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COASTAL PLAIN AND FLUVIAL DEPOSITS IN THE TERTIARY OF CENTRAL AND NORTHERN SPAIN

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Coastal plain and fluvial deposits in the Tertiary of central and northern Spain / Herman Moechtar.
Utrecht: Faculteit Aardwetenschappen, Universiteit Utrecht.
(Geologica Ultraiectina, ISSN 0072-1026; no. 123) Proefschrift Universiteit Utrecht
Met lit. opg.
Met samenvatting in het Nederlands.
ISBN 90-71577-77-5
Trefw.: sedimentologie; Spanje.

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KUSTNABIJE EN FLUVIATIELE AFZETTINGEN IN HET TERTIAIR VAN CENTRAAL EN NOORD SPANJE

(met een samenvatting in het Nederlands en in het Indonesisch)

Proefschrift ter verkrijging van de graad van doctor
aan de Universiteit Utrecht
op gezag van de Rector Magnificus, Prof. Dr. J.A. van Ginkel
ingevolge het besluit van het College van Dekanen
in het openbaar te verdedigen
op dinsdag 17 mei des namiddags te 12.45 uur

door

Herman Moechtar

geboren op 11 mei 1950 te Bukittinggi in Indonesië

promotor: Prof. Dr. D. Eisma

co-promotoren: Dr. P.L. De Boer
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ACKNOWLEDGMENTS

I am grateful to Prof. Dr. M.J.R. Wortel, Prof. Dr. B.H.W.S. de Jong, and Prof. Dr. S.H. White for their support to finish my study at the Utrecht University after almost 5 years under supervision of Prof. Dr. S.D. Nio. Prof. Dr. S.D. Nio initiated this study and introduced me to the field. I would like to thank my Promotor Prof. Dr. D. Eisma for his comments on this thesis.

I am indebted to my Co-Promotor Dr. P.L. de Boer who contributed to this research in the final stage. His help and support was very valuable. He introduced me into sedimentology during fieldwork in Indonesia in 1984 and 1985. This gave me the motivation to continue my study. I am greatly indebted also to my Co-Promotor Dr. Niek Molenaar for his valuable criticism, reviews, and corrections of the text. I am grateful to Dr. Andre van Gelder for his help and comments on part of this thesis. He also introduced me the sedimentology of the Ebro Basin. Moreover, many thanks are due to Drs. Henk de Haas for his help and discussions, particularly for his constructive criticism of Chapter 3. I would like to thank Prof. Dr. A. Tobi and Dr. M.J. van Bergen for their support and attention during my study. I also want to thank Drs. Koos de Vries and Drs. Jan Jaap van den Hurk for their cooperation in the field and for the use of some of their data. Many thanks are also due to Drs. Jeroen de Graaff, Drs. Johan ten Veen, Drs. Maarten Prins, Mr. C.M. van Bemmel, A.M. Pouw-v.d Dolder, Marjan Reith and Tom Zalm. The figures and photographs were produced by the drafting and photography department of the Institute of Earth Sciences in Utrecht.

I wish to thank Prof. Dr. John A. Katili (former Director General of Geology and Mining Resources, Indonesia), Ir. Salman Padmanagara (Director of PIO-ADB, Geological and Mineral Survey Project, Directorate General of Geology and Mineral Resources, Indonesia), and Dr. Ir. Adjat Sudradjat MSc, Deputy Chief of this project (at present: Director General of Geology and Mineral Resources, Indonesia). They supported me during my study in The Netherlands. I am grateful to my former Directors Dr. M. Untung and Dr. R. Sukamto, my Director Dr. Ir. Irwan Bahar, my former Division Heads Drs. R. Wikarno, Ismail Usna MSc (at present: Director of the Marine Geological Institute, Indonesia). They granted me permission to complete my study. Many thanks are also due to the Director of the Volcanological Survey of Indonesia Dr. Ir. Wimpy S. Tjetjep for his help during my study.

I wish to thank many persons for their assistance and help: the State University of Utrecht in The Netherlands; the Indonesian Geological Research and Development Centre; PIO-

ADB, Geological and Mineral Survey Project, Directorate General of Geology and Mineral Resources, Indonesia; and the Embassy of The Republic of Indonesia in The Netherlands, especially the former Attaché of Education and Cultural Affairs Ir. F.B. Mawengkang, the Attaché for Education and Cultural Affairs Drs. Fuad M. Salim, Mr. Supriyono, and Mrs. Ami Kusuma-Soetopo. Many thanks to Steve Silas.

Funding for this research was provided by The Government of The Republic of Indonesia through PIO-ADB, Geological and Mineral Survey Project, Directorate General of Geology and Mineral Resources, Indonesia during 1988-1990; The Dutch Government through The Netherlands Organization for International Cooperation in Higher Education (NUFFIC) from March 1991 until April 1992. From May 1992 until the completion, this research was funded by The Government of the Republic of Indonesia through of the Embassy of The Republic of Indonesia in The Hague.

I like to thank our family, M. Indra Kesuma, Retina Sri Sedjati, Indra and Wahyu, Rio Sutomo, E. Arjakusumah, and Rainy who visited me in Utrecht.

Last but not least I wish to mention the moral support of my wife Rifaida Eriningsih Setiani.

To my parents Sarinah and Moechtar R.M (Alm.),
my brother Muso Indra Kesuma,
my wife Rifaida Eriningsih Setiani,
and our son Rio Alcanadre.

SAMENVATTING (Summary in Dutch)

De stijl van rivier afzettingen is het gevolg van de complexe interactie tussen interne en externe factoren. Interne factoren zijn het meanderen van een rivier en het verlaten en zijdelings verschuiven van geulen, als gevolg van de activiteit van de rivier zelf. Externe factoren zijn het klimaat, de tektoniek en de zeespiegel, die op een grotere schaal invloed hebben op de uiteindelijke rivier afzettingen. Het klimaat bepaalt de hoeveelheid water in de rivier en de regelmaat waarmee het wordt aangevoerd, het type sediment en de vegetatie. De tektoniek en de zeespiegel bepalen de helling vanaf het brongebied tot aan de zee en daarmee de energie van het systeem. Binnen deze randvoorwaarden zijn het klimaat en ook de interne factoren van invloed in op het rivier systeem.

De tektonische activiteit en verwering in het brongebied bepalen de beschikbaarheid van klastische sedimenten, terwijl daarbij de tektonische activiteit ook de bodemdaling in het bekken bepaalt. Bodemdaling en veranderingen van de zeespiegel bepalen de mogelijkheid van sediment accumulatie.

De externe en interne factoren tezamen bepalen de uiteindelijke vorm en opbouw van de sedimentaire successie. Vlechtende riviersystemen vindt men veelal in gebieden met onregelmatige debiet, steile hellingen en grote hoeveelheden grofkorrelig sediment. Meanderende rivier systemen vindt men daarentegen meer in vlakkere gebieden met een meer regelmatige afvoer en grotere hoeveelheden gesuspendeerd materiaal relatief te opzichte van grofkorrelig materiaal. Dit laatste wordt voornamelijk langs de bodem van de rivier rollend vervoerd. Ook de aard van het overstromingsgebied is van belang, zowel wat betreft de aard van het sediment als wat betreft de begroeiing. Tezamen bepalen die de weerstand tegen geulerosie en het zich zijdelings verplaatsen van de rivier.

Er zijn dus een aantal onderling gerelateerde variabelen die de ontwikkeling van een rivier systeem en dus de aard van de uiteindelijke sedimentaire opeenvolging bepalen. In het verleden werd vaak gedacht dat klimaatsinvloeden niet erg belangrijk zijn en alleen een signaal geven dat ondergeschikt is aan dat van de tektoniek. De tektoniek werd gedacht het depositionele systeem te beheersen. In de laatste tijd wordt steeds meer ingezien dat het klimaat een onafhankelijke en zeer belangrijke impact heeft op de rivier processen en op de uiteindelijke sedimentaire produkten.

Vooral het inzicht dat astronomische invloeden leiden tot periodische veranderingen van het klimaat op aarde, heeft ervoor gezorgd dat ook de herkenning van dat signaal en het begrijpen ervan grote stappen voorwaarts hebben gemaakt.

Bij veel studies van riviersystemen is gebleken dat fluviatiele sedimentaire successies veelal een cyclisch karakter hebben. In deze studie wordt getracht de oorsprong van dergelijke cyclische sedimentatie patronen te begrijpen.

In de laatste tijd is veel aandacht geschonken aan werk van Vail et al. (1977) en Posamentier & Vail (1988) over de invloed van zeespiegel bewegingen op sedimentatie patronen. Veranderingen in de hoogte van de zeespiegel hebben de potentie om cyclische sedimentaire sequenties te produceren. In continentale afzettingen is het effect van zeespiegel bewegingen afwezig of marginaal, alhoewel de zee wel degelijk enige invloed kan hebben op het lokale klimaat. In dergelijke gevallen ligt het veel meer voor de hand om in geval van cyclische sedimentaire opeenvolgingen te denken aan de invloed van tektoniek en klimaat. Veranderingen in de tektoniek en in het klimaat, de laatste vooral in termen van neerslag, worden verondersteld belangrijke veranderingen teweeg te kunnen brengen in rivier systemen.

De mogelijke effecten van de klimatologische omstandigheden op rivier afzettingen zijn in het verleden gedocumenteerd en bediscussieerd door bijvoorbeeld Le Tourneau (1985) en Olsen (1989). Perlmutter & Matthews (1989) laten zien dat Milankovitch invloeden periodiek verschuivingen van de klimaatgordels op aarde veroorzaken. Lokale veranderingen van het klimaat worden gedacht effect te hebben op rivier afzettingen, omdat daardoor zowel de rivier afvoer als ook de hoeveelheid vervoerd materiaal verandert en rivier systemen zich daaraan aanpassen. Het is daarom mogelijk dat ook fluviatiele afzettingen een cycliciteit kunnen vertonen met tijdsperioden uiteenlopend 20.000, 40.000, 100.000, 1.3 miljoen en 2 miljoen jaar (cf. De Boer & Smith, 1994).

Rivier sedimenten, in het bijzonder grofkorrelige geulafzettingen zoals zandstenen en conglomeraten, behoren tot de belangrijkste reservoir gesteenten voor olie an gas. Vooral de geometrie van geulafzettingen van puinwaaiers langs gebergten en de rivier systemen in riviervlakten is belangrijk. Het is daarbij van belang hoe grove, permeabele geulopvullingen ten opzichte van elkaar en van fijnkorrelige, niet permeabele afzettingen op het overstromingsgebied, voorkomen. Vooral de onderlinge connecties tussen geulopvullingen en de afmetingen van afzonderlijke geulopvullingen bepalen de kwaliteit van een reservoir. Deze eigenschappen worden, indirect, vooral bepaald door de tektoniek, het klimaat, de zeespiegel, of een combinatie van deze drie factoren. Wanneer de factoren die deze eigenschappen bepalen beter worden begrepen, kan ook de potentiële kwaliteit en het gedrag van reservoir gesteenten beter worden voorspeld.

Het doel van deze studie is het herkennen en interpreteren van cyclische opeenvolgingen in rivier sedimenten aan de hand van rivier afzettingen in drie gebieden in Spanje.

Rivier afzettingen in de drie verschillende gebieden in noord Spanje zijn onderzocht: het Loranca Bekken, waar tektoniek en klimaat een rol speelden, het Ebro Bekken, waar alleen het klimaat varieerde terwijl de tektonische activiteit redelijk stabiel bleef, en het Tremp-Graus Bekken, waar naast marine invloeden ook tektonische activiteit en klimaat een duidelijk cyclisch signaal in de sedimentaire kolom teweeg hebben gebracht. In het Tremp-Graus Bekken heeft de nabijheid van de zee soms tot duidelijke mariene beinvloeding geleid. In ieder van deze drie gebieden is een cyclisch patroon duidelijk in de sedimentaire opeenvolging herkenbaar. Deze cycli worden geïnterpreteerd in termen van grootschalige veranderingen van tektonische activiteit met daarop gesuperponeerd kleinschalige veranderingen van het klimaat. Veranderingen van de zeespiegel spelen alleen een rol in het Tremp-Graus bekken.

RINGKASAN (Summary in Indonesian)

Berbagai aspek "sedimentologi" didaerah ini, secara umum telah dilakukan pengamatannya oleh beberapa peneliti sebelumnya. Khususnya, penafsiran akan model dan fasies endapannya telah dikembangkan.

Rangkaian asal-usul pembentukan fasies batuan sedimen, dapat diyakini adalah merupakan suatu rekaman peristiwa yang terjadi secara teratur dalam kurun waktu geologi, yang erat kaitannya dengan "stratigraphy". Atau dengan kata lain, adalah merupakan suatu proses dari siklus rangkaian kejadian (cyclicity in time) yang terbentuk secara teratur. Sasaran utama penelitian ini diantaranya adalah mencoba mengkaji perkembangan dari fasies endapan di daerah "coastal" dan endapan "fluvial".

Untuk mencapai sasaran penelitian, telah dilakukan pengamatan di 3 (tiga) cekungan endapan "alluvial-fluvial", yaitu:

- Endapan "fluvial" di cekungan Barbastro termasuk utara cekungan Ebro (Spanyol utara) (bab 2).
- Endapan "alluvial" kipas Tortola di cekungan Loranca (Spanyol tengah) (bab 3).
- Endapan "alluvial" di cekungan Tremp-Graus (Spanyol utara) (bab 4).

Hasil pengelompokkan atau tataan secara geometri dari bangunan tubuh batupasir (sandstone bodies) di ketiga cekungan tersebut diatas, selanjutnya diharapkan akan dapat memberikan suatu ide dalam menelusuri genetik rangkaian ataupun hubungan fasies batuannya didalam menelaah siklus proses pengendapannya. Selain itu, diharapkan pula dapat digunakan sebagai parameter dalam mempelajari endapan "coastal plain" dan "fluvial" di tempat lain.

Cekungan Ebro yang proses pengendapannya berlangsung pada Eosen akhir hingga awal Miosen, suplai endapannya terutama bersumberkan dari pegunungan "Pyrenees" di sepanjang utara cekungan (Hirst and Nichols, 1986). Cekungan Loranca yang ditutupi oleh endapan berumur Oligosen akhir hingga Miosen awal, memperlihatkan pola singkapannya yang membujur dan menyebar berarah utara-selatan. Cekungan ini terbentuk didaerah depresi (a marginal molasse-filled depression), yang berawal dari sebuah perkembangan kejadian perlipatan antara "Iberian range" dan "Cratonic block" dari "Castilian Meseta" (Diaz-Molina _et al._, 1985). Cekungan Tremp-Graus di selatan pegunungan "Pyrenees", ditutupi oleh endapan yang berumur Mesozoik hingga Tersier. Pembentukan pegunungan ini erat kaitannya dengan tahap konvergen yang terjadi di

akhir Kapur hingga Miosen, yang ke arah utaranya dibatasi oleh adanya pengaruh tekanan/ dorongan (underthrusting) lempeng Iberian dibawah Eurasia (Muñoz, 1991).

Cekungan Ebro

Secara umum, tubuh batupasirnya memperlihatkan bentuk yang lurus memanjang (sheet) yang kenampakkannya dapat sederhana (simple) atau rumit (complex). Rangkaian endapan mana dapat dibedakan menjadi 23 tubuh batupasir (A.1 hingga I.2) terdiri dari tipe fasies endapan "braided stream", "low" dan "high sinuosity". Dari rangkaiatan urut-urutan pembentukan bangunan tubuh batupasir ini, selanjutnya dapat dikelompokkan menjadi 4 (empat) interval. Adanya perulangan dari perubahan tipe fasies endapan tubuh batupasir tersebut, dapat memberikan petunjuk akan keterkaitannya terhadap perubahan peredaran iklim. Ini memberikan suatu keterangan bahwa selama proses pengendapan berlangsung telah dipengaruhi oleh adanya suatu perubahan atau pergantian iklim yang erat kaitannya dengan siklus Milankovitch. Perubahan mana kemungkinan adalah efek dari suatu peralihan jalur iklim utama (major climatic belts).

Simpulan yang diperoleh dari pembahasan bab ini adalah bahwasanya faktor utama dari pembentukan pada sifat dan corak endapannya adalah berhubungan erat dengan suatu siklus peralihan iklim, "subsidence", dan suplai endapannya. Kemungkinan ini adalah merupakan suatu bagian atau secara penuh, astronomi dikaitkan sebagai adanya gejala pergantian dari peralihan suatu iklim.

Cekungan Loranca

Fasies endapan "alluvial" di cekungan ini dapat dibedakan menjadi fasies berbutir halus dan kasar. Fasies berbutir halus dicirikan oleh endapan-endapan dataran banjir dan "crevasse", "playa-lake", dan "lacustrine". Sedangkan facies berbutir kasar adalah merupakan hasil dari suatu endapan "fluvial", terdiri dari fasies endapan "low-sinuosity" (straight), "high-sinuosity" (meandering), dan "braided stream". Dari pola rangkaian pembentukan endapannya, selanjutnya dapat dibedakan menjadi 7 (tujuh) kelompok interval (interval A.2-A.8) yang satu sama lainnya mempunyai perbedaan sifat dalam banguntubuh endapannya. Adanya pergantian ataupun perubahan fasies tersebut, dapat ditafsirkan sebagai hasil dari pemberhentian suatu aktifitas sungai. Sedangkan sejarah pembentukan tubuh alur endapan "fluvial" ini, adalah merupakan suatu hasil proses pengikisan yang berkaitan dengan perubahan maksimum dari suatu perioda aktifitas sungai.

Ciri daripada urutan pembentukan rangkaian pengendapan diatas, selanjutnya dapat memberikan suatu model dalam pengujian adanya pengaruh suatu peralihan iklim dari suatu perkembangan proses pengendapan di cekungan. Kesimpulan mana dapat disebutkan bahwa adanya perubahan susunan bangunan batupasir ke arah tegak dari sistem endapan kipas Tortola, akan berhubungan erat pada perubahan dari suatu daur pergeseran iklim yang menghasilkan beberapa bangunan interval. Ini adalah sebagai petunjuk bahwa astronomi erat hubungannya dengan suatu perubahan iklim.

Cekungan Tremp-Graus

Fasies batuan di daerah in dapat dibedakan menjadi fasies batuan endapan "alluvial" berbutir halus dan kasar, dan tubuh batupasir hasil endapan "fluvial". Sasaran utama penelitian ditujukan dalam menganalisa hubungan fasies batuan tersebut secara tegak dan mendatar dalam urut-urutan stratigraphynya.

Adanya perbedaan secara 3 dimensi dari tubuh bangunan batupasir didaerah ini telah didiskripsi dan didiskusikan. Endapan mana terdiri dari endapan-endapan tubuh "alluvial-sheet" termasuk endapan "proximal" dan "distal" kipas "alluvium", "low-sinuosity channel" (ribbon-like alluvial bodies), dan "high-sinuosity channel" (tabular alluvial bodies).

Dari rangkaian dan urut-urutan bangunan endapan diatas, selanjutnya dapat dibedakan menjadi 5 (lima) kelompok endapan (Interval 1-5). Dari hasil pembahasan dapat disimpulkan bahwa setiap interval tersebut adalah merupakan hasil suatu siklus "progradational" dari aktifitas kipas "alluvium" sepanjang tepian (margin) utara cekungan. Gejala mana adalah efek dari suatu perulangan aktifitas di daerah tektonik (tectonic regime). Oleh karena itu, sebuah rangkaian pembentukan siklus endapan di cekungan ini akan memberikan kala yang panjang (long-term) dari suatu sistem pengendapan pada peristiwa "subsidence".

Efek dari suatu peristiwa tektonik dan perubahan iklim adalah merupakan faktor utama dalam peramalan suatu bentuk rangkaian endapan dalam setiap interval. Ciri ataupun sifat dari fasies endapannya telah didiskusikan pada bab 4. Dapat disimpulkan bahwa rangkaian dari pola lapisannya (stratal pattern) berhubungan erat dengan tektonik pada evolusi pembentukan cekungan, termasuk bentuk cekungan dari perkembangan suatu "subsidence". Akan tetapi tidak ditemukan adanya permukaaan erosi utama (major erosional surfaces) atau torehan lembah (incised valley). Gejala mana memperkuat dugaan bahwa rangkaian dari interval yang terbentuk adalah termasuk sebagai sebuah sistem "progradational" kipas "alluvial" seperti yang telah dikemukakan. Efek dari siklus iklim telah ditafsirkan pula, yang kisarannya antara lebih kering (more arid) hingga lebih lembab (more humid). Adanya relatif "sea-level changes" yang memberikan efek terbentuknya endapan laut dan pantai, adanya perubahan garis pantai kearah timur, dan juga secara lokal ke utara di saat suplai endapan menurun, gejala mana diperkirakan hanyalah merupakan suatu proses dari relatif "sea-level changes" yang berkaitan erat dengan keadaan lokal didaerah ini.

Sumber utama dari pembentukan kumpulan endapan di daerah sumber yang menerus ke arah cekungan "subsidence" adalah berasal dari efek aktifitas tektonik "uplift" yang berupa "thrust-sheet loading" didaerah ini. Dalam sekala besar, rangkaian endapan ini dapat menghasilkan suatu siklus "coarsening-upwards" pada sistem utama dari "progradation" kipas "alluvium". Rangkaian siklus yang ukurannya lebih kecil dapat pula terjadi dalam sistem tersebut, seperti 5 siklus interval dalam penelitian ini.

Banyak faktor yang berpengaruh dalam menelusuri suatu proses pembentukan dari "fluvial systems". Dimasa lalu penafsiran akan adanya dampak dari suatu peralihan iklim pada proses pembentukan di endapan "alluvial-fluvial" sangat langka penerapannya, dimana perhatian umumnya tercurahkan pada tektonik sebagai faktor penyebab utamanya. Pada akhir-akhir ini, para ahli telah mensepakati bahwa faktor iklim adalah merupakan hal yang terpenting, dikarenakan merupakan kekuatan tersendiri (independence force) dalam menelusuri sifat dari fasies endapan "alluvial-fluvial" tersebut. Terutama dalam pengakuan serta pengujian akan siklus Milankovitch, yang akhir-akhir ini sangat pesat perkembangannya.

Bukti akan adanya pola-pola dari suatu siklus rangkaian urut-urutan pengendapan pada sistem "alluvial" telah dikemukakan oleh berbagai peneliti. "Eustatic sea-level changes" yang dicetuskan pertama kali oleh Vail et al. (1977), adalah merupakan faktor pengontrol yang ampuh dalam mempelajari endapan laut dan peralihan (lowland). Hal mana dikarenakan "sea-level changes" dapat memberikan suatu tanda-tanda adanya suatu perulangan dalam daya tampung endapan, yang memberikan suatu rangkaian siklus urut-urutan pengendapannya. Di daerah kontinental, efek daripada "sea-level changes" tersebut jelas sulit ditelusuri. Oleh karena itu, suatu perulangan dari tektonik dan peralihan dari suatu iklim, adalah faktor terpenting dalam mempelajari endapan "alluvial-fluvial" baik didaerah luar jangkauan "sea-level changes" maupun dalam pengaruh "sea-level changes".

