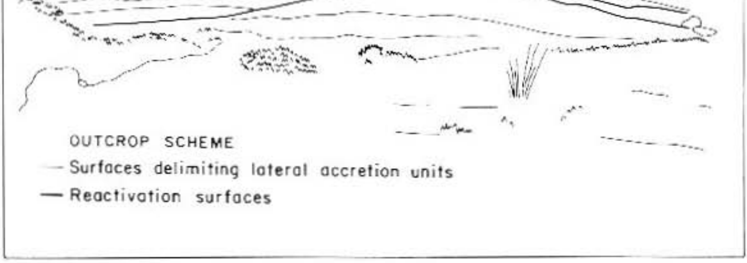
Km 5.1 between Huete and Moncalvillo de Huete (figs. 9 and 10). Parking site to the left of the road.

Coming back to the stop 2 locality and in a higher stratigraphic situation, meander loop complex deposits can be observed (parts 1 and 2 of this stop). Further in the stratigraphic succession, the differences between subunits 1 and 2 of the Tórtola depositional system can also be examined (part 3).

Part 1

Example of meander loop complex. The outcrop is three-dimensional and changes in the direction of migration can be seen on the top of the deposit (fig. 20). The mode of lateral accretion of the point bars has been interpreted using meander growth-patterns of the Beaton river (Hicking, 1974). Figure 21 shows a reconstruction

Figure 20.
Exposed meander loop surface.



of the meander loop growth (Díaz-Molina, 1979a; Díaz-Molina et al., 1985). In recent meandering river deposits it has been observed that the elongation of simple loop does not continue indefinitely, and that they develop a second arc on their perimeters (Brice, 1974). The successive point bar deposits tend to develop when the channel bend reaches a critical curvature, increasing the radius of channel curvature (Hicking, 1974).

The entire body has an upwards-convex upper boundary, and this geometry corresponds to the last point bar, which has been cut longitudinally.

Part 2

The next sandy body has a lateral extent of 400 m and a maximum thickness of 10 m. The body as a whole was formed by both the lateral and vertical stacking of high sinuosity river deposits (Díaz-Molina et al., 1985). The body has a nearly flat erosional base, and its upper surface undulates.

Two distinct aggradational levels can be distinguished (1 and 2, fig. 22). Both aggradational levels were formed predominantly by meander loop deposits.

Figure 21
Development of secondary arcs.

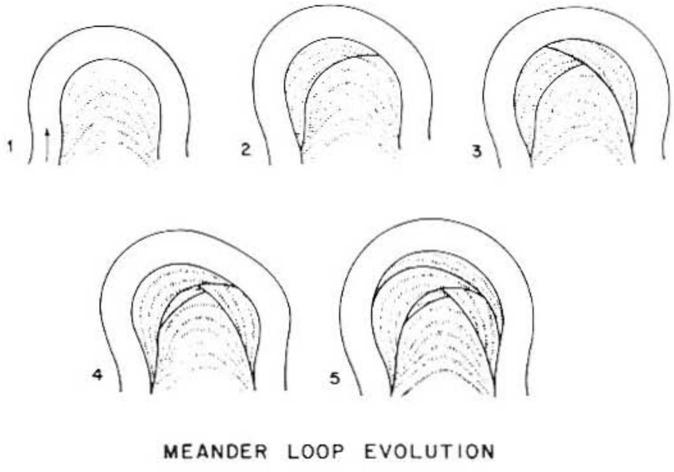
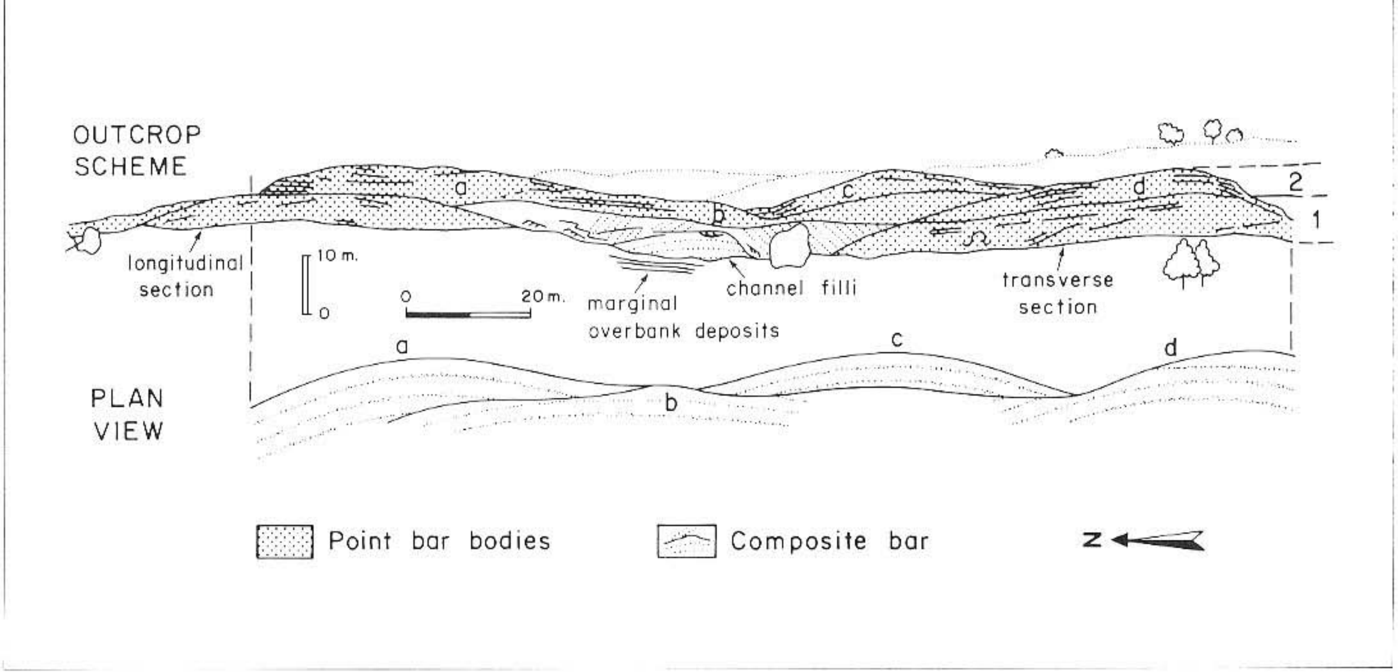