Efek akan adanya suatu peralihan iklim di daerah endapan kontinental telah dipelajari berbagai peneliti, diantaranya oleh Le Tourneau (1985) dan Olsen (1989). Akan tetapi suatu pemikiran ataupun ide adanya efek pada perulangan suatu priodik iklim pada sistem "fluvial" adalah relatif baru. Perlmutter & Matthews (1989) memperkenalkan suatu hipotesa dalam "cyclostratigraphy", yang dipusatkan pada pemikiran bahwa siklus Milankovitch dapat ditelusuri dari suatu periodik pengendapan tertentu. Siklus mana dapat diakibatkan oleh adanya suatu perubahan atau perulangan dari suatu peredaran utama atmospir (major atmospheric circulation cells). Oleh karena itu, akibat adanya perubahan atau perulangan iklim tersebut, akan memberikan suatu tanda-tanda bentuk dan perbedaan gaya di endapan "fluvial", seperti adanya perubahan hasil ataupun terhentinya suatu proses pengendapan (Perlmutter & Matthews, 1989). Waktu dan lamanya suatu siklus tersebut sudah barang tentu akan mempunyai suatu kisaran. Suatu pergantian siklus dalam bangunan "alluvial style" di suatu cekungan mungkin dapat memberikan keterangan akan keadaan lamanya pembentukan sebuah interval yang kisaran kurun waktunya antara 20 ka, 40 ka, 100 ka, 1,3 Ma dan 2 Ma (Milankovitch cycles cf. De Boer & Smith, 1994).

Simpulan akhir penelitian ini memberikan suatu masukan, bahwa:

- Cekungan Ebro semata-mata mewakili endapan "fluvial" yang relatif tinggi, tapi stabil akan tektonik dalam hubungannya dengan "uplift" dari suatu aktifitas di pegunungan "Pyrenees" dan sesar naik. Pola siklus endapan "fluvial" mempunyai sifat siklus Milankovitch yang erat kaitannya dengan pergantian suatu iklim (climatological changes).
- Cekungan Loranca memberikan kejelasan bahwa tektonik dan iklim adalah faktor utama dalam menelusuri rangkaian dan sebaran fasies endapannya.
- Cekungan Tremp-Graus menunjukkan bahwa aktifitas tektonik dan perubahan iklim berpengaruh sangat besar pada pembentukan siklus rangkaian serta urut-urutan pengendapannya.

CHAPTER 1

INTRODUCTION

Fluvial depositional style is the sedimentological response to a complex interaction of autogenic and allogenic controls (Allen & Allen, 1990). Autogenic mechanisms, such as meandering, avulsion, and channel shifting, are internally generated processes, specific to a particular fluvial system. They work on a smaller scale than allogenic mechanisms. Primary allogenic controls are climate, tectonics and sea level. The climate determines water and sediment availability as well as the type of vegetation. Tectonics and sea level set boundary conditions to the energy of the system. Tectonic setting and tectonic activity also influence the source area and thus the potential availability of sediment. Autogenic and allogenic processes together determine the final depositional style. Braided streams are favoured by high and variable discharges, steep slopes and large amounts of bedload. High-sinuosity, meandering systems are common in regions with a low gradient, perennial and regular discharge and a relatively high percentage of suspended load relative to bedload. The nature of the floodplain deposits and of the vegetation are important, particularly the coherence of the floodplain deposits and their resistance to erosion.

The development of a specific fluvial system is thus controlled by a number of partly interrelated variables. Processes as well as products of fluvial systems are influenced by these factors. In the past, climatic influences were generally considered to cause only minor overprints on a largely tectonically controlled system. In recent years it has generally become accepted that climate is also an important and independent force in determining the facies characteristics of a fluvial system. Particularly the recognition of Milankovitch cyclicity as a driving force behind the development of repetitive sequential patterns has helped in understanding the impact of climatic factors on sedimentation (e.g., Goodwin et al., 1986; Olsen, 1989; Bull, 1991).

In many alluvial studies, cyclic patterns have been recognized in alluvial sequences. The origin of long-periodic, non-autocyclic, recurring patterns in fluvial sedimentation is the problem central to this study. Since the work of Vail et al. (1977) and Posamentier and Vail (1988) it is clear that eustatic sea-level changes are an important control on sedimentation in marine and lowland sediments. Sea-level changes have the capacity to produce repetitive, cyclic sequences. In purely continental environments, however, the effect of sea-level changes is much less obvious. In areas at a significant height above sea level only indirect effects of sea level may be left, such as changes in the precipitation regime. Changes in tectonic regime and climate should have an important influence upon fluvial sedimentation in areas outside the direct influence of the sea and changes of sea level.

The general effects of climate on continental clastic deposition are well documented and discussed in several research studies (e.g., Le Tourneau, 1985; Olsen, 1989). Ideas about the effect of periodically recurring climatic changes on fluvial style are relatively new. Perlmutter & Matthews (1989) proposed the hypothesis of cyclostratigraphy, which centres around the idea that Milankovitch cyclicity induces a periodic, latitudinal shift of the major atmospheric circulation cells. The resulting changes in climatological conditions will ultimately be translated into different styles of fluvial sedimentation through changes in sediment yield and discharge (Perlmutter & Matthews, 1989). The eventual cyclic changes in alluvial style and basin architecture thus may show recurrence intervals of approximately 20 ka, 40 ka, 100 ka, 1.3 Ma and 2 Ma (Milankovitch cycles cf. De Boer & Smith, 1994).

Alluvial sediments, in particular coarse-grained sandstone and conglomerate bodies, are a major type of hydrocarbon reservoir rock. Especially the geometry of coarse-grained sediment bodies, representing deposition from fluvial channels on alluvial fans and in floodplains, is very important. Fine-grained floodplain deposits often are impermeable with respect to the coarse-grained channel deposits, thus forming seals for fluid flow. The geometry and three-dimensional geometric arrangement or architecture of coarse-grained deposits with respect to fine-grained deposits form the main purpose of most of the current alluvial research. The size of the coarse-grained deposits and the degree to which they interconnect are important. A better understanding of the controls on alluvial architecture will lead to an improved modelling of fluid flow. The already mentioned and often observed cyclic arrangement of alluvial depositional elements is still poorly understood. Either tectonism, sea level, and climate or all of them combined are mentioned as important variables.

In this study, alluvial deposits in three areas in northern Spain have been studied in detail: the Loranca Basin, the Ebro Basin and the Tremp-Graus Basin.

The Ebro Basin represents purely fluvial deposition with a relatively high, but steady tectonic activity in relation to uplift of the Pyrenees and thrust fault activity. The cyclic pattern in the fluvial deposits can be attributed to Milankovitch related climatological changes.

In the Loranca Basin both tectonic activity and climatological conditions controlled the style of the alluvial channels and the alluvial facies distribution.

Also in the Tremp-Graus Basin both tectonic activity and climate influenced the alluvial deposition and caused cyclic alluvial deposition, while, moreover, the depositional system was influenced by changes of relative sea level..

The aim of this study is to recognize cyclic patterns in the alluvial sedimentary successions in the three above mentioned areas in Spain. Moreover, the depositional cycles will be interpreted in terms of changes in tectonic activity, in climatological conditions and in sea level.

CHAPTER 2

FLUVIAL DEPOSITIONAL STYLE AND CLIMATE CYCLES IN THE NORTHERN EBRO BASIN (OLIGOCENE-MIOCENE), SPAIN

INTRODUCTION

Fluvial deposits in the northern part of the Ebro Basin in northern Spain (province Huesca) have been studied (Fig. 1). Good exposures along the Rio Alcanadre allowed a detailed study of coarse-grained fluvial channel bodies. This river cuts a north-south oriented valley through the Oligocene-Miocene fluvial deposits of the Ebro Basin (Fig. 2). Exposures along both valley walls show a large variability of facies, providing evidence for changes in depositional mechanisms during the accumulation of the studied sequence. Data from exposures at both sides of the river valley have been recorded on detailed logs. Fluvial channel bodies, showing different fluvial styles, are separated by laterally persistent floodplain deposits. Their geometry allowed to lateral correlations to be made of channel bodies by tracing the floodplain layers in the field. Drawings and panoramic photographs have been made to help reconstructing the fluvial architecture. Systematic facies changes in the sequence have been analyzed in order to test whether they might be the product of Milankovitch-induced climatic fluctuations and/or of changes in tectonic regime.

The following activities were undertaken:

- reconstruction of the environmental processes during formation of the deposits with the emphasis on the coarse-grained fluvial channel bodies,
- characterization of laterally persistent intervals characterized by a specific fluvial style, and
- establishment of vertical changes in fluvial style in order to draw conclusions about possible Milankovitch-related climatic changes and changes in tectonic regime.

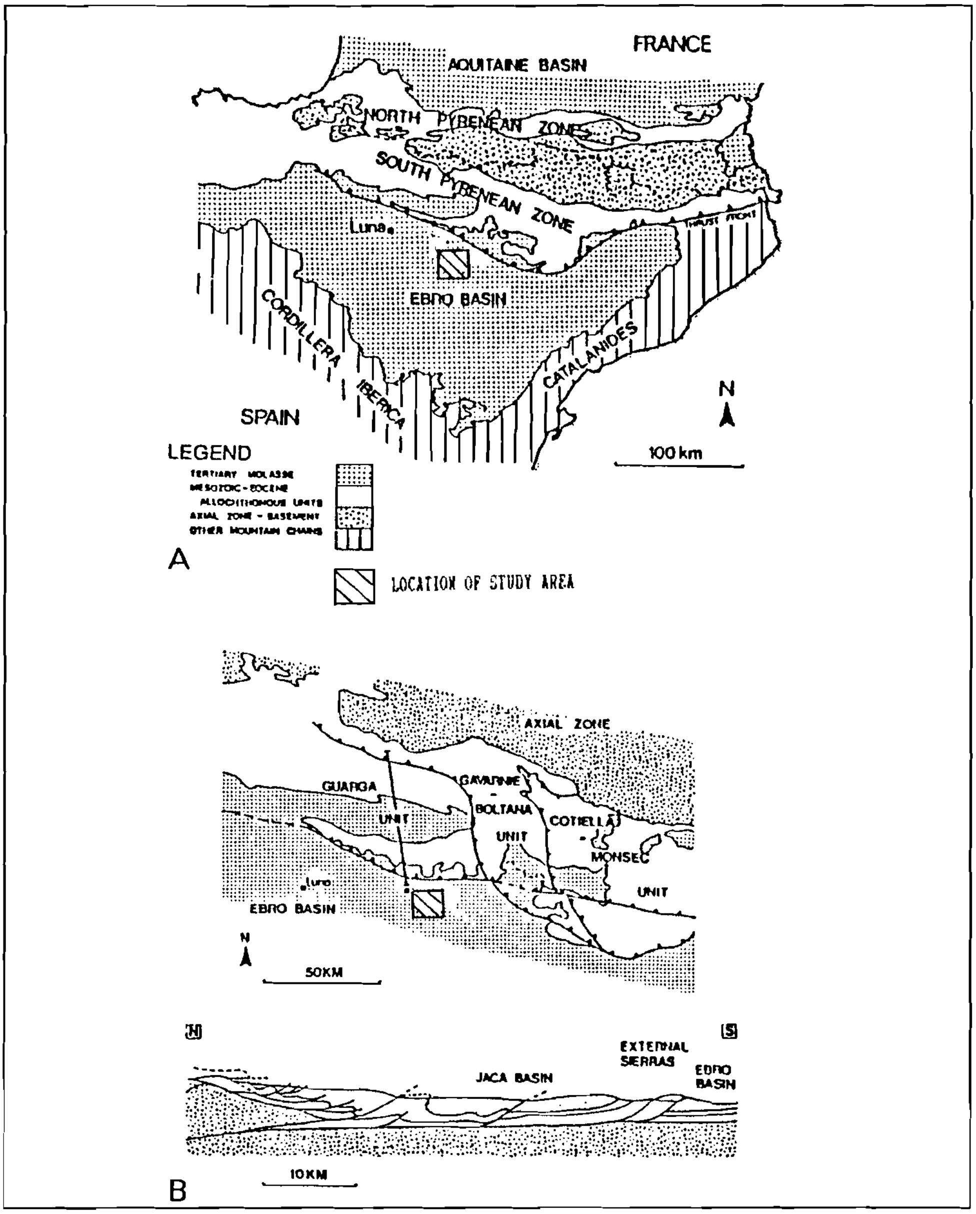


__Figure 1__. (A) General geological setting of the studied area in the Ebro basin. (B) Outline of the structural units of the western part of the south Pyrenean zone. (C) Cross-section through the southern Pyrenean zone along a N-S line north of Huesca, showing the piggy-back nature of most of the Tertiary deposits and the external Sierras forming the northern margin of the Ebro Basin. Slightly modified after Hirst & Nichols (1986).

GEOLOGICAL SETTING

The Pyrenean Foreland Basin

The general structural setting in the Pyrenean foreland Basin is well known from the work of, amongst others, Seguret (1972), Choukroune & Seguret (1973) and Garrido-Megías & Cámara (1983). Hirst & Nichols (1986) included new seismic data in their description of the structural setting (Fig. 1). Three major thrust units, the Guarga, the Gavarnie-Boltaña and the Cotiella-Montsec occur in this area. Each thrust sheet has a decollement horizon in Triassic marls and evaporites, whereas their main bodies consist of Mesozoic to early Eocene limestones. The limestones are covered by late Eocene and younger Tertiary terrigenous as well as marine clastic sediments (Seguret, 1972; Mattauer & Henry, 1974). The thrust movements started in the Eocene and continued through the Oligocene. Therefore, at least the lower part of the clastic Tertiary cover has been affected by tectonic activity. The thrust activity is noticeable in the tectonic as well as in the sedimentary facies pattern (Hirst & Nichols, 1986).

Nijman & Puigdefàbregas (1984) distinguished three basins: the Jaca basin, the Ainsa basin, and the Tremp-Graus basin. These basins were connected during the early Tertiary, forming the southern foredeep basin of the Pyrenees. Tectonic activity in the Eocene and Oligocene caused south to south-west movement of the main thrust sheets. The thrust movements caused a change in the configuration of erosion and sedimentation areas. The older pre-Oligocene molasse basin fills became subjected to partial erosion, and the foredeep shifted southward to form the Ebro basin.

The boundary between the erosional realm of the folded and faulted Southern Pyrenees and the undistorted depositional Ebro basin can be drawn along the leading edges of the Guarga Unit in the west and of the Cotiella-Montsec Unit in the east. The Ebro basin sediments in the area near Barbastro have been folded into an asymmetrical anticline. This is partly due to the movement of the frontal edge of the underlying thrust sheet during deposition of the Ebro basin sediments, and partly to late Tertiary diapirism. There is a distinct morphological and sedimentological break between the folded southern Pyrenean zone and the Ebro basin.

The Ebro basin

The Ebro basin is a large Tertiary basin with an areal extent of nearly 45,000 km^2 (Fig. 1). The triangular depression has at present an altitude of 100 m to 700 m above sea level, and is surrounded by mountain chains. The northern (Pyrenean) range is a major mountain belt, reaching altitudes of 3400 m. It has been tectonically active since late Cretaceous times. Major folding and faulting took place during the Eocene and early Oligocene. Vertical uplift continued through the Oligocene and Miocene. The southwestern (Iberian) and southeastern (Catalonian) ranges are younger and have been less active. As a consequence these areas have been less elevated. Studies of the Ebro basin were published by, amongst others, Riba (1971), Friend et al. (1979) and Riba et al. (1983).

The cross sectional shape of the basin is asymmetrical. The sedimentary fill has a considerable thicknesses near the northern basin margin (4 km according to Hirst & Nichols, 1986), and shows a gradual thinning to a few hundred metres (Riba, 1971) towards the south. The northernmost basinal sediments have locally been overthrusted by Mesozoic and Paleogene thrust nappes, whereas the southern Oligocene-Miocene sediments gradually overstep the older basement rocks. The exposed Oligocene-Miocene sediments comprise a range of continental clastic facies, continental limestones and evaporites. They have not been folded and have hardly been affected by faulting, except for some marginal tectonic features, e.g., the diapiric anticline near Barbastro.

Throughout the Oligocene and most of the Miocene, the Ebro basin had an internal drainage. Clastic sediments were deposited at the basin margins whereas the basin centre is dominated by chemical deposits, mainly gypsum (Allen & Matter, 1982).

Clastic sediment was mainly supplied from the Pyrenees along the northern margin. A number of conglomeratic alluvial fan bodies developed along the northern margin of the Ebro basin (Hirst & Nichols, 1986). They were deposited by rivers spreading radially from point-sources within the surrounding orogenic belt (Allen & Matter, 1982). More distally, the clastic facies consists of fluvial sandstones with well-developed fluvial channel bodies with variable width to depth ratios. Puigdefàbregas (1973) distinguished ribbon and sheet sandstone bodies. These sandstone bodies are surrounded by fine-grained sandstones and mudstones, interpreted as overbank and floodplain deposits. Minor amounts of lignite are also present (Williams & Birnbaum, 1975).

Towards the centre of the basin these clastic sediments grade into thinly bedded lacustrine limestones and marls, and evaporites deposited in and around lakes (Hirst & Nichols, 1986). This chemical facies zone (cf. Allen & Matter, 1982) occupies the axial parts of the basin. Apart from limestones, a main component of this chemical zone is formed by gypsum and gypsiferous mudstones.

The Barbastro/Rio Alcanadre area

The northern margin of the Barbastro area (Fig. 2) is formed by several mountain ranges of folded Triassic to Eocene formations of the Gavarnie Nappe. The folds are principally NW-SE directed, but may locally pass into E-W or N-S oriented folds. The anticlines are tight, whereas the synclines are more open. These mountain ranges mainly consist of Upper Cretaceous and Eocene fossiliferous limestones, although sandstones and shales occur as well (Reille, 1971). Thick deposits of Oligocene-Miocene calcareous conglomerates are found especially along the northern margin of the Ebro Basin. Apart from showing primary depositional (fan)slopes, they have been locally tilted in connection with syn- or post-sedimentary movements of the nappe. The conglomerates pass basinwards into fluvial sandstones and shales. The horizontally lying sediments have been disrupted by an asymmetrical antiform south of the city of Barbastro. This antiform is considered the surface expression of the frontal ramp of the underlying Gavarnie Nappe. Some isolated remnants of the Triassic and Cretaceous basement pierce through the Oligocene-Miocene cover near Barbastro (Fig. 2).

The Barbastro Evaporite Formation (Van Gelder & Dronkert, 1982) crops out in the core of the anticline near Barbastro (Fig. 2b). It consists of fine-grained sandstones, siltstones, and mudstones, interbedded with gypsum layers. The formation is considered to be of late Eocene to early Oligocene age, since in eastern Catalunya the underlying Igualada marls are of Bartonian to Priabonian age (Barnolas et al., 1981; Alvarez Sierra et al., 1990).

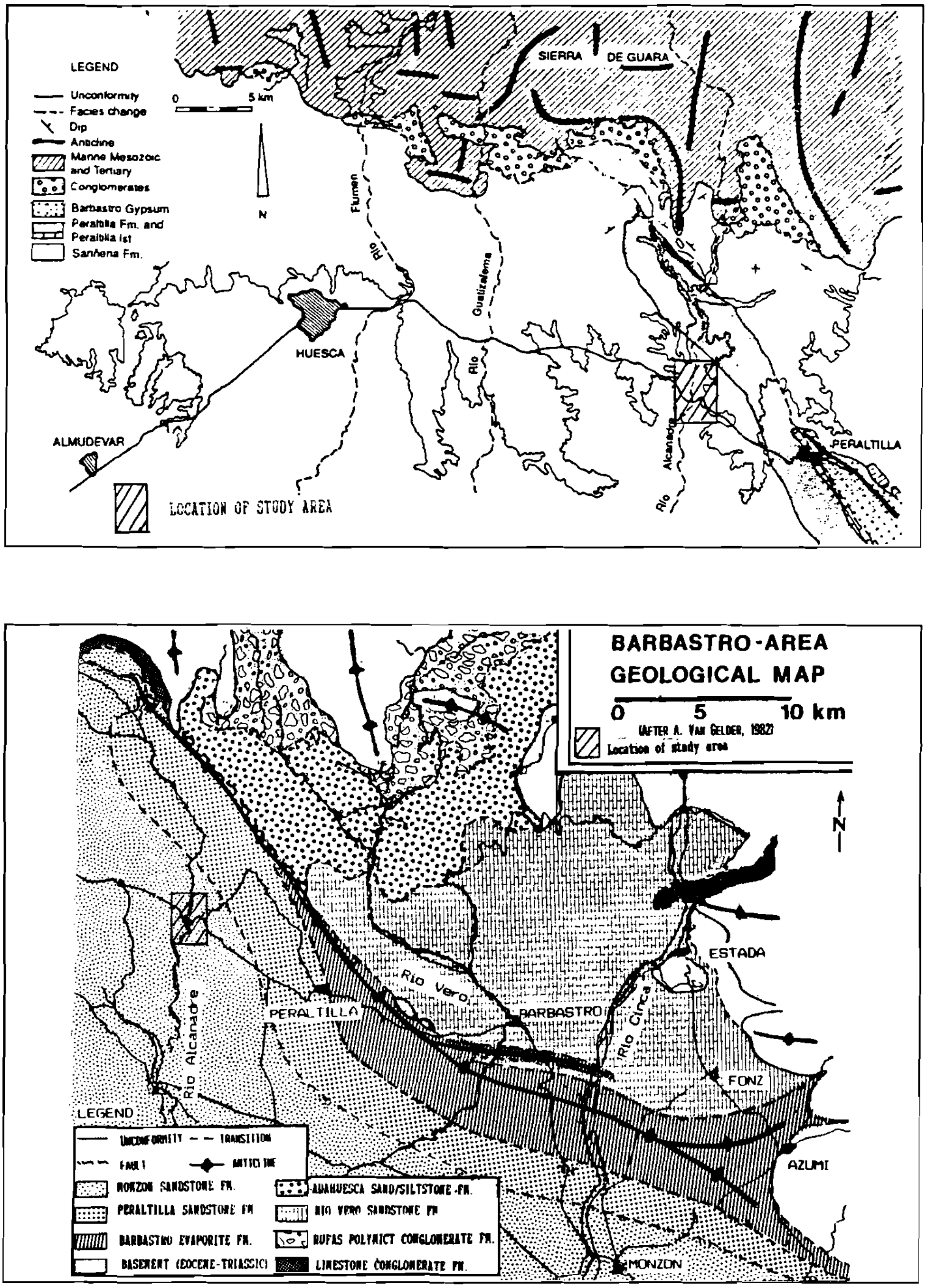


Figure 2. *Location of the studied area. (A) Overview. (B) Simplified geological map of the area, after Van Gelder (1982).*

On top of the Barbastro Evaporite Formation, the Peraltilla Siltstone Formation is found. Its base is formed by a 15 m thick succession of thinly bedded lacustrine carbonates (Alvarez Sierra et al., 1990), which show a rapid transition into overlying detrital sediments. The Peraltilla Siltstone Formation, with some lacustrine limestones at its base, consists of a more than 700 m thick alternation of yellowish-brown lens-shaped sandstones, and reddish fine-grained siltstones, sandstones and mudstones (Crusafont et al., 1966; Spuy, 1983). The entire formation slopes towards the south and southwest, as a result of the updoming of the anticline south of Barbastro. The Peraltilla Siltstone Formation is conformably overlain by the essentially flat lying Monzon Sandstone Formation.

The Monzon Sandstone Formation (Van Gelder, 1968) comprises sheet-shaped sandstones, separated by overbank and floodplain deposits. The sandstone bodies are characterized by poorly sorted, yellow to brown-coloured, medium to coarse-grained sandstones. Floodbasin deposits comprise fine- to very fine-grained sandstones, siltstones and mudstones. The horizontal orientation of the beds and the low percentage of fines result in a fairly monotonous topography: a gently undulating area with hardly any outcrops. Good exposures are only present along the river incisions of the Rio Cinca and the Rio Alcanadre. The latter exposures are subject of this chapter.

Only a very rough indication for the age of the Monzon Sandstone Formation (Fig. 3) can be given, because age determinations are absent in the investigated area. In the vicinity of Huesca (Fig. 3) the age of a vertebrate fauna was determined as Aquitanian (Crusafont et al., 1966). These authors sited the locality in the Peraltilla Formation. Field correlations of van Gelder (1968), however, suggest a position in the lower part of the Monzon Formation. New data on palaeoecology and biostratigraphy in the north-central part of the Ebro Basin, including the Barbastro area, have been published by Alvarez Sierra et al. (1990). Their stratigraphical interpretation of the fossiliferous localities is represented in Figure 3. The Monzon Sandstone Formation is equivalent to the "Formatiente Santa Cilia" being the lowermost part of the Sarinena Formation. Because the contact between the Monzon Sandstone Formation and the Peraltilla Formation is concordant and gradual, the Monzon Sandstone Formation may be of late Oligocene to early Miocene age. It is obvious that this age is only an approximation.

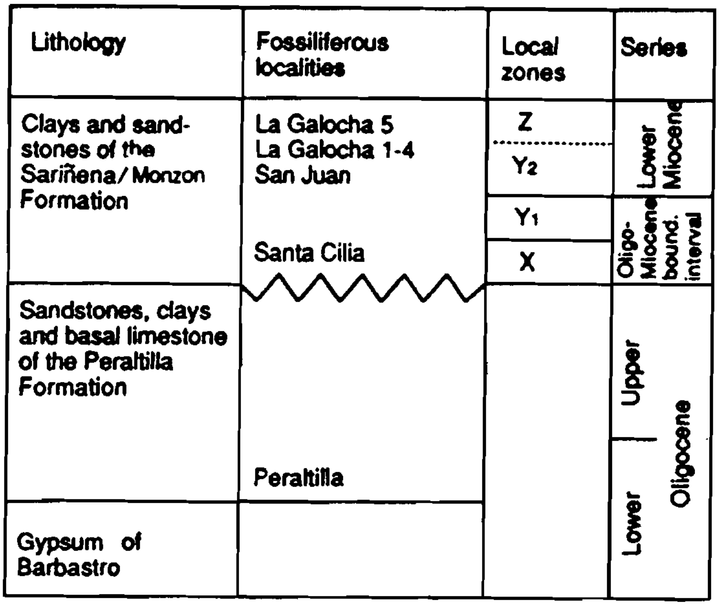


Figure 3. Correlation chart between the mammal fossil localities and the formations in the studied area, after Alvarez Sierra et al. (1990).

SEDIMENTOLOGICAL FRAMEWORK

General characteristics of the Monzon Formation

The Monzon Formation consists of fluvial channel sandstone bodies separated by floodbasin and overbank deposits. The channel bodies are sheet-like or elongate and may be simple or complex. There is evidence for vertical as well as for lateral accretion. Generally the sandstones are sheet-shaped. The thickness of these cycles varies from 1.5 m to 30 metres. The lateral extent and width/thickness ratios are high and often cannot be measured within the outcrop. The coarse-grained bases contain channel-fill cross-bedding with intraformational clasts, trough cross-bedding, and sometimes planar sets of large-scale cross-stratification. The fine-grained upper parts consist of yellowish sandstone beds with small-scale cross-lamination, with rootlets and locally strong bioturbation. Mudcracks indicate intermittently dry periods. The formation as a whole shows a coarsening-upwards trend in the sense of an increasing percentage of coarse-grained channel bodies.

Fluvial facies variability

The Monzon sandstone bodies are subdivided into braided stream facies (multistorey channel sandstone bodies), low-sinuosity facies, and high-sinuosity facies.

Braided stream facies

Generally, the braided stream facies is characterized by an erosive, though not very strongly incisive base. The channels vary in size. The basal parts of the channel fills consist of very coarse to coarse sand with fine gravel, showing cross-bedding and/or even lamination. Sometimes a single basal unit contains several sub-intervals, each being separated from the underlying unit by a basal scouring surface with very coarse sands. The sandstone bodies as a whole show a multistorey upbuilding, indicating the presence of multiple channel systems (cf. Wolman & Leopold, 1957; Rust, 1978). In this study, braided streams are regarded as being multi-channelled according to the definitions of Miall (1977) and Rust (1978). Schumm (1968) defined braided streams as single channel bedload rivers with relatively permanent and vegetated islands at low water stands.

This facies type is here explained as a system of channels, with mobile gravelly or sandy bars emerging at periods of low discharge (Rust, 1978; Ethridge, 1985). Collinson (1970) and Smith (1971) noted that sandy bars are not limited to braided channels, although they have been described most frequently in braided systems. The low percentage of fine-grained channel-fill sediment (claystones and siltstones) and the low number of storeys in the channel bodies found in the present study, suggest regular abandonment of channels in a multichannel system composed of individual low sinuous channels (cf. Leeder, 1978).

Part of the braided stream facies in the Monzon Formation is dominated by even-laminated sandstone sheets. This points to a relatively irregular discharge and possibly ephemeral streams (Schumm, 1977; Miall, 1985). Miall (1985) interpreted this facies, which he called sheet-braided deposits (Model 12 of Miall, 1985), as deposition from sheetfloods on a fluvial plain with a highly flashy discharge. The typically laminated sand sheets are interpreted as the product of flash floods, depositing sand under upper flow regime, plane-bed conditions (Miall, 1977, 1984; Rust, 1978; Tunbridge, 1981; Sneh, 1983). A distinction between dry and relatively small fans with ephemeral discharge and wet fans with more perennial and regular discharge was proposed by Schumm (1977).

The system described here, obviously represents the first type, considering the size of the fan as well as the ephemeral character of the deposits.

Another type of braided stream facies is characterized by multiple channels with gently to strongly erosive bases and cross-bedded to even-laminated sandstones. This conforms to Model 11 of Miall's (1985) sheet-braided deposits, which is interpreted as having been formed in a distal braidplain environment. Both major facies types are typical for arid regions where runoff and sedimentation are controlled by flash floods.

Low-sinuosity facies

This facies is made up of coarse- to medium-sized, poorly sorted sandstones, with a distinct fining-upwards grain-size trend. Towards the top the sandstones grade into fine-grained siltstone and mudstone layers. Erosion at the base is not very deep. Usually, a coarse-grained and conglomeratic lag with intraformational clasts is found on the erosional surface. Sedimentary structures in the coarse-grained basal part are mainly channel-fill cross-bedding, trough cross-bedding and sometimes planar cross-bedding. Because there is no indication of storey-wise upbuilding, this type of deposit is interpreted as having been formed in a straight channel (cf. Wolman & Leopold, 1957), or in a single low-sinuosity channel (cf. Rust, 1978). The sandstone bodies typically have a ribbon-like geometry.

High-sinuosity facies

The fluvial channel bodies of this facies show a fining upwards trend in grain size and consist of poorly to medium-well sorted, coarse to fine sandstones. Sandstones are trough cross-bedded with tangential foresets. Channel bodies as a rule have a sheet-like geometry and are multistorey. They are separated by thin, yellowish brown sandstone, siltstone and mudstone layers, interpreted as floodplain deposits. Occasionally these fine-grained floodplain layers show small-scale cross-lamination and even-lamination, but more often the physical structures have been destroyed by bioturbation, rootlets and other soil forming processes. Desiccation cracks and wave ripples have not been observed.

A strong indication for the meandering character of the deposits is the fact that they are organized in cycles, in which grain size and scale of sedimentary structures decrease gradually upwards to the fine-grained overbank and floodplain deposits. Low-angle lateral accretion surfaces are present, providing additional evidence for the high-sinuosity character of the channels (Allen, 1965a; Moody-Stuart, 1966). The laterally accreted bodies represent pointbars of meandering streams (cf. Visher, 1965; Allen, 1970; Miall, 1985). However, Nanson & Page (1983) stated that lateral accretion is not necessarily restricted to pointbar deposition, but may also occur on concave benches in confined meandering river systems, where an appreciable quantity of fine-grained material is deposited. Of course, this will occur only if channels are filled up rapidly upon gradual abandonment.

PALAEOCURRENT DIRECTIONS

Palaeocurrent directions have been measured in 133 cross-stratified beds. Since it was difficult to determine the exact orientation of channels due to outcrop conditions, measurements of channel axes proved to be unreliable and have not been used. Figure 4 shows rose diagrams of several sandstone bodies (D3, E4 and F). The rose diagram of sandstone body D3 shows a rather random pattern, whereas the diagrams of E4 and F show a more unimodal distribution. The rose diagrams of sandstone bodies E4 and F show palaeocurrents mainly towards the SSW.

DESCRIPTION OF FLUVIAL BODIES IN THE MONZON FORMATION

The architecture of the Monzon Formation has been studied on the eastern and western wall of the valley eroded by the present Rio Alcanadre (Fig. 5). A number of different sandstone bodies (23) has been identified and traced laterally as far as outcrop conditions allowed. Correlations are shown in Figure 5a. The character of the sandstone bodies (Fig. 5b) will be described below, starting from the southern part of the outcrop.

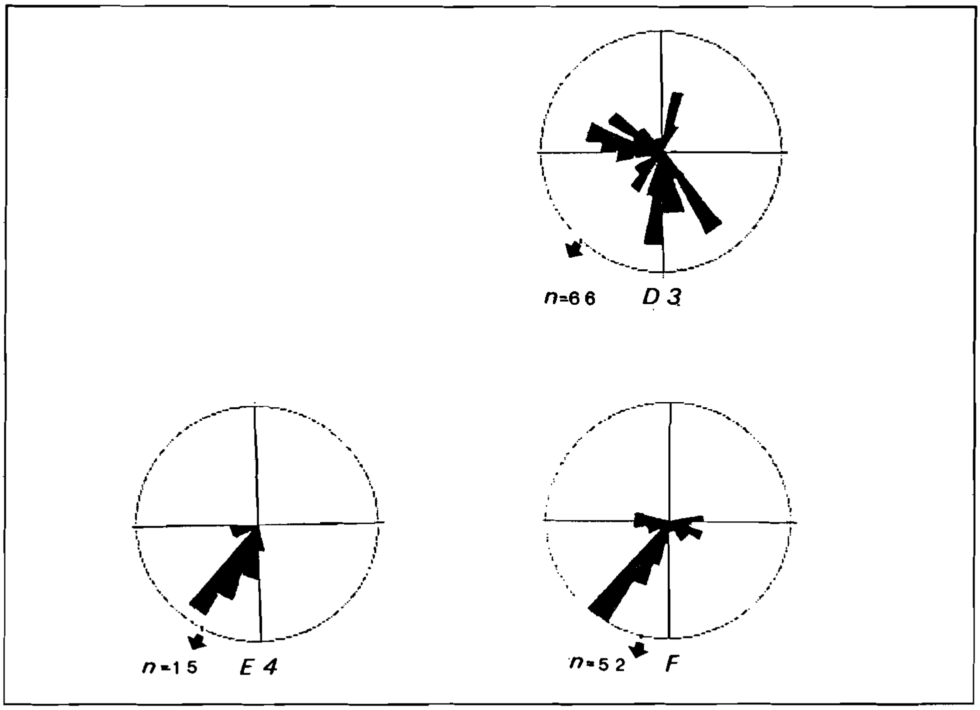
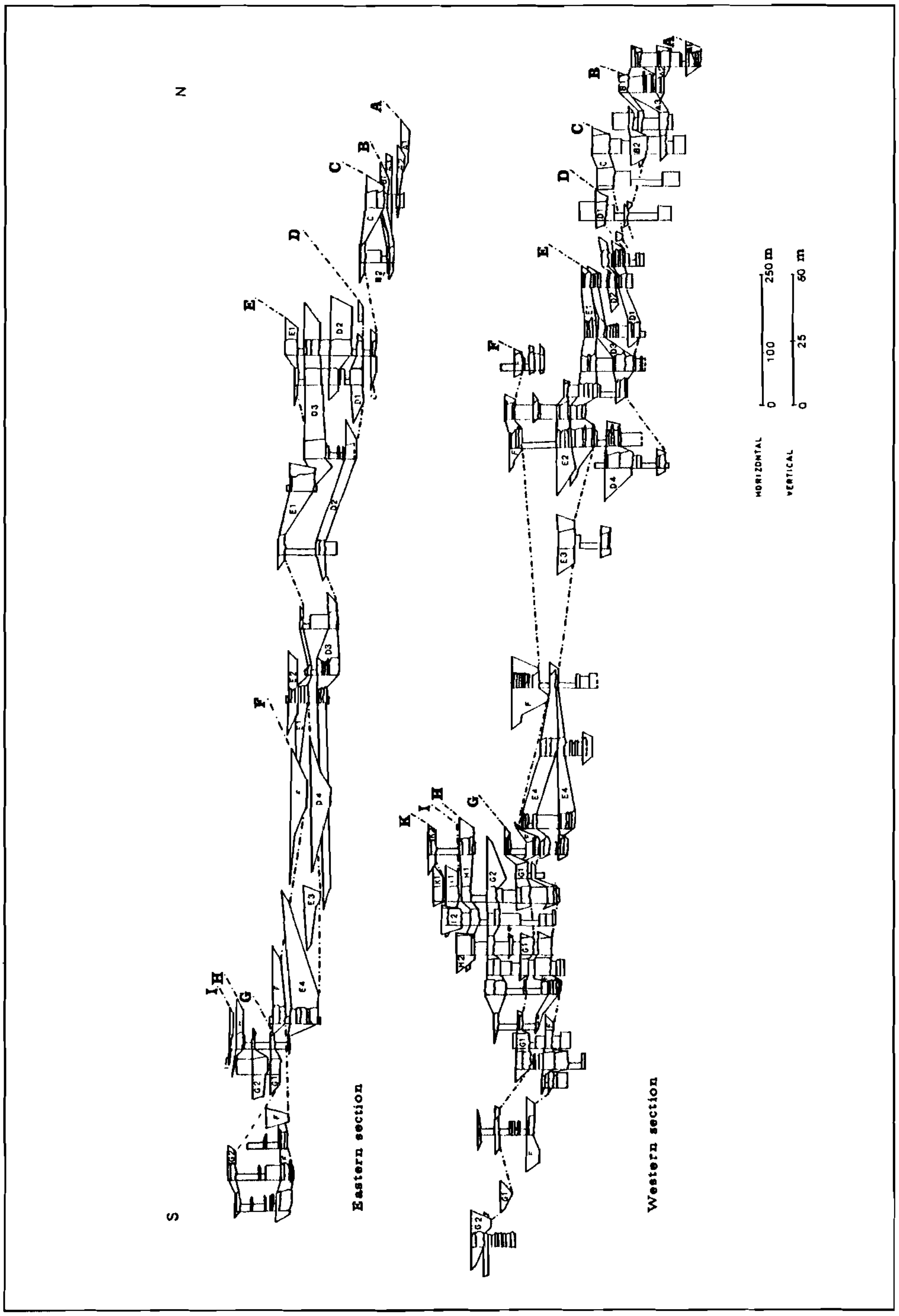


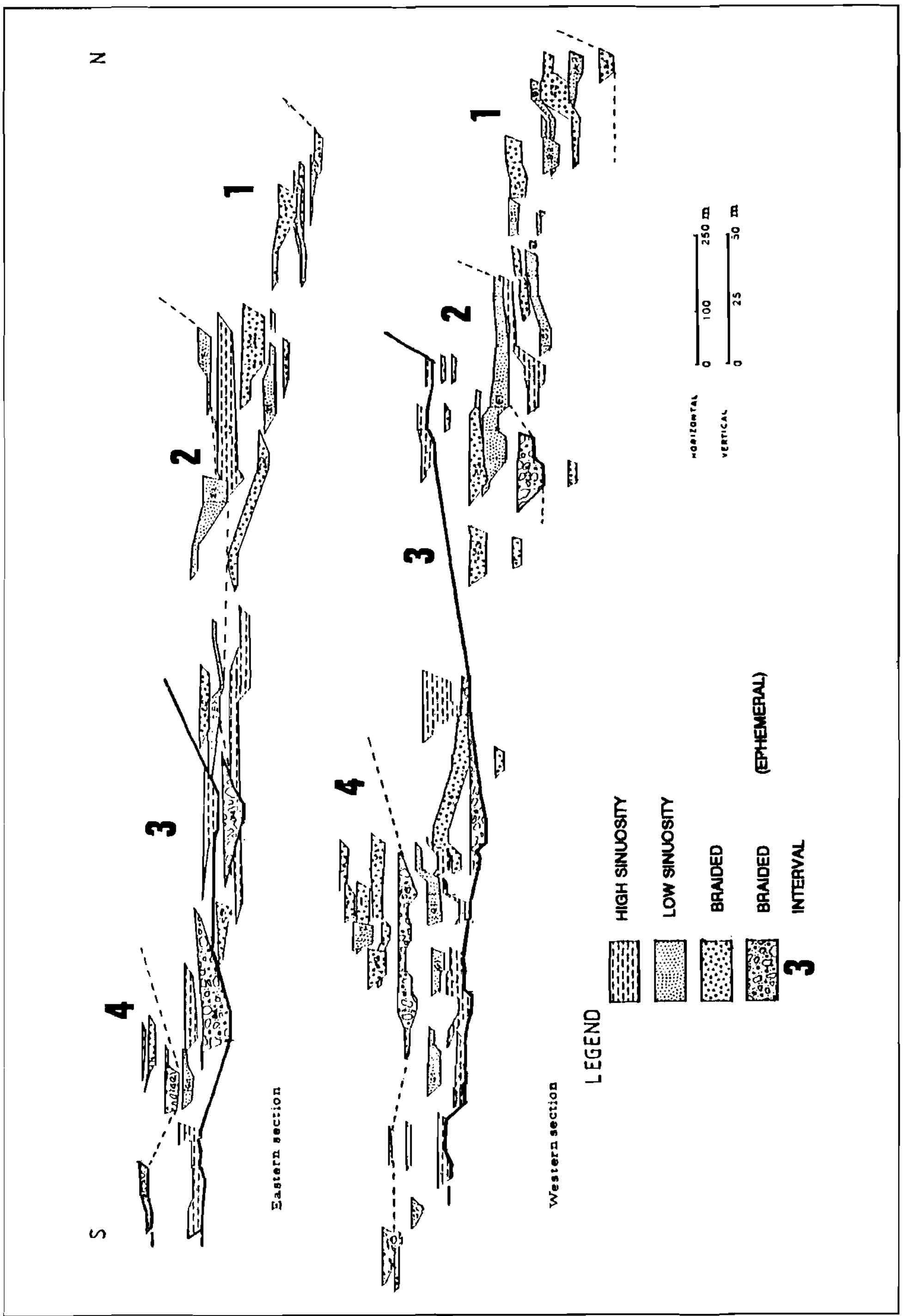
Figure 4. Current directions measured in sandstone bodies D3, E4 and F.

Interval 1 (channel bodies A1 to D3)

In the lowest part of the exposure, a braided channel body (body A1) has been identified (Fig. 6). The base is moderately erosive. The channel body has been deposited by several low-sinuosity channels with scouring and eroding bases covered by very coarse-sized sandstone to fine gravel conglomerate with low-angle cross-bedding. Towards the top this changes to massive medium-sized sandstone. The uppermost part of this sandstone body shows evidence of pedogenic processes and soil formation. The overbank deposits consist of fine to very fine, greyish, even-laminated sandstone and mudstone layers.

Figure 5. *See pages 37 and 38. (A) The architecture of the measured fluvial sequence in the Monzon area on the eastern and western wall of the valley cut by the present Rio Alcanadre. The figure shows the various sandstone bodies numbered A through K. (B) An interpretation of the fluvial sandstone bodies in terms of the style of the fluvial channels.*


N
S
Eastern section
Western section
HORIZONTAL
0 100 250 m
VERTICAL
0 25 50 m


S
N
Eastern section
Western section
1
2
3
4
LEGEND
HIGH SINUOSITY
LOW SINUOSITY
BRAIDED
BRAIDED (EPHEMERAL)
3 INTERVAL
HORIZONTAL 0 100 250 m
VERTICAL 0 25 50 m

The next coarse-grained unit (A2) (Fig. 6) is a single, low-sinuosity channel body which is moderately erosive at the base. The basal part consists of very coarse sandstone to fine gravel conglomerate, usually with flat internal erosional surfaces. The sandstone is moderately sorted and generally massive, with an upwards decrease in grain size. In some places planar cross-bedding and small-scale cross-lamination can be observed. The overlying floodplain deposits are medium-grained and contain soil horizons.

The overlying braided channel body (A3) shows a moderate to strong erosion surface overlain by fine gravel and very coarse-grained sandstone (Fig. 6 and Fig. 7). It shows well-developed low and high-angle trough cross-bedding, and small-scale cross-stratification. Sandstone body A3 seems to have been deposited by a river with a more continuous discharge. The floodbasin deposits are characterized by a succession of soil horizons with very fine sandstones at the base and the top, and overbank deposits of usually non-laminated mudstones in between.

A succession of low-sinuosity channel bodies (B1 and B2) follows on top of body A3. The basal parts of the channel deposits are very coarse to coarse-grained sandstones, and they show a fining-upwards grain-size trend. They are mainly massive and slightly erosive at the bases, sometimes with even lamination and cross-bedding. Floodplain deposits do not occur between the two sandstone bodies. The floodplain sediments above sandstone body B2 comprise unlaminated siltstones and mudstones.

Transport and deposition was subsequently taken over by a braided stream (C). The resulting sandstone body is moderately erosive at the base, and is usually dominated by low-angle cross-bedding and trough cross-bedding, sometimes with even-lamination. The floodplain deposits above sandstone body C are made of soil horizons and thin layers of mudstone.