Figure 22
Lateral and vertical stacking of high sinuosity river deposits.



Level 1 shows the largest lateral exposure. Most of the sections cut point bar sequences, but one may correspond to a channel fill. The channel fill is a sandy composite bar deposit (fig. 22), formed by a flow channelized between two meander loop bodies. It could represent a type of chute bar deposit filling an abandoned meandering channel. The convex up arrangement in the bar foreset is interpreted as an accretionary feature resulting from the migration of a linguoid crest.

Level 2 can be examined on the top, and the lateral accretion patterns of the four point bar deposits forming the whole body have been represented in figure 22. The sketch in figure 22 indicates a relatively straight segment of a channel around a large bend.

Part 3

In subunit 2 palaeochannel deposits are less frequent and in general they have smaller dimensions. The colour can be used to distinguish different types of sand bodies in subunit 2. Thus, in subunit 1 sandstone deposits are dark yellow-gray, while in the subunit 2 they are white to light gray bodies. At stop 10, the relationships between sandstone colours and diagenetic events are analysed.

Sandstone extrabasinal components (quartz, K-feldspar and calcareous rock fragments) are present in similar amounts in subunit 1 and 2. Only a slight increase in quartz components is observed in some samples of subunit 2. This composition is characteristic of a Mesozoic rock provenance (sandstone and limestone), and reveals a constant composition of source area.

However, some important differences in intrabasinal grain content between both units exist. First, there is an increase of carbonate intrabasinal grains in subunit 2; and secondly, non-carbonate intrabasinal grains (gypsum grains) appear only in subunit 2.

Diagenetic features also distinguish the sandstones of different units. Scarce cementation in subunit 1 sand bodies takes the form of calcite mosaics. In contrast, abundant cementation by dolomite and gypsum has ocurred in subunit 2 palaeochannels, belonging to the Tórtola fluvial system. Gypsum cement gives a light-gray and white colour to subunit 2 sand bodies.

Remanent primary and a little secondary porosity appears in subunit 1 sandstones. Sandstones of subunit 2 have very little porosity because of occlusion of pores by gypsum cement. However, some sand bodies have suffered an intense decementation by the solution of gypsum cement. These deposits are very dark grey in colour.

STOP 10

Channels with gypsum cement and decemented sand bodies in subunit 2 (Tórtola fan).