The overlying, small low-sinuosity channel body (D.1) shows a slightly erosive base (Figs. 8 and 9). The grain size decreases upwards from coarse to fine sand. It is a simple channel body without storeys, and it has a generally massive appearance. Some low-angle cross-bedding occurs. A few thin intercalations of sandstones, siltstones and mudstones have been interpreted as overbank deposits. Sandstone body D2 (Figs. 8 and 9) is characterized by gentle erosion at the base, followed by coarse- to medium-grained sandstone with low-angle cross-bedding. It comprises several depositional storeys separated by erosional surfaces. This sandstone body has been interpreted as a multi-channel, low-sinuosity channel. The overlying floodplain deposits consist of thin

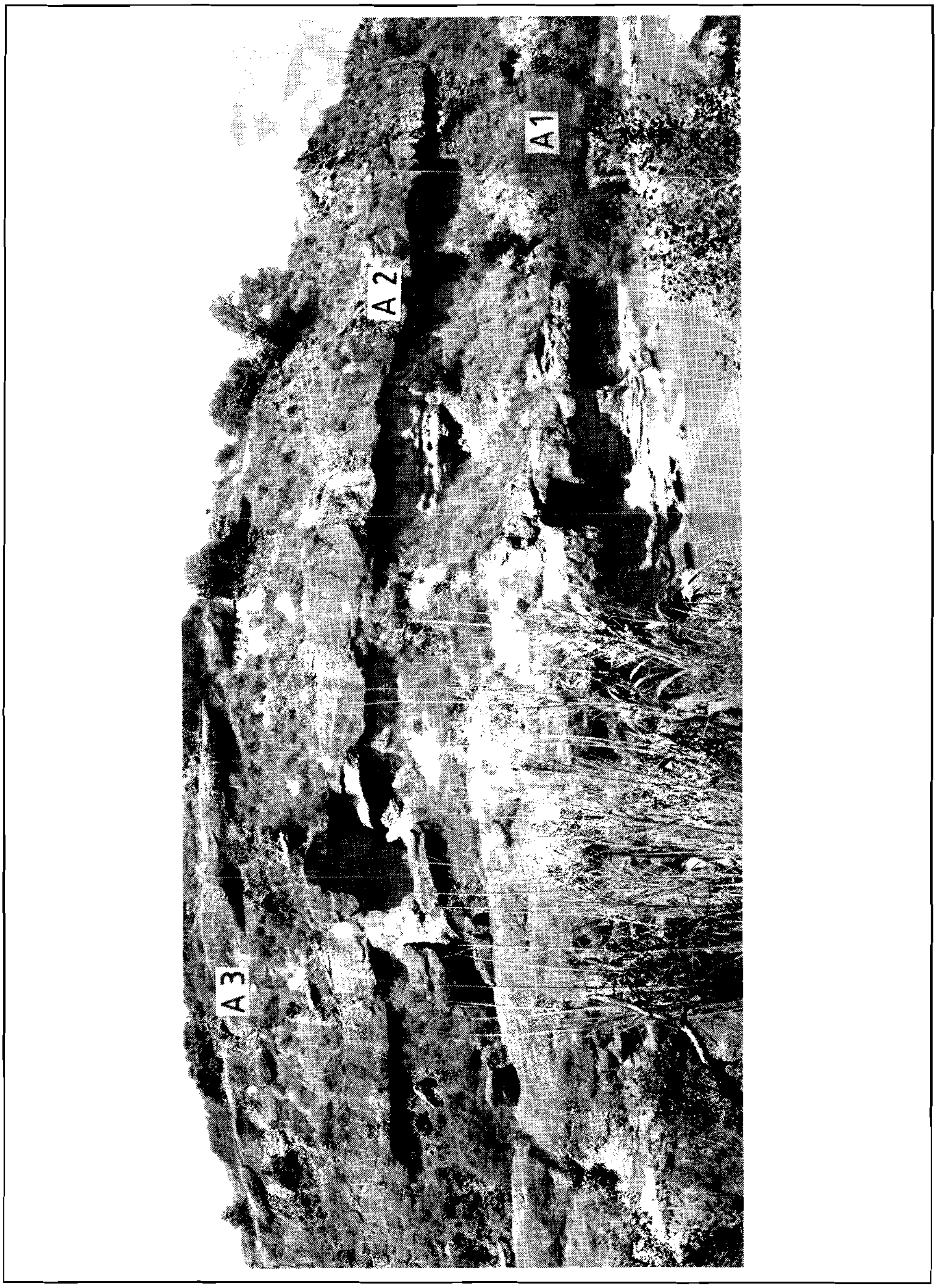


Figure 6. *Overview of the series in the lower part of the western valley wall showing sandstone bodies A1 to A3.*

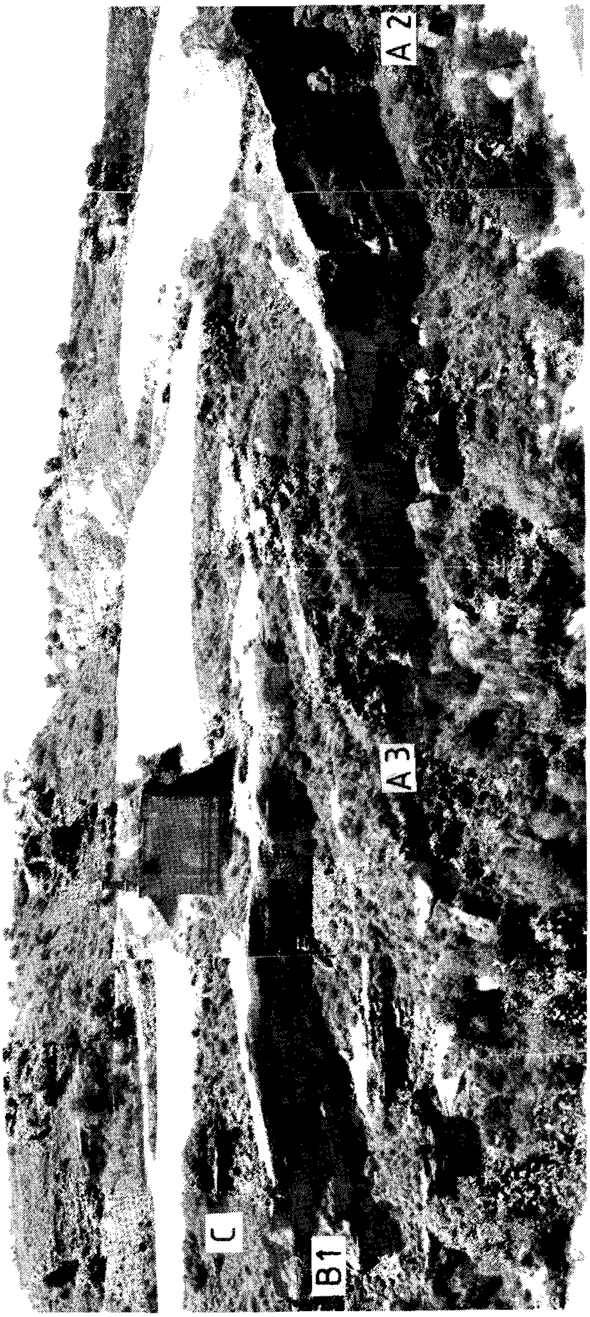


Figure 7. *Intervals A2 to C along the western side of the valley of the Alcanadre River.*

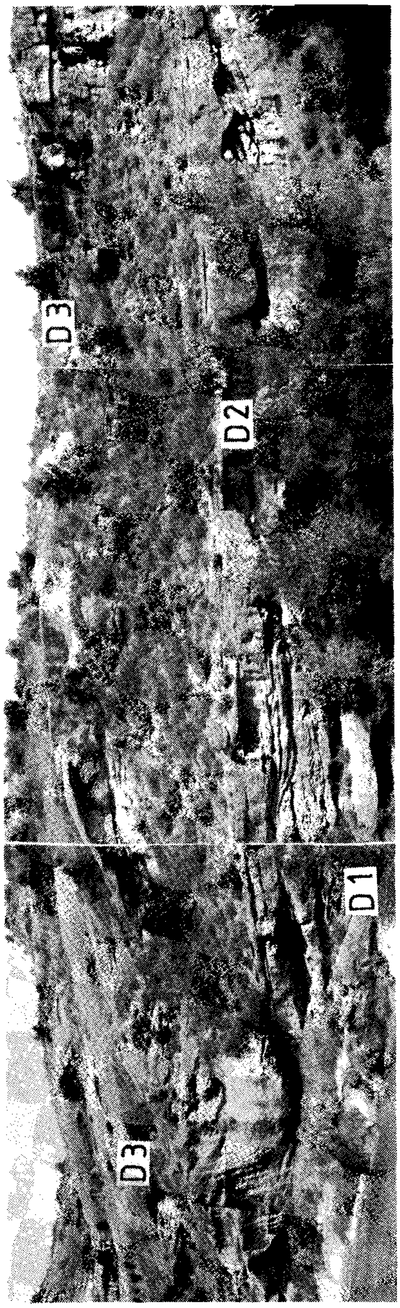


***Figure 8**. Overview of units D1 to D3 along the western side of the valley of the Alcanadre river.*

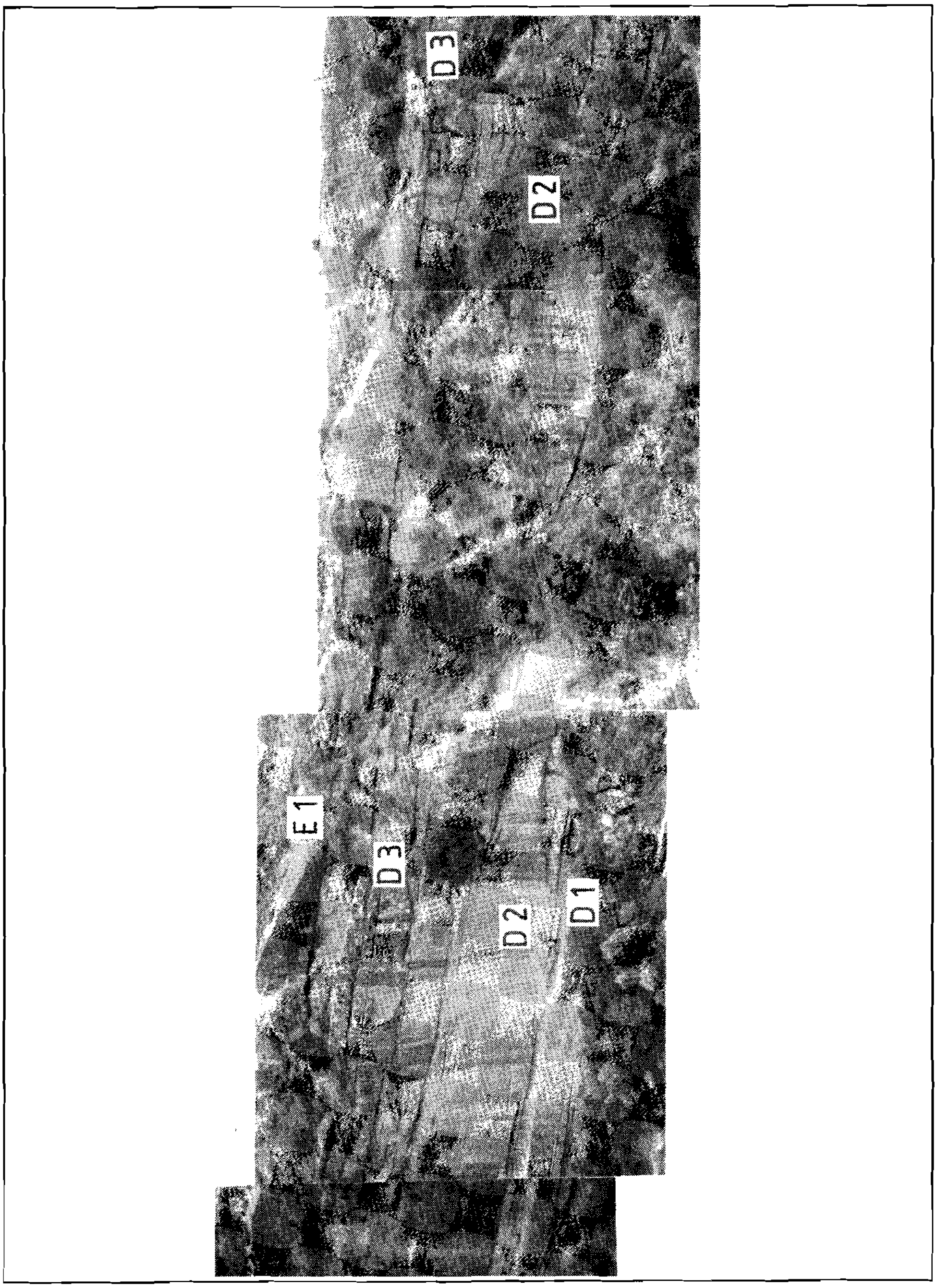


Figure 9. *Overview of units D1 to E1 along the eastern side of the valley of the Alcanadre River.*

layers of sandstone, siltstone and mudstone. Towards the top, a thick soil horizon is present. A high-sinuosity system (D3) was formed subsequently (Figs. 8 and 9). The main channel body shows moderate erosion at the base, followed by coarse to medium sandstone, fining up to very fine sandstone. The intercalated lateral accretion unit is characterized by trough and planar cross-bedding. The floodplain deposits of sandstone body D3 are dominated by soil formation.

Interval 2 (channel bodies D4 to E3)

The size and thickness of the channel bodies in this interval decrease gradually in the upward direction (Fig. 10). The lower boundary is characterized by proximal, ephemeral braided channel bodies (D4), followed by a low-sinuosity channel body (E1) and braided channel bodies (E2 and E3).

Sandstone body D4 has a shallow erosional base, followed by coarse to very coarse sandstone. It is even-laminated and contains gravel lenses. It is a multistorey channel body comprising several channels. It was probably deposited by an ephemeral braided stream, in which run-off was influenced by occasional flash floods. The overlying floodplain deposits are dominated by a succession of soil horizons.

The low-sinuosity channel body E1 begins with a shallow to moderately erosional surface, which is followed by low-angle to trough cross-bedded sandstone. The sandstone body is a single-storey channel body, as indicated by grain sizes and structures which remain similar throughout the unit. A succession of soil horizons and fine-grained overbank sediments with even-lamination and cross-lamination characterizes the overlying floodplain deposits. Two small braided stream channel bodies (E2 and E3) can be recognized on top of sandstone body E1 (Fig. 5). They are characterized by poorly sorted, coarse- to medium-grained multiple channels with shallow to moderately erosional bases. The units are usually massive, but sometimes even-lamination or cross-bedding can be recognized. The floodplain deposits between the sandstones contain soil horizons.

Interval 3 (channel bodies E4 to G1)

Interval 3 is composed of a succession of fluvial channel bodies which increase in size in the upward direction (Fig. 10). The interval starts with an ephemeral channel body (E4). It is followed by high-sinuosity channel bodies (F). The extensive high-sinuosity channel activity, indicates a period with more regular discharge. On top of these, low sinuous channel bodies occur (G1), whereas the dimensions of the channel bodies decrease. Thin overbank deposits mark the end of this depositional period.

Sandstone body E4 consists of two storeys, each of them showing a similar succession of facies. The unit begins with gently erosional, multiple channelled bases, which are covered by medium to coarse, poorly sorted, evenly laminated sand. The upper parts of the units show distinct internal erosional surfaces, even lamination, and low-angle cross-bedding. The storeys are separated and covered by fine-grained, evenly laminated overbank sands and mudstones. The succession of lithotypes indicates that deposition of the lower parts of the units was dominated by ephemeral flash-floods, whilst going up a more regular braided stream facies developed. The increased size of the vertically stacked sandstone units indicates an increasing rate of net aggradation for this part of the Monzon Formation.

Sandstone body F shows all characteristics of a high-sinuosity river system with a well-developed fining upward grain-size trend, lateral accretion surfaces, and crevasse splay deposits. Seven different meander loops could be distinguished along the studied outcrop. According to Miall's (1985) classification this would be a sandy, mixed-load meandering river (his Model 6). The pointbar accretion deposits usually have a simple geometry with some small-scaled bedforms superimposed. The sandstone body has a moderately erosional surface at the base. In the southern area the laterally accreted beds were eroded by a single thalweg channel, which belongs to the same system. The two channel bodies are separated by floodplain deposits consisting of fine to very-fine sand with even-lamination and cross-lamination.

Sandstone body G1 comprises four storeys. They have moderately to strongly eroding bases, fining-upward grain-size trends, and they are low-angle cross-bedded. They are thought to have been formed in relatively deep, low-sinuosity rivers with sandflats or shoals showing foresets, and isolated linguoid or transverse bars (cf. Miall, 1985). The floodplain deposits above this channel show thin overbank sands without bioturbation.

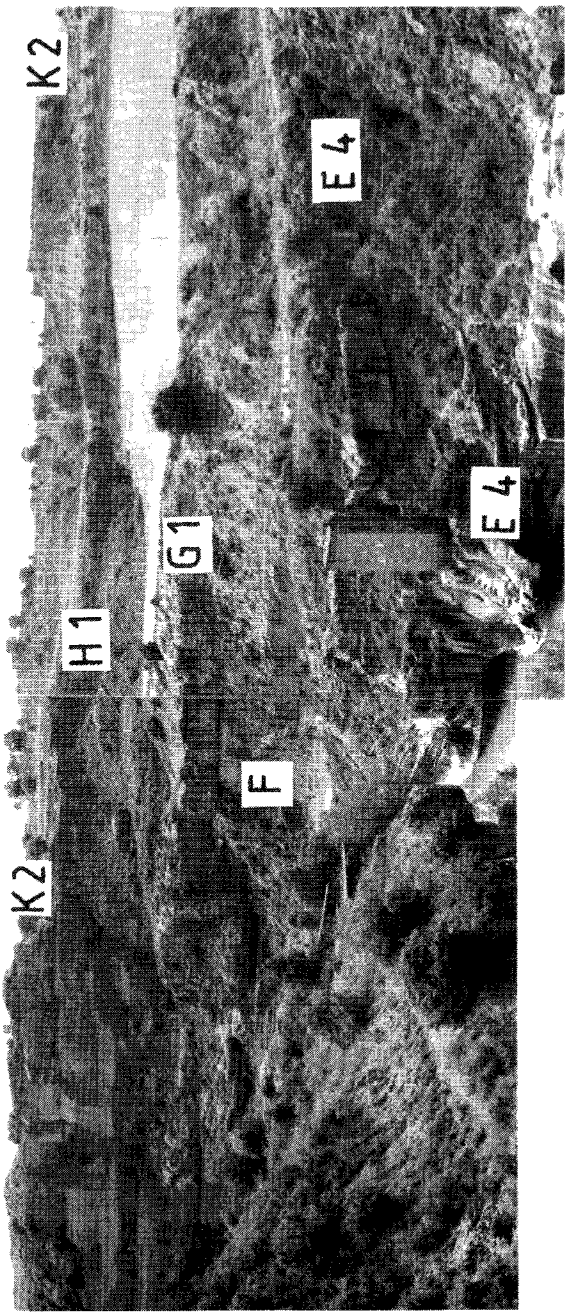


Figure 10. *Overview of units E4 to K2 along the western side of the valley of the Alcanadre River.*

Interval 4 (channel bodies G2 to K2)

Interval 4 is characterized by a gradual decrease in the size of the channel bodies. The lower part of the interval comprises laterally extensive ephemeral braided channel bodies, deposited during periods with relatively irregular discharge. On top of this, braided channel bodies (H1, H2 and I1) and a low sinuous channel body (I2) occur. The top of this interval is formed by sandstone bodies K1 and K2. On top of these channel bodies thin overbank deposits occur, which show evidence of pedogenic processes.

Sandstone body G2 has a shallow erosive base, followed by fine gravel and coarse to medium sandstone with parallel lamination and rare cross-bedding. Channel body G2 consists of three subunits, separated by thin, bioturbated overbank deposits. The depositional setting is supposed to represent the distal braidplain with ephemeral channel deposition (cf. Miall, 1985). Laterally stacked braided channel bodies are found in the overlying interval with bodies H1 and H2. They are characterized by shallow erosional bases, and composite bedforms with trough cross-bedding. These bedforms are probably bars. The floodbasin sediments display soil formation. Evidence for bioturbation has not been found. The next sandstone body (I1) represents a braided channel facies. It is covered by a low-sinuosity channel body (I2). No floodplain deposits occur between the two. The overlying floodplain deposits are dominated by paleosols and thin layers of overbank deposits. These deposits have not been bioturbated either. The uppermost braided channel bodies (K1 and K2) are characterized by moderate erosion at the base, even-lamination and rare cross-bedding. The floodbasin deposits contain soil and thin overbank deposits. On top of these bodies thin overbank deposits occur, showing evidence of pedogenic processes.

DISCUSSION: CAUSES OF THE FLUVIAL FACIES CHANGES

The development of a specific fluvial system is controlled by a number of partly interrelated parameters. These parameters include tectonic, climatic, hydraulic, and geomorphologic conditions (Flores, 1985). Processes as well as products of the fluvial system are influenced by these factors. In the past, climatic influences were generally considered to cause only minor overprints on a largely tectonically controlled system. In recent years it has become increasingly accepted, however, that climate is an important, and independent force in determining the facies characteristics of the system. Particularly the recognition of Milankovitch cyclicity as a driving force behind repetitive sequential patterns has helped in understanding the impact of climatic factors on sedimentation (e.g., Van Houten, 1964; Olsen, 1989; Perlmutter & Matthews, 1989).

The descriptions of the fluvial succession in the previous pages clearly show that in the course of time the depositional conditions have changed in a regular way. Low- and high-sinuosity and braided systems alternate on the scale of metres to tens of metres. An attractive, appropriate hypothesis is to ascribe such alternations in style of fluvial deposition to (regular) changes in the character and intensity of sediment supply. Three possible causes for such changes in sediment supply are regular changes in climate, regular tectonic pulses, and autocyclicity.

Climate

In an inland basin with internal drainage, it seems logical to suppose that climate is one of the major factors controlling both runoff as well as sediment availability. Therefore, the style of a fluvial body depends not only to the tectonic regime, but also to the climate. When direct influences of sea-level controlled base level changes are lacking, any regular, cyclic changes in fluvial style can possibly be related to cyclic changes in the climate. The studied area was, for most of the Tertiary, at a palaeolatitude of around 30 to 35°N. This is the latitude, which numerous studies have shown to be sensitive to climatic changes (see review by De Boer & Smith, 1994). Therefore, during the Miocene, the Ebro Basin could well have experienced the influence of slight climatic changes related to Milankovitch cycles and related shifts of the major climatic belts.

The geographic position of the major climate belts depends on the solar radiation and the topography of the Earth's surface. Periodic variations in the geometry of the Earth's orbit around the Sun, combined with its spinning axis, produce cyclic changes of climate. These cycles have recurrence intervals of approximately 20 ka, 40 ka, 100 ka and 400 ka and longer (Milankovitch cycles) (Berger, 1988). Provided that the sedimentary basin in question is located in a position which is sufficiently sensitive to such variations, the local climate will vary between predictable endmembers. For instance Perlmutter & Matthews (1989) have developed a detailed model for continental basins based on the assumption that the geographic position of a basin during a particular time interval dictates the climatic influence on basin stratigraphy. As yet, the stratigraphic time control in the studied succession is insufficient to decide if orbitally induced climate changes are the cause of the observed cyclicities.

Tectonics

The area is part of the north-central margin of the Ebro Basin, and is limited by the thrust fold of the Sierras Marginales (Sierra de Guara). This suggests that the sediment supply in this region is directly related to syn-sedimentary tectonic activity in the north and that subsidence was tectonically controlled.

From the relationships between the sandstone bodies, i.e., their vertical and lateral stacking pattern, it can be inferred that this region was subject to moderate subsidence. Blakey & Gubitosa (1984) showed that sandstone/mudstone ratios are important for the interpretation of depositional conditions. They related sandstone geometry and interconnectedness to frequency of avulsion, and subsidence rate. The stacking of the channels in the Monzon Formation is not very close and only shallow erosion during channel establishment has been observed. Following Blakey & Gubitosa (1984) this points to deposition in a moderately rapid subsiding basin. In addition, the stacking pattern of the channel bodies remains comparable throughout the whole succession. This suggests that subsidence and thus tectonic activity remained fairly stable during deposition of the succession. Therefore, the observed cyclic patterns in the style of the channel bodies and in the type of alluvial processes involved likely have not been the result of pulsating tectonic processes.

Autocyclicity

Fluvial systems which can develop under conditions of great stability of external parameters, such as tectonic relief and climate, may develop autonomic cyclic depositional patterns. This is due to lateral shifts of channel belts, which occur when these have built up to a sufficient height above the neighbouring floodplain (cf. Bridge & Leeder, 1979). However, any change in external parameters will either promote an early lateral shift or retard it, thus imposing external tuning of the system. The strength of such tuning depends on the strength of the external forcing factor.

Autocyclicity of alluvial systems implies that at any time all subenvironments are active within the system. With increasing influence of external forcing factors, certain subenvironments may increasingly be more dominantly represented in the fossil record on the cost of other ones. Also, with increasing influence of external forcing factors, the autocyclic aspect of the succession produced will diminish. It is obvious that the large lateral extension of the different lithological units in the above described succession does not favour an explanation in terms of autocyclic processes.

CONCLUSIONS

As outlined above, astronomically induced climate changes are an obvious mechanism for producing regular changes in precipitation and runoff, and the consequent character of fluvial systems such as the above described one in the Ebro Basin. The palaeolatitude of the area was at 30° to 35°N. From many parts of the sedimentary record (cf. De Boer & Smith, 1994), it has been shown that this latitude is indeed sensitive to astronomically induced climatic and oceanographic changes. However, the time control of the studied succession is obviously insufficient to decide whether the fluvial cycles indeed did result from orbitally forced changes of climate. On the other hand, if the argument is reversed, and if the cycles are assumed to be the product of such changes, this would imply that, at an average cycle thickness of about 10 m, the mean rate of deposition would have been of the order of 0.5 m/ka. For common foreland basins this is a slightly high, but fair value (cf. Allen & Homewood, 1986). In addition, it should be noted that the cycles described may include multiple cycles in which floodplain deposits in between coarse-grained intervals have either not been deposited, or have been eroded prior to deposition of the subsequent coarse-grained body. In case of erosion and amalgamation of cycles, the real number of cycles could have been higher. The average thickness of the cycles would then have been less, leading to a lower average sedimentation rate in case of the above hypothesis.

The present insights into the palaeoclimate of the Southern Pyrenees during the Oligocene-Miocene and the lack of sufficiently detailed stratigraphic control do not allow to preclude any of the three possible mechanisms in this stage. However, considering the fact that regular short-term tectonic changes and autocyclic control are unlikely causes for the observed cycles, it seems likely that indeed climate has been a major and independent control on fluvial style and architecture of the studied alluvial sequence.

CHAPTER 3

ALLUVIAL DEPOSITS OF THE OLIGOCENE-MIOCENE TÓRTOLA FAN, LORANCA BASIN, SPAIN

INTRODUCTION

The late Oligocene to early Miocene sedimentary alluvial succession of the Loranca Basin (Figs. 1 and 2) consists largely of coarse-grained channel bodies embedded in fine-grained floodplain deposits. Deposition occurred on two large alluvial fan systems. The southernmost fan, the Tórtola Fan, developed during a rise of base level, finally culminating in the deposition of gypsum with bioturbation structures and chert (Díaz-Molina et al., 1989). Different types of high- to low-sinuosity and braided channel fills are excellently exposed. Low and high-sinuosity palaeochannel fills show longitudinal and transverse bars. Moreover chute channels and meander loops of high-sinuosity channel fills are exposed in three dimensions. Braided channel fills show typical composite bedforms such as alluvial "island" sand flats.

The different low and high-sinuosity channel fill systems form stratal patterns showing a cyclic vertical alternation of coarse (fluvial) and fine-grained sediments. Lateral facies changes over short distances and the lateral discontinuity of the sandstone bodies demand a dense network of sections in order to produce a reliable reconstruction of the geometry of the sandstone bodies.

The objectives of this chapter are twofold:

- to reconstruct the facies development during the late Oligocene to early Miocene, and
- to interpret sedimentary cycles and related changes of sedimentary facies through time.

METHODS

All alluvial sandstone bodies have been mapped, and logs on 1:100 and 1:500 scales were made. The fine-grained floodplain deposits are brightly coloured and could be traced over large distances. Most sandstone bodies occur isolated within the floodplain deposits. Identification and correlation of the various sandstone complexes and successions (Figs.

3 and 4) are based on their occurrence within specific laterally extensive, brightly coloured fine-grained overbank complexes. Correlation was further enhanced by the occurrence of playa lake deposits and calcareous or gypsiferous paleosols, forming local key horizons. The stratigraphic succession and vertical and lateral facies relationships could be reconstructed mainly based on tracing of the floodplain fines. Palaeocurrents were measured and rock samples were studied petrographically. Attention was focused on the sandstone bodies below and above a marker bed, the Pozo marker bed, a laterally extensive lacustrine layer.

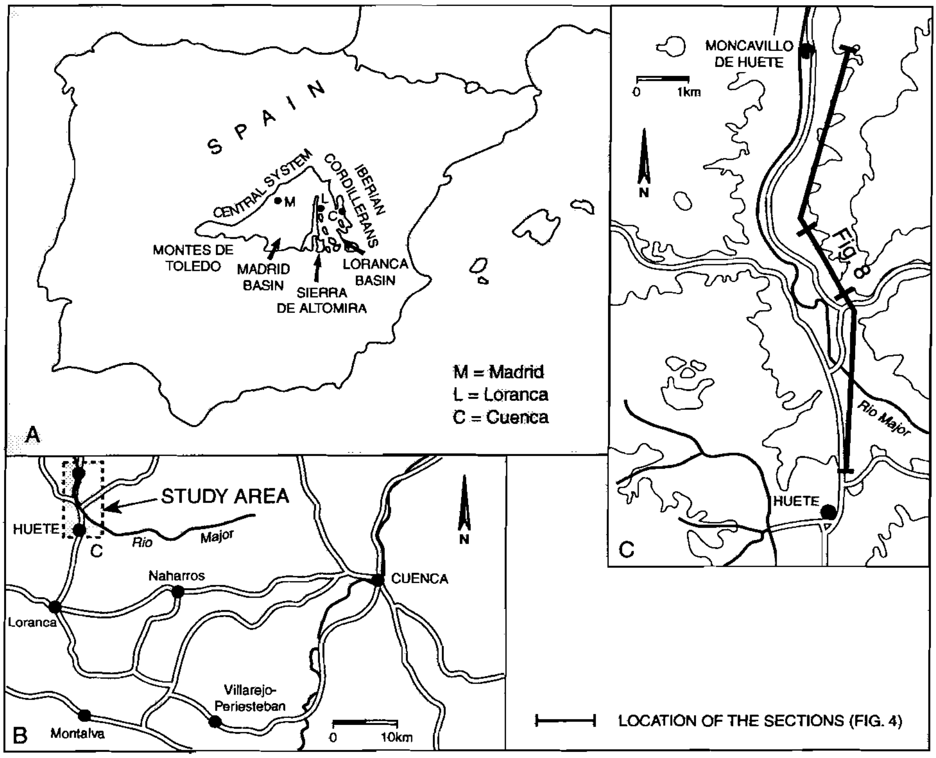


Figure 1. (A) Map of Spain with the location of the Loranca basin; B and C the studied area.

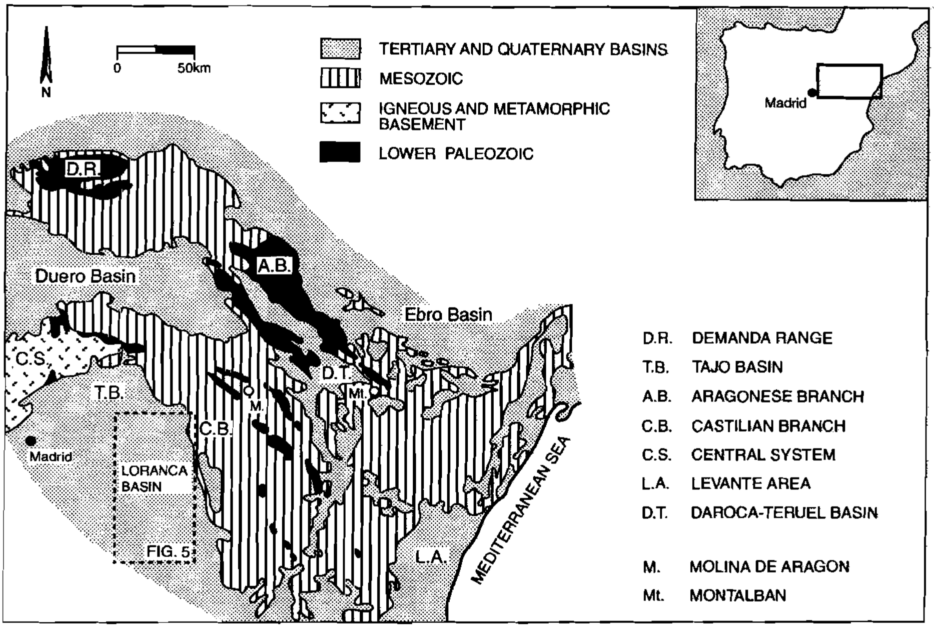


Figure 2. *Central Spain with the Loranca basin south of the Castilian Branch of the Iberian Range (after Sopeña et al., 1989).*

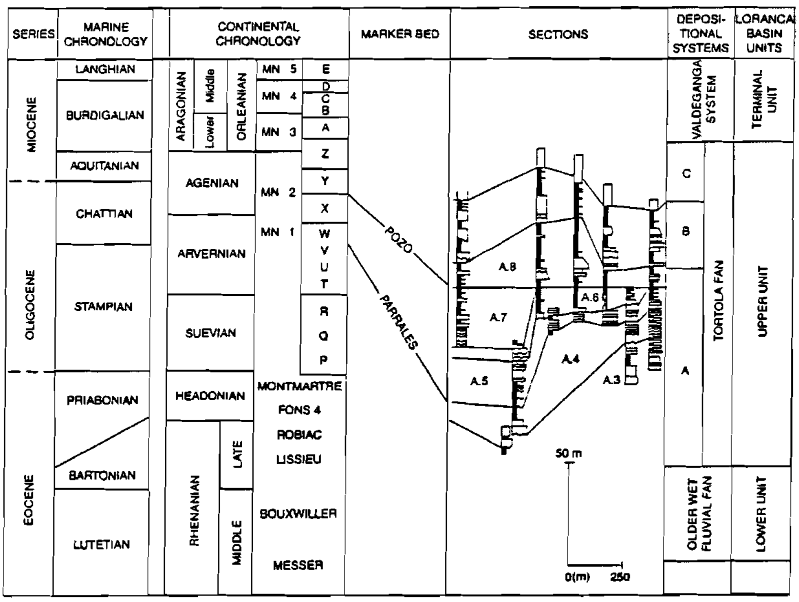


Figure 3. *Tentative correlation of the studied series with the continental stratigraphy. Paleontological data are from Daams et al. (1986) and Alvarez Sierra et al. (1987), after Díaz-Molina et al. (1989).*

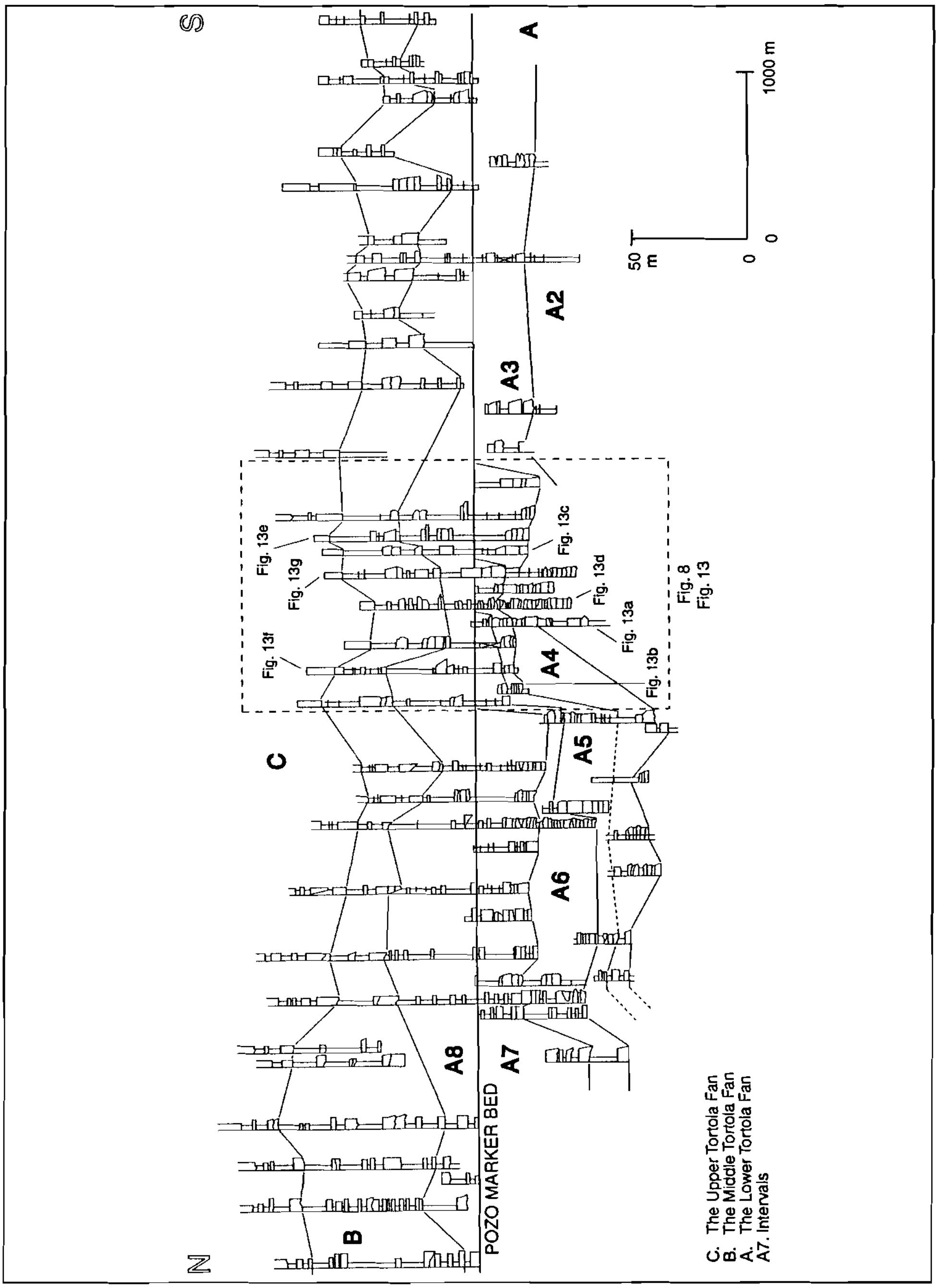


Figure 4. *Cross-section showing the correlations of fluvial intervals and the lacustrine Pozo marker bed. Measured stratigraphic sections show the coarse-grained fluvial bodies.*

GEOLOGICAL SETTING

The Loranca Basin (Fig. 1) is a marginal molasse-filled basin between the Iberian Range and the cratonic block of the Castilian Meseta (Díaz-Molina et al., 1989) (Figs. 2 and 5). The general stratigraphy of the Loranca Basin was described by Vilas-Minondo & Pérez-Gonzáles (1971), Melendez-Hevia (1971), Viallard (1973), Díaz-Molina (1974), Garcia-Abbad (1975), Díaz-Molina & Lopez-Martinez (1979) and Díaz-Molina et al. (1985) (Fig. 3).

Structural Setting

The structural setting of the Loranca Basin has been described in detail by Díaz-Molina (1989). The Iberian Range extends over about 400 km from the Cantabrian Mountains in the NW to the Mediterranean Sea in the SE. The average width is 150 km (Fig. 2). It can be divided into two branches, the Aragonese Branch and the Castilian Branch (Sopeña et al., 1989). These are separated by the Daroca and Teruel Tertiary Basins and join in the Levante area (Fig. 2). The Iberian range consists predominantly of Permian and Mesozoic sediments, resting unconformably on a basement of Hercynian metamorphic rocks. The Loranca Basin lies W of the Iberian Range (Fig. 2). Its main structural units have a N-S orientation, thus intersecting the NW-SE oriented main structural units of Central Spain (Figs. 2 and 5).

The Loranca Basin has an oval shape (Fig. 5). The western part of the basin is bounded by the thrust belt of the Sierra de Altomira (Fig. 5), where the Mesozoic cover has been thrusted from E to W over the basement and its immediate cover (Díaz-Molina, 1989). Decollement generally occurred within the Triassic Keuper deposits, and drilling results show that thrusts become younger towards the E (Díaz-Molina et al., 1989). The Loranca Basin depression formed during the Eocene to late Oligocene (Díaz-Molina et al., 1989) (Fig. 3). Slight synsedimentary folding occurred. These folds have a NNE-SSW alignment in the north, passing into a NNW-SSE orientation in the south (Fig. 5). Structural analysis allows the recognition of three stages of compression that occurred during the Tertiary (Díaz-Molina et al., 1989).

The valley of the Rio Major (Fig. 1), where the studied exposures are located, has a flat valley floor and gently dipping walls. The valley runs more or less parallel to the gently northward dipping fold axis of a weak anticline, dipping about 10° to either side (Fig. 5).

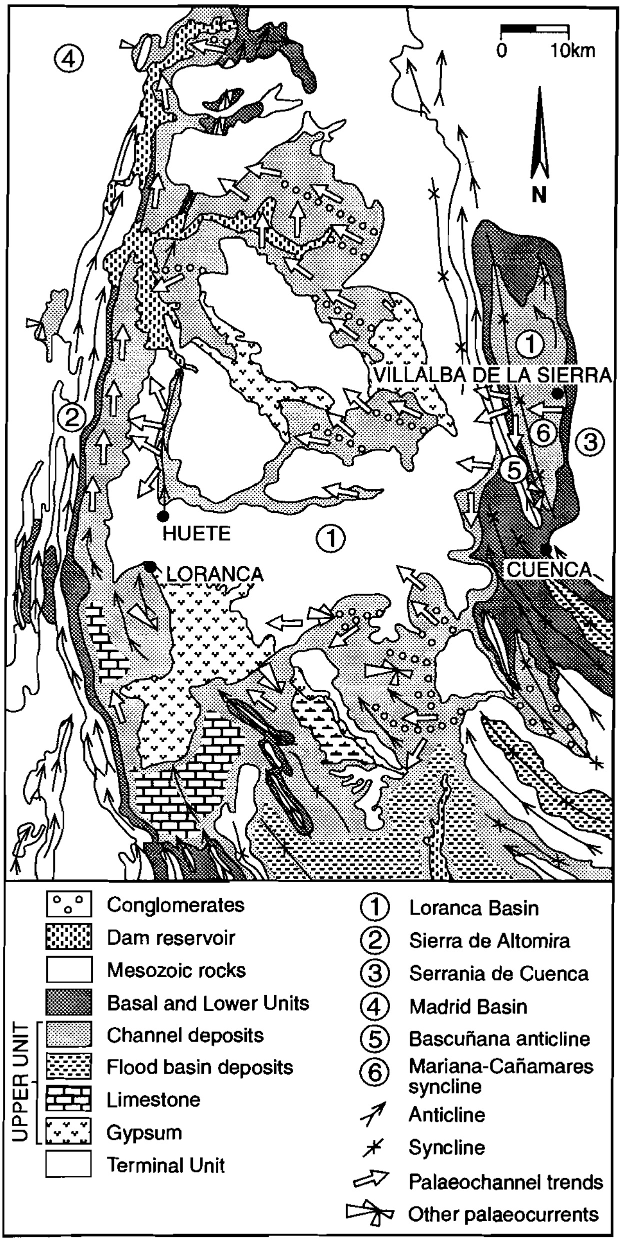


Figure 5. *Geological map of the Loranca basin. In the east, the Loranca basin is bound by the folded Mesozoic rocks of the Serrania de Cuenca. The western margin of the basin if formed by the thrust belt of the Sierra de Altomira (after Díaz-Molina et al., 1989). Arrows refer to general palaeocurrent directions.*

Stratigraphy

The stratigraphy of the Loranca Basin has been studied in detail by Melendez-Hevia (1971), Vilas-Minondo & Pérez-González (1971), Viallard (1973), Díaz-Molina (1974), Garcia-Abbad (1975), Díaz-Molina & Lopez-Martinez (1979) and Díaz-Molina et al. (1985). The last authors distinguished a Lower Unit, an Upper Unit, which is largely formed by the Tórtola Fan system and is subject of the present study, and a Terminal Unit (Figs. 3 and 4). Both the Lower and Upper Units consist of alluvial deposits. The Lower Unit accumulated when the basin was more extensive than at present. Deposition of the Lower Unit is considered to have been synchronous with the Eocene to Late Oligocene deformational event during which the Sierra de Altomira and the Loranca Basin were formed (Figs. 5 and 6) (Díaz-Molina et al., 1989).

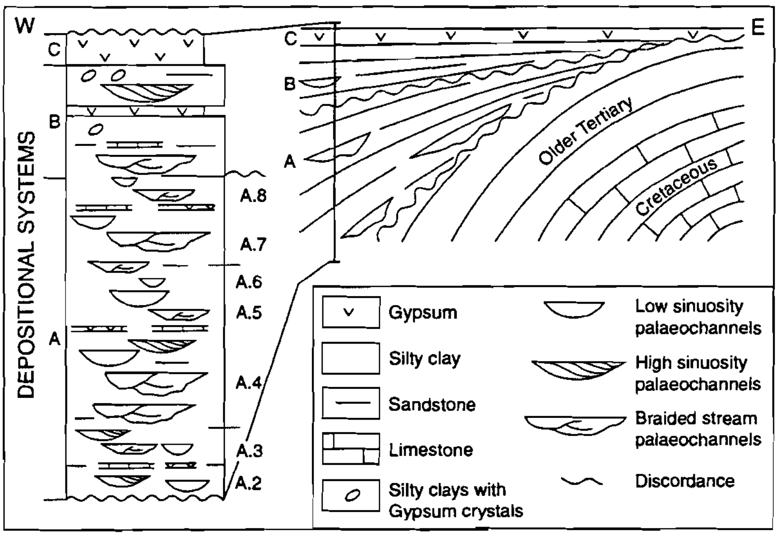


***Figure 6**. Stratigraphic subdivision and nature of the Tórtola Fan (forming part of the Upper Unit), subdivided in the Lower (A), the Middle (B) and the Upper Tórtola Fan (C). Syndepositional tectonic deformation of Mesozoic and older Tertiary basement is evident from the unconformities and the onlapping nature of the fan deposits shown in the schematic cross section (modified after Díaz-Molina et al., 1989).*

The Upper Unit, which is subject of this study, is mainly built by two large alluvial fan systems, the Tórtola Fan in the south and the Villalba de la Sierra Fan in the north. The maximum thickness of the Upper Unit is 800 m. Previously the Tórtola Fan was considered to have been the only supply system filling the Loranca Basin from the late Oligocene to the early Miocene (Díaz-Molina et al., 1985). More recently studies of Díaz-

Molina et al. (1989) indicate that another fan system, the Villalba de la Sierra Fan, was located north of the Tórtola Fan (Figs. 5 and 7).

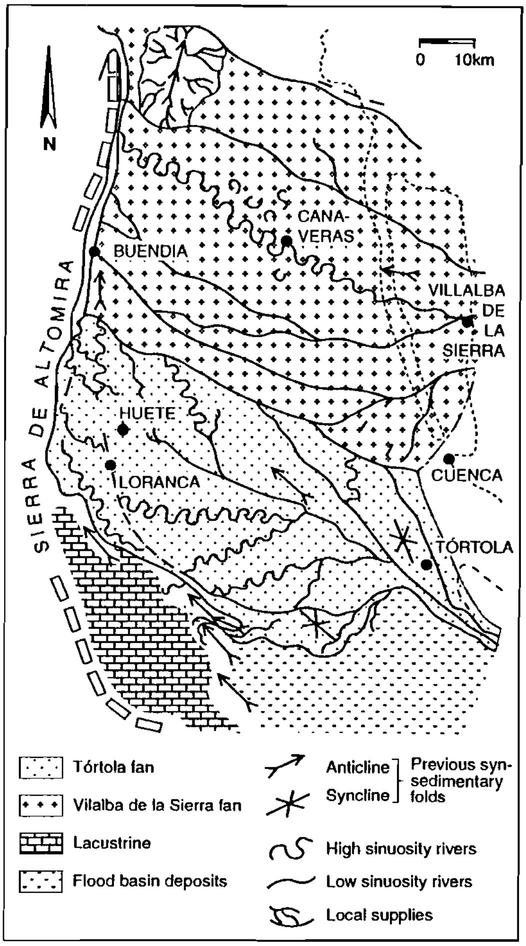


***Figure 7.** Reconstruction of the Tórtola and Villalba Sierra alluvial fans; after Díaz-Molina et al. (1989).*

The Tórtola Fan depositional system onlaps over the older tectonically deformed sediments in the basin (Fig. 6). Tectonic deformation occurred during deposition of this unit, as shown by progressive discordances on the flanks of anticlinal folds and along the basin margin (Díaz-Molina et al., 1989) (Figs. 5 and 6). Based on differences in clastic composition, the alluvial sedimentary succession of the Tórtola Fan can be subdivided into three depositional systems (Fig. 3), the lower Tórtola Fan (A), the middle Tórtola Fan (B) and the upper Tórtola Fan (C).