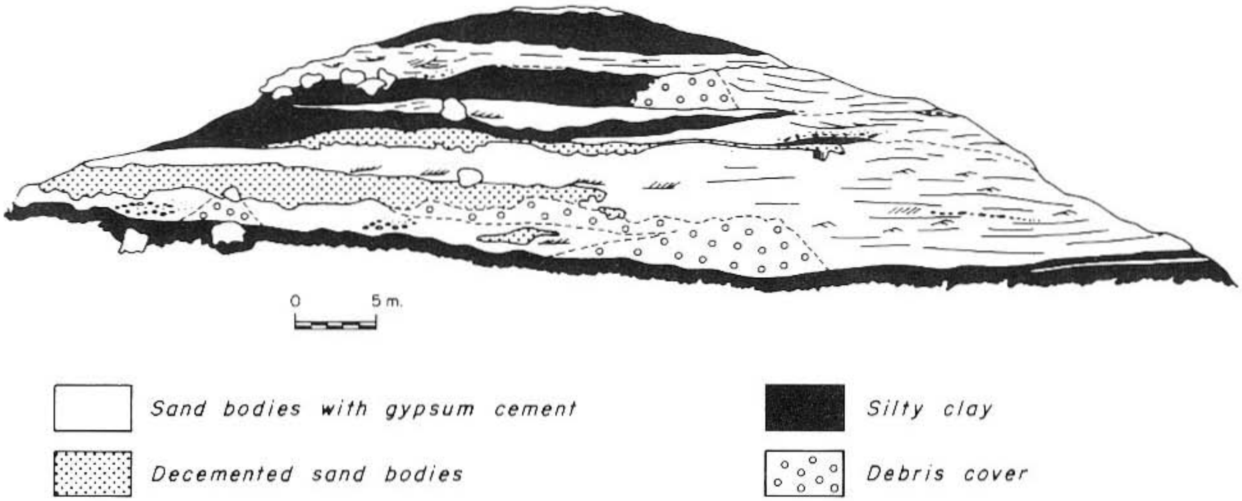
Km 9.3 between Huete and Moncalvillo de Huete figs. 9 and 10). Parking site on a track to the right in km 9.2.

At this locality a palaeochannel with two types of sand bodies can be examined. The particular interest is the contrast in diagenesis between: a) gypsum cemented sand bodies, and b) decemented sand bodies (fig. 23).

Gypsum cemented channels are characterized by light-gray colour. The texture of the cement is poikilotopic and crystals vary in size between 8 and 10 cm. The internal structures of the sand bodies are well preserved, and the cementation affects deposits of all grain size. However, coarse bottom deposits, characterized by a higher content of intrabasinal grains, have more conspicuous dolomite cement.

Decemented sand bodies appear associated with the gypsum cemented deposits. In the exposure these deposits are dark-gray in colour. They show an irregular like-lens shape lacking stratigraphic control. Primary structures are poorly preserved. Often the upper surfaces of these bodies are discontinuity surfaces (e. g.: surfaces between minor channels). However, the lower surfaces are more irregular and cut depositional structures. Frequently, decementation affects the deposits of whole channels. This decementation was due to the solution of gypsum cement and its replacement, which took place in a telodiagenetic stage.

Figure 23
Channel with gypsum cement and decemented sand bodies.



STOP 11
The «tubular chert»: key for the interpretation of gypsum sequences in the Loranca Basin.

Km 5.4 between Huete and Mazarulleque (figs. 9 and 10). Parking site by the road.

At this stop it is possible to see an exposure of gypsum, clays and chert, which correspond to the top of subunit 2 and subunit 3.

Within the stratigraphic succession (fig. 24) there are two main lithofacies of gypsum: powder and compact. The «powder gypsum» is poorly compacted and porous, and in several cases, small and large scale cross-stratification is present. The gypsum crystals are well differentiated and have lenticular shapes. Their lengths usually vary between 50-500, although they may be as long as 1 mm. Scarce pedotubules are contained in this lithofacies.

In the compact gypsum, the rock is composed entirely of vertical and subvertical pedotubules (Brewer, 1964). The gypsum crystals have the same size and shape as those described in the previously-mentioned lithofacies, although they are frequently welded. A longitudinal section through a pedotubule shows successive meniscus laminations. These laminations are formed by lens-shaped gypsum crystals aligned in the direction of their maximum length which is parallel to the crystallographic axis (fig. 24). In cross-section, the gypsum crystals are arranged around a vertical axis, often like concentric bands.