Earlier studies (Díaz-Molina et al., 1985) show variations in the nature of the type of river deposits depending on the location in the fan body. In a down-stream direction the Tórtola Fan system sediments show a general change from braided to low- and high (meandering)-sinuosity river types. However, locally the different types of fluvial channels alternate.

Palaeocurrent analysis and measurements of channel orientation revealed a SE-NW orientation alignment of the palaeochannels (Figs. 7 and 15). The main detrital components are extrabasinal (derived from sources outside the basin), non-carbonate grains including monocrystalline quartz, K-feldspar and microcline. A small number of extrabasinal carbonate grains occurs (micritic and sparry limestones and dolostones). Intrabasinal carbonate grains are mainly derived from paleosols. The modal composition of extrabasinal detrital grains reflects a high proportion of recycled sedimentary rocks derived from the Mesozoic succession of the Iberian Range. Apart from clastic gypsum, the average sandstone composition of the Tórtola Fan deposits remains similar throughout the sequence, pointing to one source area surrounding the basin in the east (Díaz-Molina et al., 1989). The content of intrabasinal carbonate grains is variable due to local reworking of paleosols.

Díaz-Molina et al. (1989) stated that during deposition of the lower Tórtola Fan (A) (Fig. 3) the fan system showed it greatest activity. The depositional system A consists mainly of channelized conglomerate and sandstone bodies along with layers of siltstone and silty claystone, thin sandstone and siltstone sheets and fresh water limestone layers. In the southern part of the basin silty clay, limestone, and gypsum form the dominant lithology. This depositional system shows a slight upward decrease in palaeochannel dimensions and in abundance of channels (Díaz-Molina et al., 1989). The sediments have been transported by various kinds of mass-transport mechanisms and ephemeral streams.

Differences between the Middle Tórtola Fan deposits (B) and the Lower Tórtola Fan deposits (A) have been related to a decrease in tectonic activity (Díaz-Molina et al., 1989). The Upper Tórtola Fan deposits are characteristically dominated by clastic gypsum, which was supplied from weathering gypsiferous late Cretaceous and Triassic rocks (Díaz-Molina et al., 1989).

The Upper Unit is covered by the Terminal Unit consisting of fluvial, alluvial and lacustrine gypsum and fine-grained clastic deposits extending throughout the basin (Díaz-Molina et al., 1989). The lithology indicates that during the Early Miocene the basin was not drained and that all water supplied to the basin evaporated.

Palaeoclimate

The palaeoclimatological conditions were reconstructed on the base of faunal (micromammal) evidence found in the Upper Unit in the Loranca Basin. A relatively warm and humid climate was proposed with a progressive trend to more arid conditions towards the top of the Upper Unit (Daams & Van der Meulen, 1984; Díaz-Molina et al., 1985; Lacomba & Morales, 1987; Lacomba, 1988; Díaz-Molina et al., 1989) (Fig. 3). Drier conditions are reflected by the increasing number of fine-grained floodplain deposits with pedogenic gypsum crystals in the upper part of the studied succession (Figs. 8, 9 and 10).

The Tórtola Fan

The Lower Tórtola Fan (A)

The Lower Tórtola Fan (A) comprises alluvial channel sandstone bodies embedded in fine-grained floodplain deposits. Deposition of coarse-grained sediments mainly resulted from deposition in channelized streams. The Lower Tórtola Fan succession is divided into several intervals (A1 to A8). Intervals A2 to A8 are the subject of this study (Figs. 4 and 6). Each of these intervals consists of a succession of genetically related fine-grained floodplain deposits with paleosols and alluvial channel sandstone bodies. The successive intervals (A2 to A8) are separated locally by lacustrine intervals or soil intervals (Figs. 8a and b). The sandstone bodies in this system show a simple vertical stacking without severe erosion of older sandstone bodies. The bodies comprise single and multistorey channel bodies. In the latter case, alluvial deposition occurred during several, successive phases of river discharge, and subunits usually are separated by an erosional surface.

Channel sandstones, fine to very coarse sand, consist of moderately to well-sorted, well-rounded quartz and feldspar grains and rock fragments, cemented by calcite or gypsum. Usually the lower part of the sandstone bodies is yellow to white or light-grey.

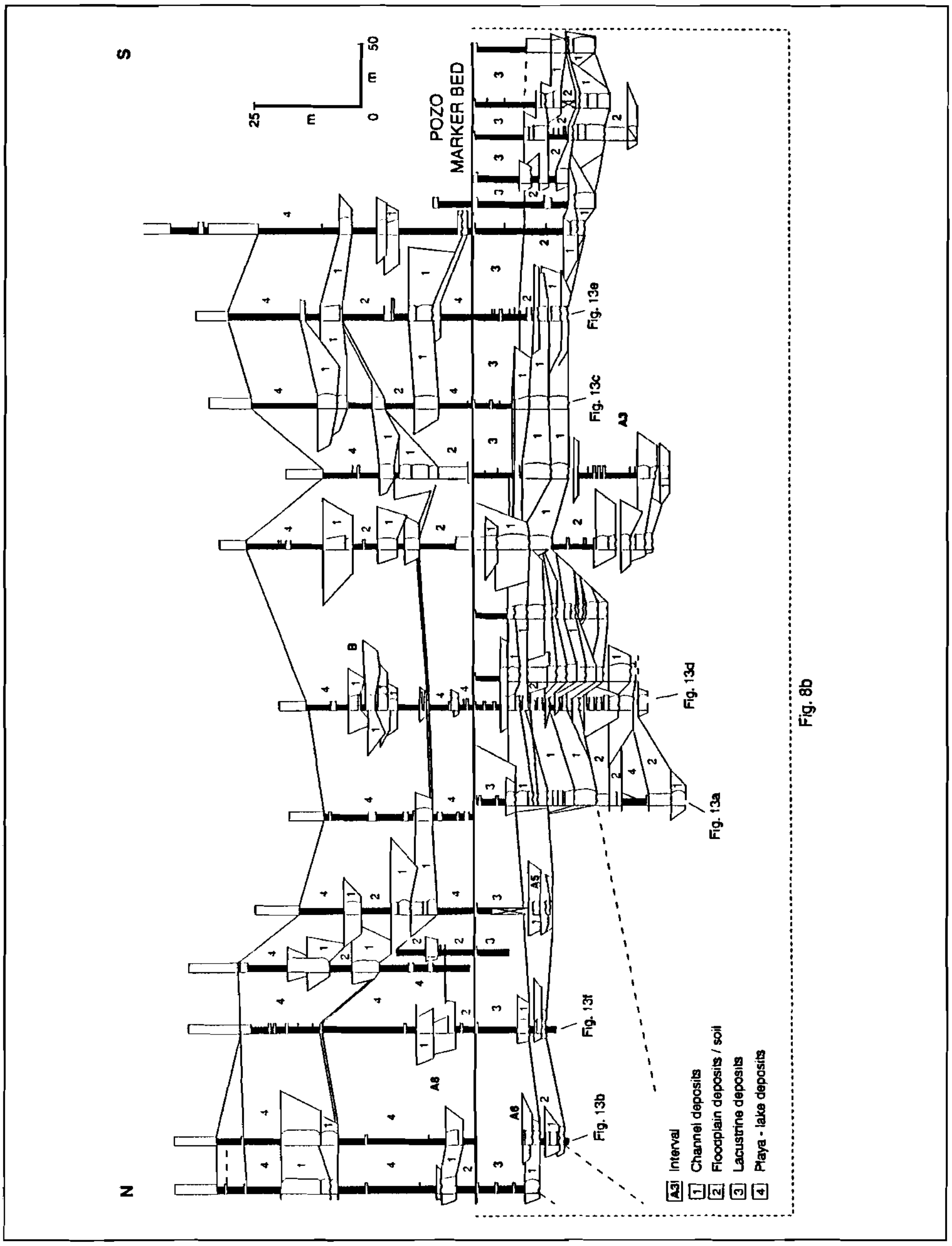


Figure 8 A. *Correlation of logs of measured stratigraphic sections in the central part of the research area. Logs show the facies and the relationships between floodplain, fluvial and lake systems. Location of cross section is given in Figure 4.*

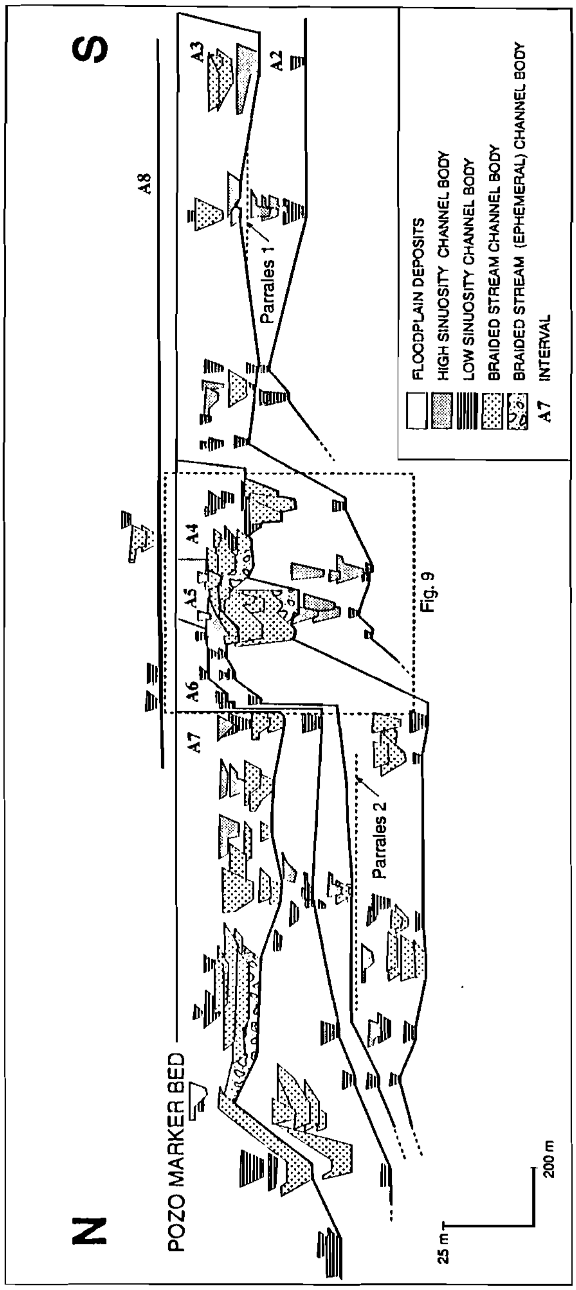


Figure 8 B*. Enlargement of part of Figure 8a. Stratigraphic cross section through the central outcrop belt.*

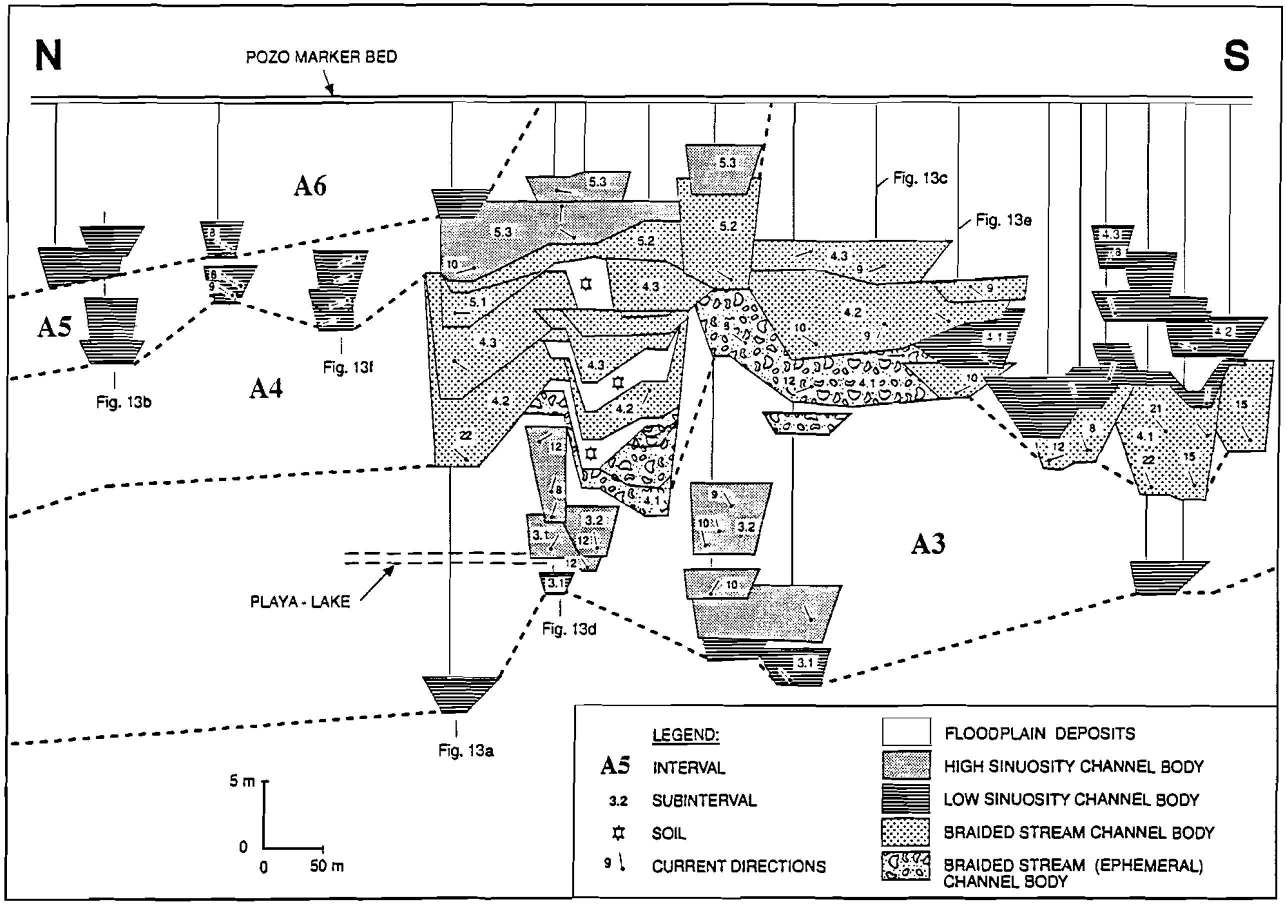


Figure 9. *Enlargement of part of Figure 8b. Stratigraphic cross section with interpretation of the type of fluvial channel bodies.*

Interval A2 is only exposed in the south. Interval A3 is exposed in the south as well as in the central part of the area (Fig. 9). Intervals A2 and A3 are locally separated by a lacustrine interval, called Parrales 1 (Fig. 8b), which has an Upper Oligocene age (Daams & Van der Meulen, 1984).

Intervals A4 to A6 are exposed in the central and northern part of the area. Interval A4 is characterized by densely stacked, large sandstone bodies separated by soil horizons. The boundary between intervals A4 and A5 is locally formed by a lacustrine interval, called Parrales 2 (Fig. 8b) (Díaz-Molina et al., 1989). Intervals A5 and A6 are separated by soil horizons. The sandstone bodies of interval A6 show a decrease in thickness in the upward direction.

The channel dimensions (depth and width) of the sandstone bodies in intervals A4 and A7 (Figs. 4, 8a and 8b) exceed those of other intervals. Interval A7 is only exposed in the north. Interval A7 and A8 are separated by an erosional surface covered by a continuous lacustrine key layer (Pozo marker bed), which can be traced as a major marker bed throughout the whole area (Figs. 4, 8 and 9). This lacustrine layer nevertheless shows a large lateral variation in lithology comprising limestones, siltstones, marls and gypsum. Interval A8 covers the whole area. The upper boundary of interval A8 again is an erosional surface which forms an unconformity (Fig. 6). On top of this erosional surface, for the first time sandstone bodies with detrital gypsum appear, which are typical for the Middle Tórtola Fan (B).

During development of the sequence of the Lower Tórtola Fan (A), the locus of alluvial channel deposition shifted towards the north. This shift, as well as the lateral extensive erosional surfaces bounding interval A8 probably can be attributed to tectonic activity.

The Middle Tórtola Fan (B)

The Middle Tórtola Fan (B) also consists of sandstone bodies embedded in fine-grained deposits. Typical for these sandstones is the presence of detrital gypsum, resulting in a grey to whitish appearance of the sandstones. Apart from this detrital gypsum, the mineralogy of the bulk of the grains is similar to that in the Lower Tórtola Fan (A). A second difference between the Lower and Middle Tórtola Fan deposits is the geometry of the sandstone bodies. The sandstone bodies of the Middle Tórtola Fan (B) usually comprise a single depositional storey, whereas in the Lower Tórtola Fan (A) multistorey sandstone bodies are common.

The thicknesses of the single storey channel sandstone bodies in the Middle Tórtola Fan range from 1 to 8 m, and the maximum width is 10 to 20 m. The grain size of the sandstone bodies is very fine to very coarse sand. Occasionally, contorted bedding occurs.

The fine-grained floodplain deposits are comparable to those of the Lower Tórtola Fan. At some intervals gypsum conglomerates are present. Such layers have a thickness of about 0.5 m and consist of gypsum crystals in a matrix of clayey marl.

The upper boundary is formed by sandstones composed of detrital gypsum of the Upper Tórtola Fan (C).

The Upper Tórtola Fan (C)

This depositional system consists of sandstone bodies composed of detrital gypsum. Occasionally, sedimentary structures such as cross-stratification have been preserved in spite of the recrystallization of much of the original clastic gypsum. Two types of clastic gypsum occur. The first type is found at the base of this depositional system and consists of massive light-brown to reddish gypsum beds with a thickness between 1.5 m and 8 m and a lateral extension of several hundreds of metres. The second type of gypsum consists of lobate massive gypsum bodies embedded in red-brown marls. This gypsum has a white colour and becomes grey upon weathering. The lobes can reach a thickness of more than 7 m and extend several hundreds of metres laterally.

Sandstone petrography of the Lower Tórtola Fan (A)

A joint study of 40 thin sections by the author and by de Vries (1990) confirms earlier work of Garcia Palacios (1974, cit. Díaz-Molina et al., 1989). The main clastic components of the sandstones are monocrystalline quartz (50-70%), feldspar (5-10%), carbonate rock fragments (25-50%) and some chert grains. The roundness and sphericity of the individual grains depend on the mineralogy. The quartz grains have a low sphericity and are very angular to subangular. Feldspars have a low to medium sphericity and are angular to well rounded. The feldspars have commonly been weathered and seritized. The carbonate rock fragments consist of mudstone grains and (extrabasinal) wackestone grains containing fossils (foraminifera) and fossil fragments (corals and pelecypods). These carbonate components are sub- to well-rounded and have a medium sphericity. Sometimes mica flakes are present (muscovite and biotite).

Generally the sandstones contain 5 to 10% micrite pseudo-matrix. This pseudo-matrix was generated during mechanical compaction and squeezing or destruction of the generally friable carbonate mudstone grains. Other features indicating compaction include penetration of quartz grains into softer grains and broken carbonate grains. Hardly any calcite-carbonate cement is present. Sometimes micrite matrix and carbonate grains have been recrystallized into sparite, and feldspars have been replaced by clay minerals.

From the thin section analysis it appears that the sandstones are micritic lithic arenites. The low sphericity and often poor roundness of the grains (especially quartz and feldspar) indicate little abrasion and weathering during transport from the source area to the sedimentary basin.

THE LOWER TÓRTOLA FAN SEDIMENTARY FACIES

The sediments in this area are fluvial and lake systems. The fluvial deposits include channel sandstones, which will be described in detail separately, and fine-grained floodplain deposits. The lake deposits include lacustrine and small playa-lake deposits. Representative sedimentological logs are shown in Figure 10.

Fine-grained deposits

Floodplain and crevasse deposits

The fine-grained intervals are dominated by ochre to brown, red-brown, orange-brownish, and grey-yellow to orange greyish laminated marls and clayey or silty marls. Frequently they have a massive appearance. Thicknesses of fine-grained intervals vary from a few centimetres in between the channel bodies, where obviously part of these fine-grained deposits have been eroded, up to several tens of metres laterally of the coarse-grained channel bodies. The channel bodies are embedded within the fine-grained deposits, which are interpreted as having been deposited on the floodplain relatively distal from the main channels.

Thin siltstone and sandstone layers are frequently intercalated in the mudstone intervals. The siltstone layers have a thickness of 0.1 to 0.8 m and show even lamination and cross stratification. Fine to very coarse-grained sandstone beds, with a thickness of 2 to 70 cm, generally show even lamination or sometimes cross-stratification. These intervals are interpreted as crevasse splay and sheetflood deposits.

The above fine-grained floodplain overbank deposits and intercalated sheetflood and crevasse silt- to sandstone deposits have reddish colours and commonly show (pedogenic) colour mottling. Vertically, colours show slight variations in hue. These colour variations are probably due to fluctuations in climatological conditions, variations in the groundwater level during and shortly after deposition, or variations in the rate of vertical accretion of the floodplain. The vertical accretion rate may be related to tectonic activity or simply to periodic avulsion of main alluvial channels.

Reddish colours indicate oxidizing conditions related to low groundwater levels. The breakdown of minerals which contain iron such as biotite and hornblende leads to the formation of clay minerals and immature iron-oxides under the influence of rain water during flooding. Buurman (1980) states that yellow colours indicate the presence of hydrated iron-oxides. This hydrated character suggests wetter climatic conditions than in the case of the red muds. Locally greyish colour mottles are observed. Iron and manganese were separated by an alternation of reduction and oxidation processes in the bottom triggered by groundwater fluctuations (cf. Buurman, 1980). The grey colours indicate a relatively high concentration of organic matter and/or pyrite. Organic matter can be preserved under reducing conditions as a result of a temporary high groundwater level or a stagnation of the groundwater flow.

Playa-lake deposits

This facies is characterized by black to grey mudstones. The beds of these badly laminated deposits have a thickness of 0.3 to 12 m. Locally, this facies consists of siltstone and sandstone beds with thicknesses ranging from 0.1 to 0.8 m. Also vertical and lateral alternations of silt, gypsum and limestone layers with thicknesses between 0.05 to 0.8 m are found. These sediments are interpreted to have formed in small playa lakes in between the floodbasin (Fig. 9). These playa lakes probably had a short life time. The lake deposits are intercalated within the fluvial channels deposits.

Lacustrine deposits

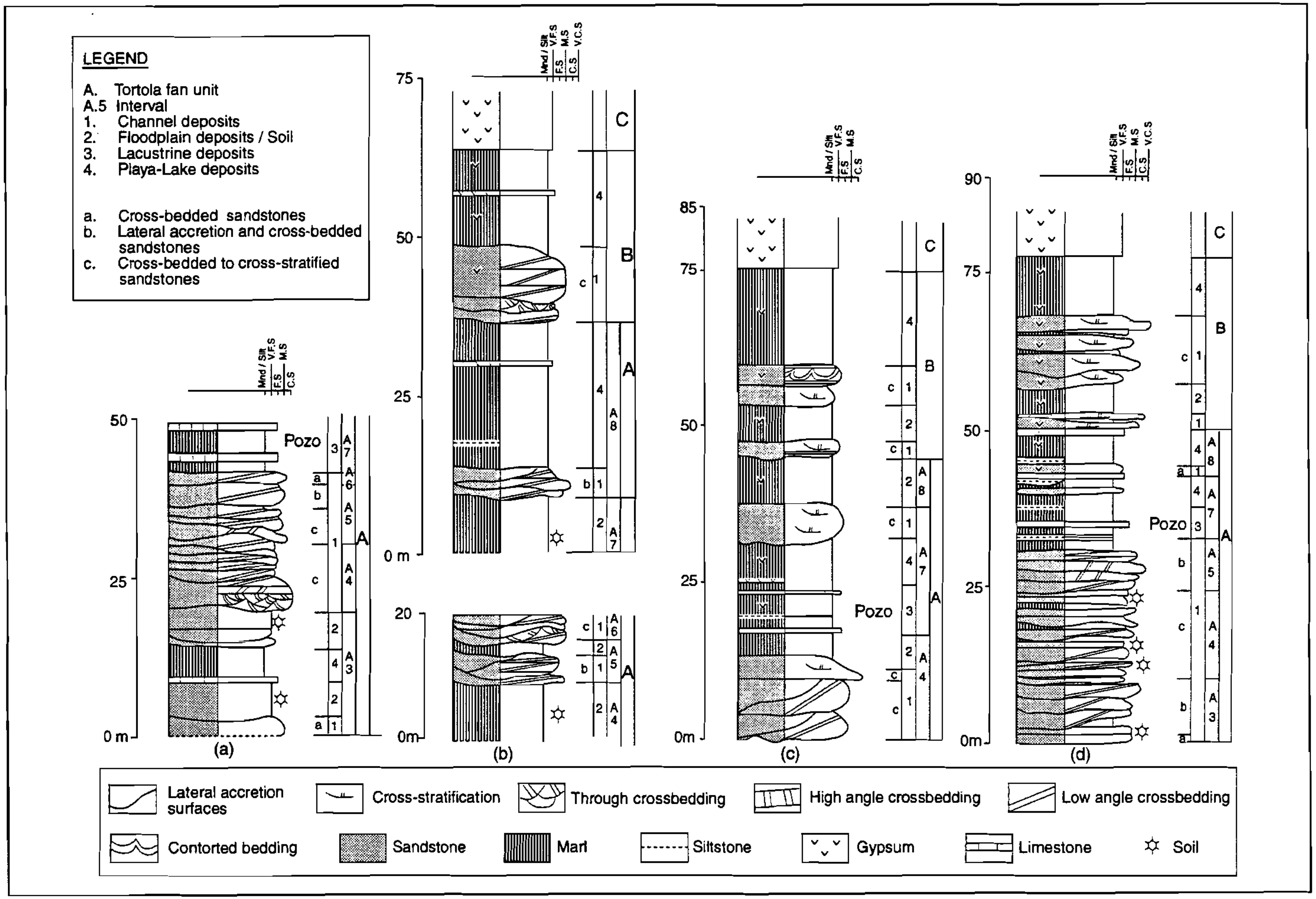
Contrary to the above playa-lake deposits, the lacustrine deposits are characterized by a greater lateral extension. They also show a similar large variation in lithology. They consist of marls, clayey-marls, siltstones, fine to very fine-grained sandstones, limestone and silty limestone beds with thicknesses of 5 cm to more than 1 m. The beds show lateral thickness variations and can be traced over long distances. Three important and laterally extensive layers occur. The lower and the middle marker beds are named Parrales 1 and 2, respectively (Díaz-Molina et al., 1989). They consist of (silty) limestones and clayey to silty marls. These lacustrine marker beds occur between intervals A2 and A3, and intervals A4 and A5, respectively. The upper lacustrine marker bed, occurring between intervals A7 and A8, is named Pozo marker bed. It consists of (silty) limestones, clayey-marls, siltstones and fine- to very fine-grained sandstones.

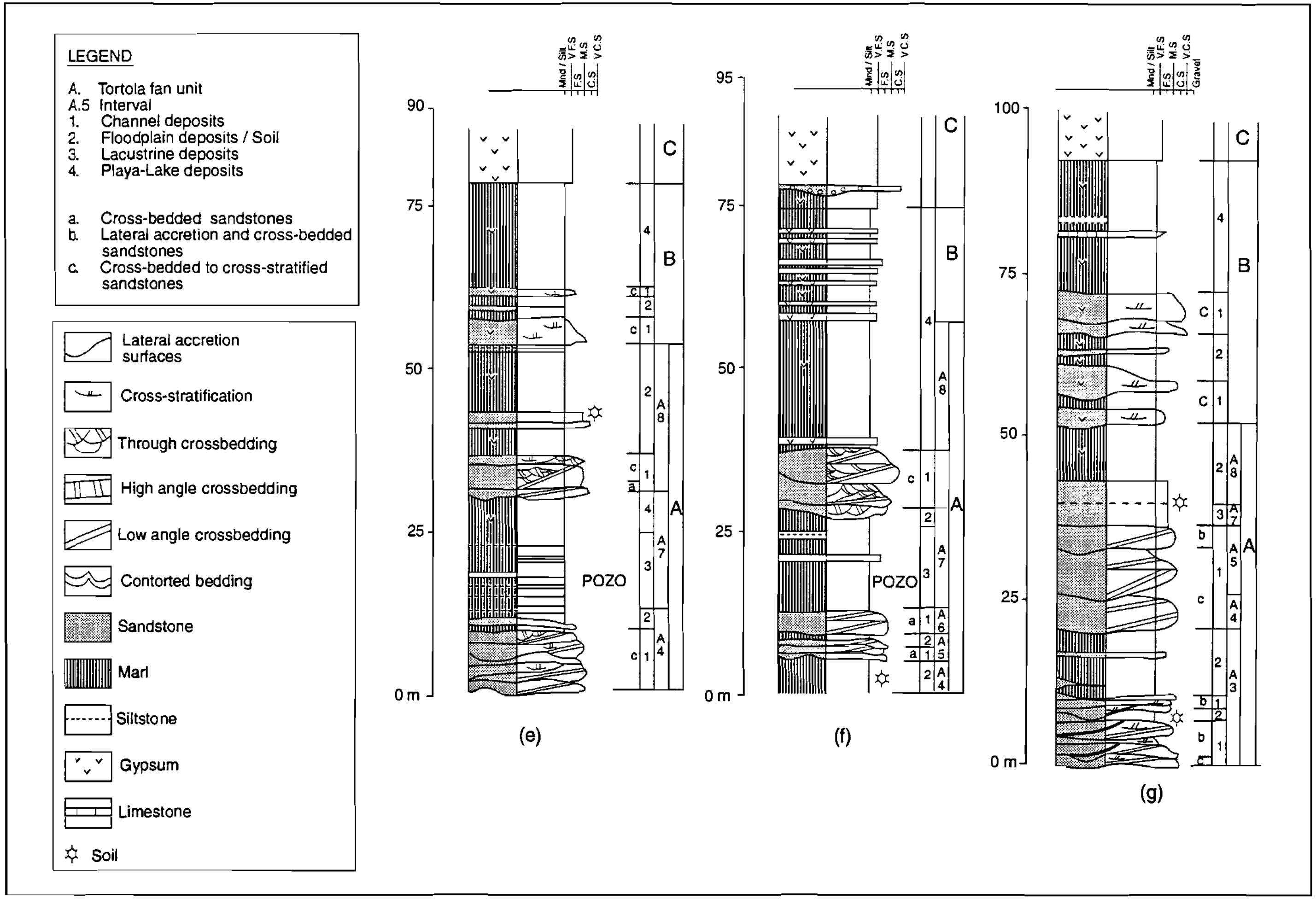
Coarse-grained alluvial facies

The observed morphology and the distribution of sedimentary structures of the coarse-grained deposits are indicative of river channel deposits. The fine-grained deposits formed from overbank deposition on the floodplain. The channel types comprise braided streams (Fig. 11 and 12), low-sinuosity (fairly straight) channels and high-sinuosity (meandering) channels (Fig. 13).

The well exposed alluvial channel sandstones are grey, ochre, yellow greyish, and yellow. They are embedded in fine-grained deposits (Figs. 8a, 8b and 9). Most of the channel bodies have a symmetrical cross-sectional shape. The sandstones are predominantly very fine to very coarse quartz sand, but occasionally clay or quartz pebbles are present.

***Figure 10**. See pages 69 and 70. Sedimentological logs (a through f) in the Lower Tórtola fan the western part of the Loranca Basin. Localities of logs are shown in Figures 8 and 9. The logs show different types of the fluvial deposits and the typical internal sedimentary structures in the studied area.*


LEGEND
A. Tortola fan unit
A.5 Interval
1. Channel deposits
2. Floodplain deposits / Soil
3. Lacustrine deposits
4. Playa-Lake deposits
a. Cross-bedded sandstones
b. Lateral accretion and cross-bedded sandstones
c. Cross-bedded to cross-stratified sandstones
Mud / Silt
V.F.S
F.S
M.S
C.S
V.C.S
Pozo
(a)
(b)
(c)
(d)
0 m
25
50
75
85
90
20
Lateral accretion surfaces
Cross-stratification
Through crossbedding
High angle crossbedding
Low angle crossbedding
Contorted bedding
Sandstone
Marl
Siltstone
Gypsum
Limestone
Soil


LEGEND
A. Tortola fan unit
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1. Channel deposits
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Lateral accretion surfaces
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Through crossbedding
High angle crossbedding
Low angle crossbedding
Contorted bedding
Sandstone
Marl
Siltstone
Gypsum
Limestone
Soil
Mud / Silt
V.F.S
F.S
M.S
C.S
V.C.S
Gravel
POZO
0 m
25
50
75
90
95
100
A
B
C
(e)
(f)
(g)

Braided Stream Deposits

Braided stream deposits are characterized by large-scale trough and planar cross-bedding and small-scale trough cross-bedding. High to low angle cross-bedding is present, but not very abundant. Occasionally parallel lamination is present. The dimensions of this type of channel body is variable. Channel depth varies from 5 to 7 m, and in case of vertically stacked channels up to 25 m. The channel width ranges from a few tens to a few hundreds of metres. Channel bars of various dimensions are found. Coarse-grained lag deposits are present at the base of the channels. Sometimes the channels show a vertical and lateral aggradation (interval A4) (Fig. 14). Mudstone and siltstone beds between the sandstone bodies are interpreted as the result of flooding events during which fine-grained sediments were deposited on the higher parts of the channel belt. This type of closely stacked multistorey sandstone bodies, separated by thin beds of fine-grained sediments (less than 10 % of the total thickness) is interpreted as the product of low-sinuosity, braided channels which were repetitively incised, and is characterized by minor imbricate channels, channel incisions on sandy bars, and the preservation of bedforms that are characteristic for alluvial islands during periods of low river discharge. Imbricate channel fills are the result of vertical aggradation.

In general, the braided stream sandstone bodies are the result of several stages of aggradation. The channels show a convex shaped base in cross section perpendicular to the flow. The channels can be divided into two groups: wide multichannel braided streams (A4.1-4.3 in Figure 14) and small braided streams (A5.2 in Figure 14).

Figure 11. *Photograph showing braided channel fills in interval A4 in the Lower Tórtola Fan in the western part of the Loranca Basin.*

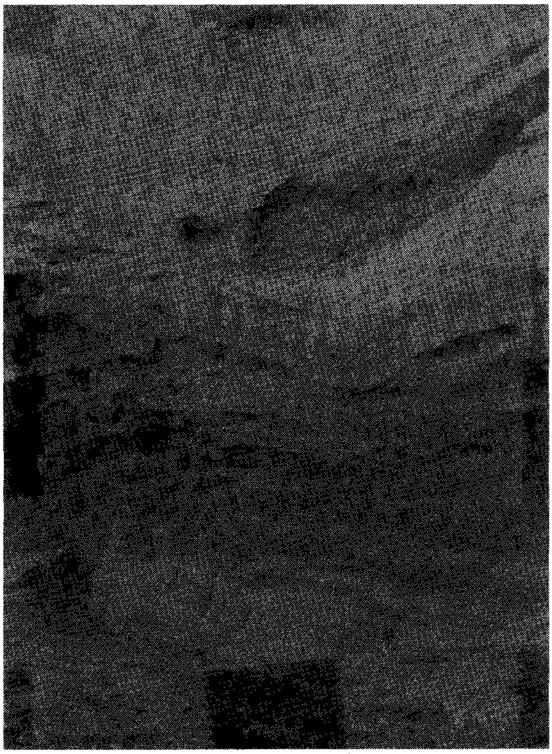


Figure 12. *Detail of Figure 11, showing internal structures in a braided channel body. The lower part of the channel body consists of medium to coarse-grained sandstones with high-angle cross-bedding and trough cross-bedding. The upper part consists of low-angle cross-bedded sandstones.*

Ephemeral braided stream deposits

In the lower part of intervals A4 (Fig. 10e) and A7 coarse to very coarse-grained, evenly laminated to small scale cross-bedded sandstones are found. These intervals show shallow erosional surfaces. The evenly laminated beds are the result of high current velocities during episodic runoff. In general these deposits are characterized by vertical aggradation and multiple concave shallow erosional bases. These deposits are interpreted as the products of ephemeral braided streams. Ephemeral systems are rarely able to establish an equilibrium fluvial style because of the short period of river discharge.

Low-sinuosity Channel Deposits

The low-sinuosity channel deposits in this region are characterized by vertically accreted sandstone bodies with a simple convex base and one or more fining-upward sequences. The channel fills mainly consist of cross-bedded to cross-stratified coarse-grained sandstones. Locally these deposits contain cross-bedding produced by almost straight-crested megaripples. Moreover these deposits are characterized by channel-lag deposits containing a great number of intraformational clasts, derived from eroded floodplain deposits. In case of vertically stacked channel bodies, the channels show moderate to severe erosion into the older channel sandstones. We agree with Díaz-Molina et al. (1989) that this type of channel deposit fits well into the classification of ribbon sandstone bodies of Friend et al. (1979). This classification gives a width to height ratio of less than 15 for the individual single channels or multistorey channel complexes. Both these characteristics are found in the research area.

The low-sinuosity channel bodies are symmetrical in shape. Occasionally individual channel bodies are stacked. These stacked channel bodies represent reactivation of river systems after periods of non-activity.

High-sinuosity Channel Deposits

Large-scale low-angle cross-bedding is a dominant sedimentary structure in many channel sandstone bodies in the area. This type of bedding is interpreted as having been caused by lateral accretion on pointbars in high-sinuosity (meandering) rivers (Figs. 13 and 14). The height of intervals with lateral accretion surfaces generally is around 1.5 m. The lateral-accretion beds are internally through cross-bedded with set heights of 0.3 to 1 m and foreset angles of 20 to 28°. Sandstones are medium to fine, moderately sorted, sub-rounded to rounded sands.

An example of lateral accretion surfaces is shown in Figure 13. This type of deposit (Fig. 14) compares well with the classical model of point bar bodies. The presence of large-scale low-angle epsilon cross-bedding provides abundant evidence for lateral accretion as the primary depositional mechanism (cf. Allen, 1965b; Moody Stuart, 1966). Channel lag deposits at the channel floor cover erosional surfaces. The preserved trough cross-stratified beds are the result of migrating sinuous-crested dunes, more or less perpendicular to the direction of lateral accretion. In the shallower parts of the channels, current ripples are the main bedforms. According to Díaz-Molina et al. (1989) meander loops have been abandoned gradually by the development of chute cut-offs or suddenly as a result of neck cut-off. Neck cut-offs resulted in the forming of oxbow lakes in which sedimentation of fine-grained sediments occurred during overbank flooding.

The undulating upper surfaces of the high-sinuosity channel deposits are interpreted as reactivation surfaces of meander loop complexes.

The high-sinuosity channel bodies are symmetrical in shape and sometimes form a stacking of individual channel bodies. These stacked channel bodies represent reactivation of river systems after periods of non-activity.

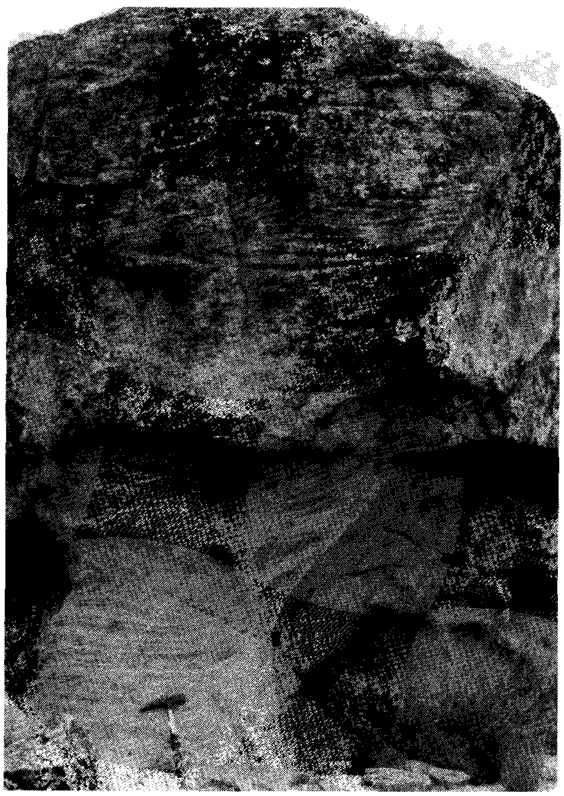


***Figure 13**. Well developed lateral accretion surfaces in high-sinuosity channel fill of interval A5 in the Lower Tórtola fan in the central part of the research area north of Huete.*

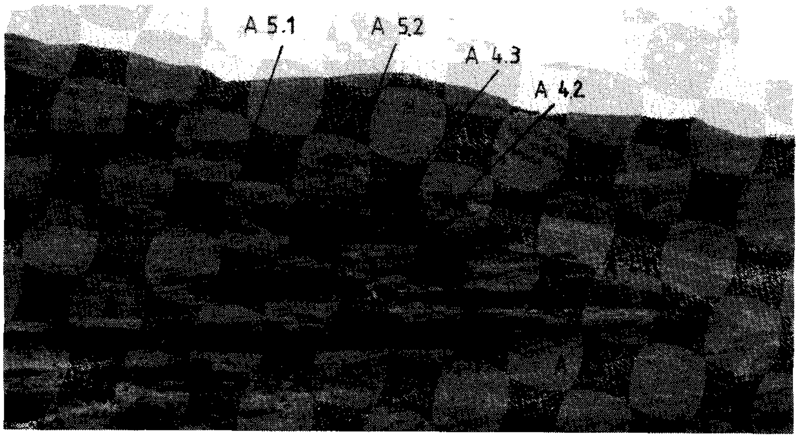


__Figure 14__. Outcrop in the Lower Tórtola Fan the central part of the studied area north of Huete with coarse-grained channel bodies of interval 3 through 5. Tabular bodies with well-developed lateral accretion bedding form the top of interval 3 (bodies A3.1 and A3.2). Braided channel bodies form most of interval 4 and the lower part of interval 5.

Palaeocurrents

426 palaeocurrent measurements have been carried out by de Vries (1990), de Haas (1991), van den Hurk (1989) and the author. Figure 15 shows palaeocurrent data of the different depositional intervals as well as the combined data. The overall palaeocurrent direction of most intervals to the NW indicates a main river system flowing from the SE to NW, which is in agreement with the palaeogeographic setting of the Tórtola Fan as well as with data of Díaz-Molina et al. (1989). The observed variations can be ascribed to the variability of flow directions in high-sinuosity channels. The random palaeocurrent pattern in interval A5 likely is the result of the low number of measurements (n=10) in this interval and the high-sinuosity character of the channels.

Fluvial Architecture

Interval A2

The sandstone bodies in this interval (Fig. 8b) are at maximum 50 m wide and have a maximum thickness of about 9 m. The lower part of this interval consists of low-sinuosity channel fills, the upper part of high-sinuosity channel deposits. The number of channels and their dimensions decrease towards the top of this interval.

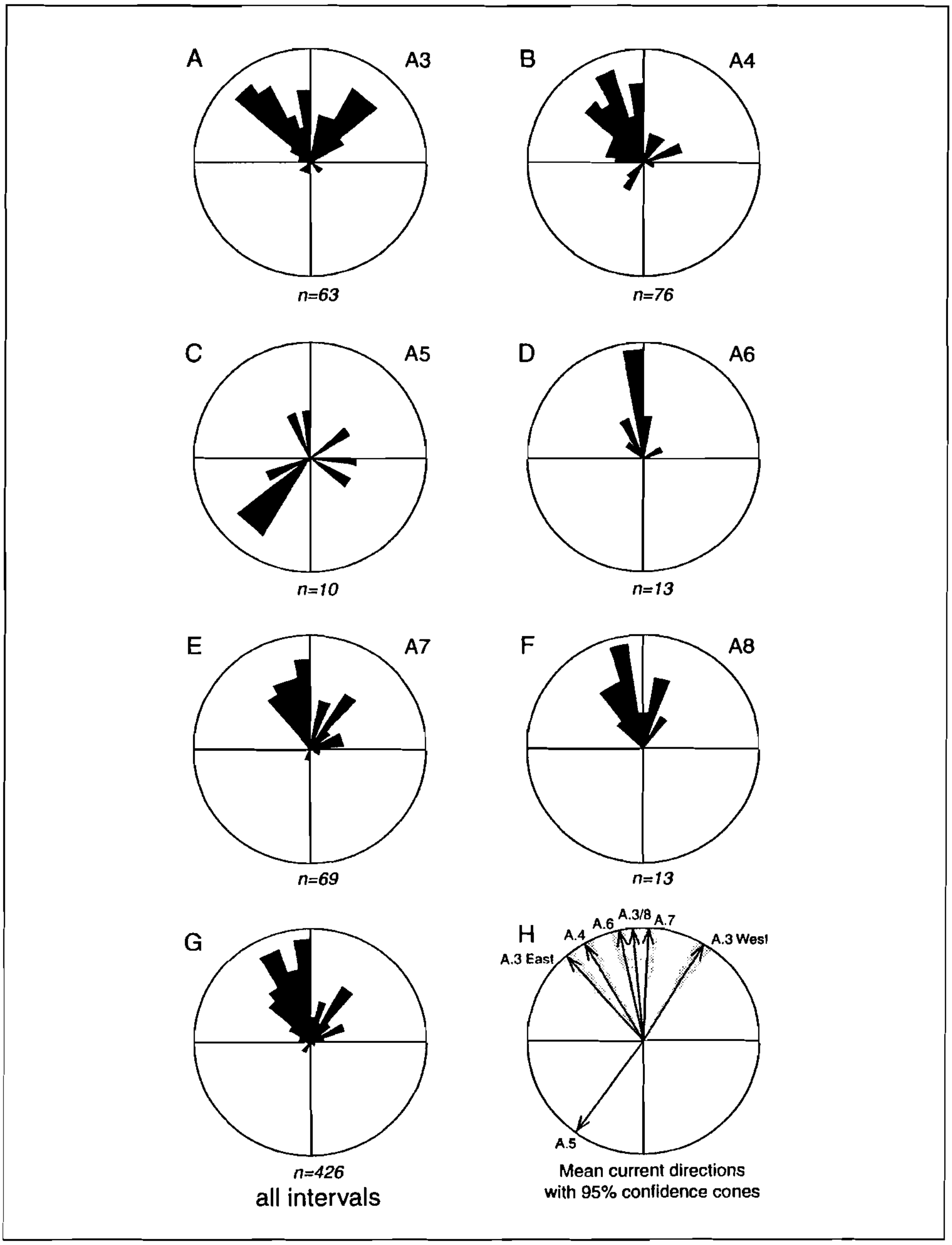


Figure 15. *Rose diagrams of palaeocurrent directions measured in intervals A3 to A8.*

Interval A3

The lower part of interval A3 is characterized by small low-sinuosity channels (Fig. 8). This interval was formed during a short period. During this period a relatively low runoff occurred and several small single channel fills were formed (lower interval A3.1). The alternation of high and low-sinuosity channels is the result of variations in runoff and concomitant increases and decreases in channel dimensions (Figs. 8 and 9). This period was succeeded by a long period in which large high-sinuosity river systems were active (upper interval A3.1). Next, a period occurred with presumably regular and relatively low runoff during which no river system was active. During this period small playa-lake deposits and paleosols were formed (Figs. 8a and b). In the southern part of the area, a single high-sinuosity channel body showing lateral accretion surfaces occurs in this interval (Fig. 8b), and in the lower part of interval A3 in the central part. This channel and the channels further to the north are differently stacked. The channel in the south shows lateral accretion, in contrast to the vertical aggradation found in the north. The channels were active during the same period. The difference in type of stacking is likely due to differences in local subsidence which was stronger in the north than in the south (Figs. 8b and 9).