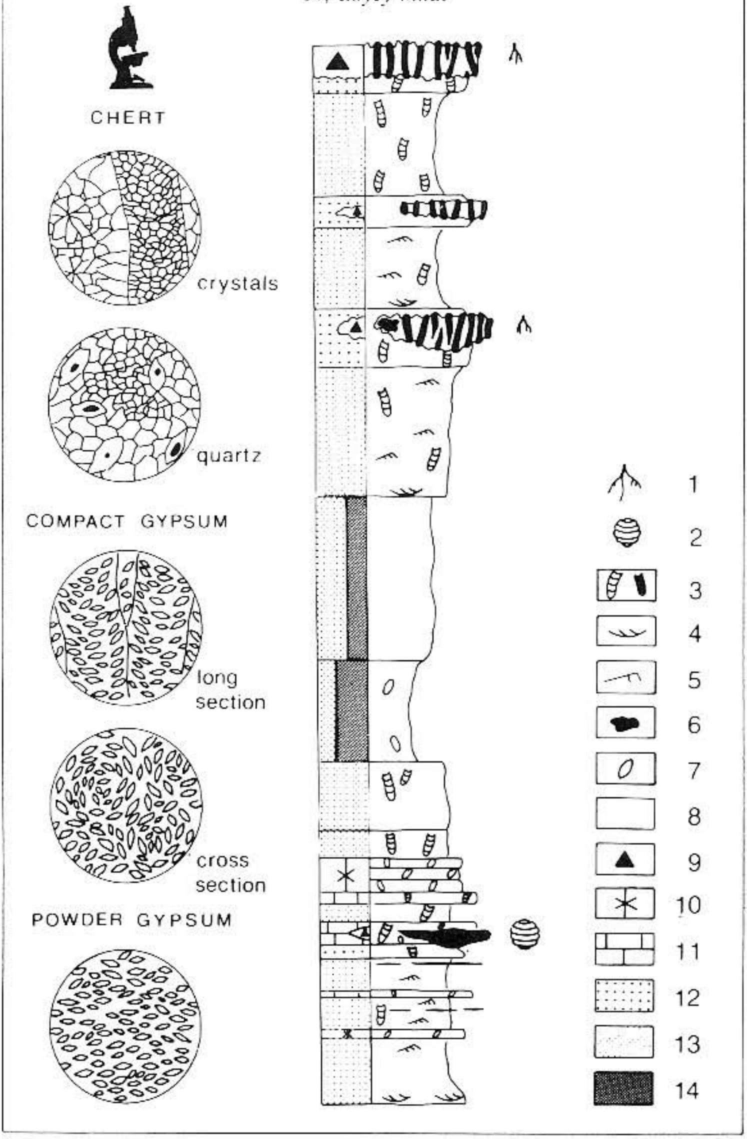
The meniscus laminations are described in the striotubules of Brewer (1964) and Freytet and Plaziat (1982), among others. They are generally considered to be animal burrows.

At the base of the stratigraphic succession (fig. 24), the lenticular gypsum crystals grew in beds of dolomicrite. These beds contain the previously-described bioturbation structures, but lack the clastic bedforms.

Chert forms uneven beds and nodules between a few centimetres and one metre in thickness. Many of them have the tubular bioturbation structures.

The distribution, dimensions and shapes of the tubules are

Figure 24
Log through gypsum deposits at the top of the Upper Unit. 1, rizholits; 2, charophytes; 3, striotubules; 4, large scale cross stratification; 5, small scale cross stratification; 6, massive chert; 7, lenticular gypsum crystals; 8, non identifiable structures; 9, chert; 10, gypsum crust; 11, dolomicrite; 12, compact gypsum; 13, powder gypsum; 14, clayey mud.



more evident in the chert than in the gypsum deposits. The tubules are usually parallel to each other but sometimes they transect each other. Occasionally they penetrate bedding surfaces. The lengths of the vertical or subvertical pedotubules vary from a few centimetres to about 40 centimetres. Their diameters range from 0.5 to about 1 cm, although there is no variation in the diameter of a single pedotubule. They do not present external ornamentation and, generally, they do not show downward bifurcation (less than 1 per thousand of the pedotubules are branched). Both features (constant diameter and no branching), together with the fact that some pedotubules have meniscus laminations, lead us to believe that most of the pedotubules are burrows (Klappa, 1980).

The quartz textures, which include length-slow chalcedony and several types of megaquartz in general, indicate only earlier deposition.

In these cherts, there are many minute organic-siliceous corpuscles and occasionally some spores. These organic-siliceous corpuscles are thought to have been formed during the early stage of the silicification. The first phase os silica precipitation contains previously dissolved organic material and then organic-siliceous corpuscles were produced. This could indicate that the environment of the gypsum host-sediments was very rich in organic matter.

The scarce structures observed in the powder gypsum pose difficulty in interpretation of the sedimentary environments. Apparently the gypsum beds with clastic structures (large and small scale cross-stratification) coexist with primary evaporite gypsum. The reworked gypsum is difficult to distinguish from the primary gypsum because the grains of the gypsarenites have the same textural features as the gypsum primary crystals. Stoops and Ilawi (1981) consider that loose gypsum sands accumulated as a result of mechanical sorting by water movements on the shore of sabkhas. These authors have also observed the compaction of powder gypsum by animal activity.

In the stratigraphic succession there are shallowing sequences, with cherts located at the top of them. These cherts appear always replacing the bioturbated gypsum. These bioturbated gypsum were affected by an early diagenetic silicification which incorporated the dissolved organic material. The silicification was produced when the sheet of water was very thin, near desiccation, or when the gypsum emerged (vadose environments).