Most sandstone bodies in the A3.2 interval (Fig. 9) show a stacking of symmetrical low-sinuosity to braided channels (cf. Rust, 1978). During periods of low discharge, currents within the braided river belt flew around mobile channel bars (Fig. 11). Successively more regular runoff resulted in the formation of small low-sinuosity channels which gradually changed into high-sinuosity channels. Individual single-storey channels show strong erosional surfaces.

The most southern exposures of interval A3 show a group of vertically stacked channel fills (Figs. 8 and 9). Probably, the short periods in which the individual channels were active are the result of frequent avulsion higher up the Tórtola Fan. After large braided systems were formed, a change in type of runoff occurred resulting in the formation of small low-sinuosity channel fills. A successive increase in runoff led to the long-term development of a high-sinuosity river system which resulted in the deposition of (groups of) high-sinuosity channel fills in the central area.

Interval A4

The lowermost part of this interval consists of channel fills interpreted as (ephemeral) braided streams. These deposits are covered by low-sinuosity channel fills. The small channel fills are vertically stacked, each fill indicating a short period of activity.

The intermediate part of interval A4 consists of braided to low-sinuosity channel fills. The channels in the upper part are larger than in the lower part. The contact between the middle and uppermost interval is occasionally erosive.

Towards the top of the interval, the dimensions of the individual channels decrease, indicating a decrease in runoff and sediment supply (Figs. 8a, 8b and 9). The uppermost part of interval A4 consists of deposits of braided streams, low-sinuosity and occasionally high-sinuosity channels. The top of interval A4 is formed by a lacustrine layer.

Interval A5

The lower part of interval A5 consists of a group of braided to low-sinuosity channel bodies. These channel bodies are difficult to correlate as a result of erosion by younger channels. The intermediate part of interval A5 consists of a group of braided stream and low-sinuosity channel bodies. The upper part of interval A5 is built up of a group of low to high-sinuosity channel bodies, one single high-sinuosity channel body and some single low-sinuosity channel bodies.

Interval A6

The lower part of interval A6 consists of isolated single and stacked low-sinuosity channel bodies. The upper part of interval A6 consists of one group of braided channel bodies, two single low to high-sinuosity channel bodies and one single high-sinuosity channel body. The group of braided composite channel bodies can be divided into two generations of individual channel bodies. The deposition of the first generation was followed by a period without any recognizable river activity. After some time a new river system developed and produced a deep erosional surface which was filled with braided channel deposits. Afterwards some smaller low to high-sinuosity channels developed.

Interval A7

The sandstone bodies of this interval, which are much larger than those in the previous interval, are the result of severe erosion followed by a sediment supply which was larger than during the previous periods. This interval consists of two parts: A7.1 and A7.2. Interval A7.1 consists of four groups of large braided channel fills and one single channel fill. One of the groups also contains an ephemeral braided channel fill. In the north of the area where this interval is exposed, also a group of low-sinuosity channel fills is present. Interval A7.2 consists of single and groups of several braided and low to high-sinuosity channel bodies. The transition from one type of channel to the other is, as is the case in the previous intervals, the result of changes in runoff and sediment transport.

Interval A8

This interval is exposed throughout the whole area. In general the individual sandstone bodies are smaller in size than those of the older intervals. Interval A8.1 consists of several single low-sinuosity channel bodies which sometimes show lateral accretion surfaces. Interval A8.2 consists of a group of braided to low-sinuosity channel bodies. Interval A8.3 finally consists of a group of low-sinuosity channel bodies. The intervals A8.1 and A8.2 are separated by playa lake deposits and soils which developed in overbank fines. Intervals A8.2 and A8.3 are separated by floodplain deposits with some soils. The top of interval A8.3 is formed by playa lake deposits and soils.

DISCUSSION: CONTROLS ON FLUVIAL ARCHITECTURE

Fluvial style is a complex response to a number of autocyclic and allocyclic controls. Large fluvial complexes tend to produce integrated drainage networks containing one or more trunk streams of the same type (i.e., meandering, braided, etc.) for large parts of the network (Galloway, 1985). The development of a specific type of fluvial system is controlled by the tectonic, climatological, hydraulic, and morphological conditions (Flores, 1985). The primary allocyclic controls include the climate, controlling weathering of parent rocks and run-off (discharge), and the tectonics, controlling basin slopes and relief of the drainage area (Allen & Allen, 1990). For a discussion on autocyclic controls see Chapter 2 (page 49).

Generally, climatological changes are considered to cause only minor overprints in a tectonically controlled depositional system rather than being a critical factor that determines the system. Olsen (1989) however, gives a model for humid environments where runoff is a combination of surface runoff, infiltration, subsurface and return flow, and changes with changing climate. Yair & Klein (1973) assume that surface runoff is dominant and solely responsible for channel initiation in present-day semi-arid depositional environments, because surface runoff mostly is a response to rainfall in dry climates. Perlmutter & Matthews (1989) made a detailed model assuming that the geographic position of a basin during a particular time interval together with regular climate changes dictates the basin stratigraphy. In their study Milankovitch oscillations are suggested to be dominant in certain cases.

Changes in channel style from low to high-sinuosity, as seen in Figures 11 through 14, are interpreted as the result of a change in the type of river discharge. The size of a single storey channel is the result of erosion and is therefore related to the maximum discharge during the period in which the channel was active. Vertical accretion in channels takes place when the discharge of the river decreases and the flow is not strong enough to maintain the channel depth adapted to erosion during maximum discharge. This results in the deposition by vertical accretion of coarse-grained sediment in the channel. When channels were abandoned due to avulsion higher up on the Tórtola alluvial fan, the channels became filled with fine-grained deposits. This process is comparable to the deposition of fine-grained sediments in oxbow lakes, forming clay plugs.

Changes in river discharge are reflected in the type of fluvial channel. Low-sinuosity ribbon type channel bodies mainly show vertical accretion. This type of channel fill, which is abundant in the studied succession, was formed when discharge was of relatively short duration and the rate of avulsion was high. Channels formed during periods of more continuous river discharge tend to be more sinuous. During periods with a regular runoff relative large channels were formed. A decrease in discharge and sediment supply resulted in the formation of relatively small sandy fills of large channels. Sinuous channels formed classic point bars in meander loop complexes, in which usually lateral accretion was dominant. When the runoff decreased or became irregular, channels did not instantaneously adapted to these new hydrodynamic conditions. Changing conditions caused vertical accretion or a change in the pointbar geometry. Complex pointbar geometries are frequently found in the studied succession and have been described in detail by Díaz-Molina (1989). The reason for this may be that periods of low runoff were too short for the river to reach a new equilibrium and to obtain a straight or braided character. A second reason may be a decrease in the sediment supply.

The formation of depressions combined with a small, regular runoff resulted in the formation of small (playa) lake deposits, of which several have been found (Fig. 8). Such depressions could form as a result of local differential compaction between sandy channel fills and fine-grained floodplain deposits after a period of relatively fixed channel position.

From the close stacking of the sandstone bodies, especially in intervals A4 and A5, it is deduced that the subsidence in this area was low relative to the amount of sediment supplied (cf. Blakey & Gubitosa, 1984). The northward shift of the main channels (Fig. 4 and 8) must have resulted from a gradual northward shift of the centre of subsidence of the Loranca basin during deposition. This trend in subsidence continued until the lacustrine deposits of the Pozo marker bed were formed.

The above mentioned features suggest that the conditions in terms of river discharge changed regularly during development of the Tórtola Fan. The base of most intervals constituting the Tórtola Fan succession consist of coarse-grained, low-sinuosity channel bodies, comprising braided to straight channel types. The upper parts generally consist of finer-grained, high-sinuosity channel bodies with abundant floodplain deposits and intercalated paleosols. In addition, the number of channel bodies and their dimensions tend to decrease towards the top of the intervals.

The intervals thus show a general pattern of development. This must have resulted from regularly changing conditions in terms of alluvial discharge. Cycles of alluvial deposition started with a highly irregular discharge, resulting in braided and occasionally ephemeral braided river systems. The river systems gradually developed into straight and finally high-sinuous channels, when the discharge became more regular. This points to a system with more or less regular variations of precipitation and discharge.

A principle motive for the initiation of this study has been to analyze whether astronomically induced climate changes have produced the rhythmic sequence in the Loranca Basin (cf. Perlmutter & Matthews, 1989). The character of the sequence certainly would fit into such a model in which quasi-periodic changes of climate produce fluvial successions with cyclic changes of the character of the fluvial deposits. However, the poor stratigraphic resolution and the frequent occurrence of depositional hiatuses prohibit, as yet, to unequivocally demonstrate the validity of such an hypothesis.

CONCLUSIONS

Vertical changes in the character of the fluvial channels in the Tórtola Fan system of the Loranca Basin can be related to changes in climate. It is concluded that:

- Interval A2 reflects a relatively humid climate with regular runoff. This resulted in the formation of a high-sinuosity river system. The continuous increase in humidity (see Fig. 9) resulted in the deposition of lacustrine deposits on top of unit A2.
- Interval A3 consists of an alternation of several high and low-sinuosity channel bodies reflecting more or less regular variations in runoff. This pattern continued during deposition of interval A4, after which an increase in humidity caused the formation of the lacustrine environment which forms the top of interval A4.
- At the onset of the formation of interval A5 the relative humidity decreased and runoff became more irregular, resulting in the formation of low-sinuosity channel bodies in the lower part of A5. Relatively more high-sinuosity channel bodies in the upper part of interval A5 must be due to an increase of the regularity and possibly also a net increase of the runoff.
- Interval A6 is similarly characterized by a change in the depositional regime which led to a succession of low-sinuosity channel bodies in the lower part and high-sinuosity channel bodies in the upper part of this interval.
- Intervals A7 and A8 again show a more or less regular repetition of periods characterized by changes in precipitation and runoff, with a lacustrine interval forming the boundary between both intervals.

Present-day knowledge about astronomical processes confirms their influence upon the climate. It seems justified to assume that regular climatological changes were an important factor controlling the character of the Tórtola Fan succession. In addition, tectonic conditions were relatively quiet during development of the Tórtola Fan. Therefore, good arguments can be furthered to state that the rhythmic variations in depositional style seen throughout this succession are due to changes in precipitation and runoff. This may be, partly or fully, due to astronomically defined climatological changes.

CHAPTER 4

SEDIMENTARY FACIES AND SEQUENCES IN THE EOCENE OF THE RIO NOGUERA RIBAGORZANA VALLEY, TREMP-GRAUS BASIN, SOUTH-CENTRAL PYRENEES, SPAIN

INTRODUCTION

The relation between tectonics and sedimentation in the Southern Pyrenean Basin is expressed in Mesozoic and Tertiary sedimentary cycles. Puigdefàbregas & Souquet (1986) described the migration of foreland basins (Eocene-Early Oligocene) of the Pyrenees in relation to their tectonic development. The depositional systems in the Paleogene of the South Pyrenean foreland basin provide an outstanding example of eustatic and tectonic control on depositional sequences, systems, and systems tracts. Marzo et al. (1988) showed that tectonic activity caused southward and northward shifts of the axial fluvial system, whereas vertical aggradation was controlled by both tectonic activity and eustasy. Burbank et al. (1992) emphasized the control of thrust development, particularly with respect to hindward-imbricating successions in the south-central Pyrenees. Phases of strong progradation of fluvial systems are interpreted to have been induced by falls of sea level (Marzo et al., 1988).

This chapter intends to outline the sedimentary facies and sequences of the Eocene (Upper Ypresian to Upper Lutetian) Lower Montañana Group in the Rio Noguera Ribagorzana valley (Tremp-Graus Basin, southern Pyrenees, Spain, Fig. 1) based on the alluvial architecture development. The aim is to analyze the character of large-scale lateral shifts of the transition zone between alluvial and shallow marine deposits and of the cyclic character of the succession and to analyze which were the main controls. This will be achieved by:

- Establishing a high-resolution stratigraphic model for the alluvial-fluvial to shallow-marine basin fill.
- Analyzing the variability in alluvial-fluvial facies and the cyclic character of the observed changes in facies.
- Discussing possible causes of the cyclic character of the succession in terms of tectonics, climate and third-order eustatic cycles.

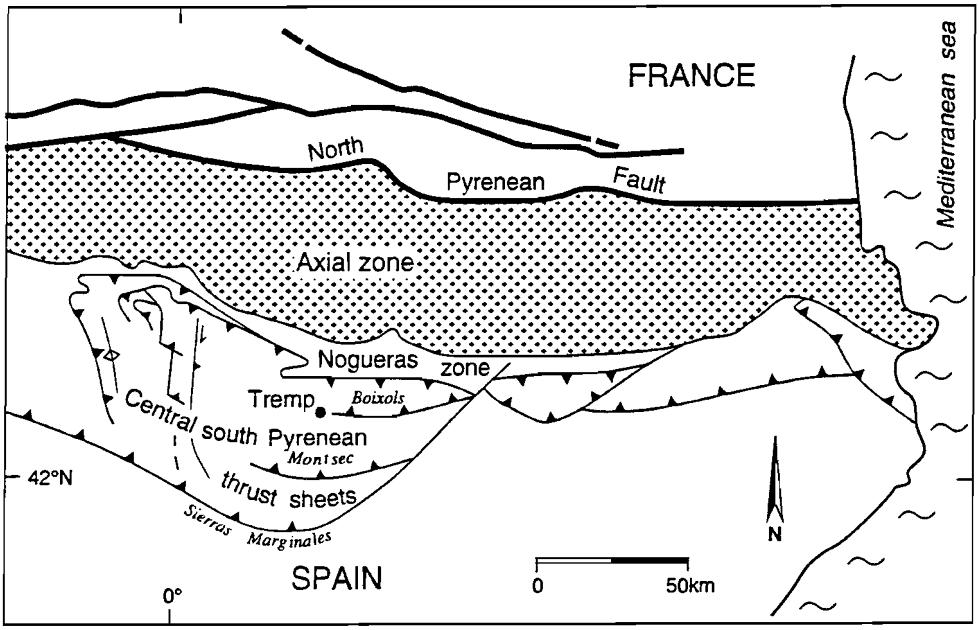


Figure 1. *Schematic map of the Pyrenees with the main tectonic units and the central south Pyrenean thrust sheets. The studied area is west of Tremp. Modified after Camara and Klimowitz (1985), Williams (1985) and Puigdefàbregas et al. (1986).*

Alluvial deposits are an important component of the stratigraphic record. They occur in a wide range of tectonic settings, and are sensitive indicators of extrabasinal controls such as tectonic activity, sea-level changes and climate. At a basin-scale, the stratigraphic architecture of fluvial deposits, the three-dimensional distribution of facies, yields information about the major allogenic controls on deposition (cf. Miall, 1992). Changes in fluvial architecture may indicate changes in topographic gradients, changes in runoff and/or relative sea-level changes. Relative sea-level changes in a given area are the result of the interplay between global eustasy, local tectonics, the rate of clastic sediment supply and isostasy (Plint, 1991).

A detailed overview of the vertical and lateral lithofacies distribution of the Eocene Montañana Group has been made in order to construct a high-resolution stratigraphic model for the deposits in the Rio Noguera Ribagorzana valley area. The deposits in the Rio Noguera Ribagorzana valley are subdivided into the Lower and Upper Montañana Formations separated by the Castissent Formation (Nijman & Nio, 1975, Van der Meulen, 1989). The Castissent Formation represents a phase of strong fluvial progradation during the upper Ypresian, separating the Lower and the Upper Montañana Group (Nijman & Puigdefàbregas, 1989). Intercalated in the fluvial and alluvial deposits of the Montañana Group, fine-grained marine intervals occur, which have a vast lateral continuity. These marine intervals represent marine flooding surfaces formed during relative sea-level highs. They have been used for correlation and for the definition of depositional sequences.

METHODS

In the field sandstone bodies and conglomerate sheets were mapped, and logs on 1 : 100 and 1 : 500 scale were made. The stratigraphic succession and vertical and lateral facies relationships were reconstructed as well as palaeocurrent patterns. The attention was focused on the coarse-grained sediment complexes. The correlation of these complexes is based on laterally extensive, brightly coloured fine-grained overbank deposits, paleosols, and marine deposits.

GEOLOGICAL SETTING OF THE TREMP-GRAUS BASIN

The Pyrenees were formed during a phase of late Cretaceous-Miocene convergence and limited northward underthrusting of the Iberian plate beneath Eurasia (Fig. 2) (Muñoz, 1991). The evolution of the Pyrenean mountains as part of the Alpine orogeny is closely related to the movement of the African and Iberian plates in relation to Eurasia (Choukroune, 1969, 1970; Seguret, 1972; Choukroune & Seguret, 1973; Mattauer & Henry, 1974; Hirst & Nichols, 1986; ECORS, 1988; Muñoz, 1991). A deep reflection seismic survey across the Pyrenees by ECORS (1988) shows the results of compressional forces which mobilized the post-Triassic sedimentary pile (Fig. 2). The onset of thrust deformation associated with the Pyrenean collision was strongly diachronous from east to west (Bentham et al., 1992). In the southern Pyrenees, basins developed during Paleocene to early Eocene times upon the southerly translating thrust sheets, partly in response to thrust wedge loading, and partly due to subduction-related flexure of the down-going Iberian plate (Muñoz, 1991). Deformation was initiated during the late Cretaceous-Paleocene. It was strongest during the middle and late Eocene, and continued through the Oligocene and Miocene. According to Sole Sugranes (1978), during the late Eocene to Oligocene strong compressional movements caused the formation of the Gavernie nappe, which carried piggy-back the Montsec thrust sheet (Fig. 1).

Structural, stratigraphic and magnetic data indicate that the main phase of motion of the Montsec thrust sheet terminated 54 Ma ago (Burbank et al., 1992). However, the Montsec thrust was reactivated in the middle Late Eocene (Choukroune & Seguret, 1973; Sole Sugranes, 1978; Williams, 1985).

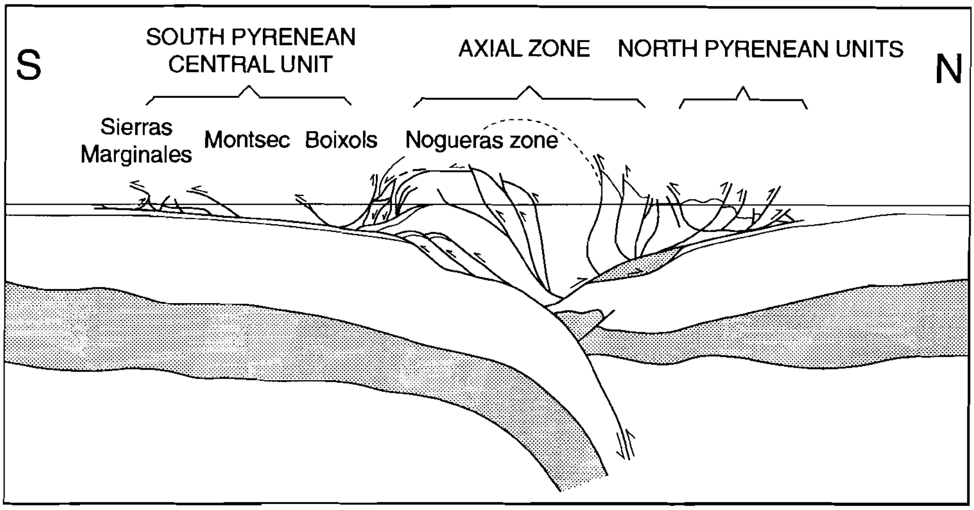


Figure 2. *ECORS profile through the Pyrenean orogen with main structural units and thrust sheets in the south central Pyrenean (ECORS, 1988).*

The Tremp-Graus Basin is a thrust-sheet-top basin on top of the Cotiella-Montsec Nappe which had its main displacement during the Eocene (Seguret, 1972; Nijman, 1981). The Tremp-Graus basin belongs to the allochthonous part of the Southern Pyrenean zone, in the south central part of the Spanish Pyrenees (Fig. 1). Tectonic activity during the Eocene had a strong impact on the sedimentation patterns within the Tremp-Graus basin.

The stratigraphy and sedimentology of the basin fill have been studied, amongst others, by Mey et al. (1968), Garrido-Megias (1968), van Eden (1970), Mutti et al. (1972), Puigdefàbregas (1975), and Nijman & Nio (1975). Detailed investigations of sedimentological aspects of the coarse clastic deposits have been carried out by, amongst others, Puigdefàbregas & Van Vliet (1978), Nijman (1981), Van der Meulen (1982, 1983), Atkinson (1983), Puigdefàbregas et al. (1986), Cuevas Gozalo (1989), Cuevas Gozalo & De Boer (1989) and De Boer et al. (1991).

The basin infilling systems and general sedimentation patterns of the Eocene foreland basin can be differentiated into two major systems (Nijman & Nio, 1975):

- The Tremp-Graus Platform, where continental, transitional and shallow marine sedimentation of clastic material and carbonates took place.
- The Ainsa deep, where clastic sedimentation occurred in a deeper marine environment.

The Tertiary formations of the Tremp-Graus Platform can be divided into four groups (Nijman & Nio, 1975):

- The Ager Group (Ypresian) that consists of a thick marl sequence with smaller and larger clastic complexes (mainly calcarenites), scattered horizons of limestone beds and local conglomeratic influxes. The facies pattern ranges from open marine to shallow marine and restricted marine (inshore to lagoonal) from the W to the E.
- The Montañana Group (Ypresian-Lutetian), which constitutes a regressive mega-sequence (progradation of a large delta complex).
- The Campodarbe Group (Upper Eocene - Lower Oligocene), which consists of continental deposits (conglomerates, sandstones, siltstones and marls) with some minor intercalations of fresh water limestones.
- The Collegats Conglomeratic Group (Upper Eocene-Oligocene), which largely consists of coarse-grained siliciclastic alluvial fan complexes.

Van Eden (1970) described the Montañana Formation, which includes all deposits in the area which are subject of this chapter. Van Eden (1970) recognized two major sedimentary environments in this formation: a floodplain environment with fluvial sandstones, conglomerates and light-coloured mudstones, and a transitional environment with littoral sediments, dark-coloured mudstones and a marine shell fauna.

Analyses of the sedimentary environments of the Lower Montañana Group, which is equivalent to the Montañana Formation of van Eden (1970), have been published by several authors (e.g., Nijman & Nio, 1975; Nijman, 1981; Van der Meulen, 1982, 1983). Nijman & Nio (1975) defined and subdivided the Montañana Group into the laterally equivalent non-marine Monllobat Formation and the marine Castigaleu Formation, the fluviatile Castissent Formation and the laterally equivalent non-marine Capella and marine Perarrúa Formations. West of the Rio Noguera Ribagorzana valley, along and west of the Isabena valley, the San Esteban and Campanúe fan sandstones and conglomerates occur (Fig. 3). The Monllobat and the Castigaleu Formations show a clear interfingering.

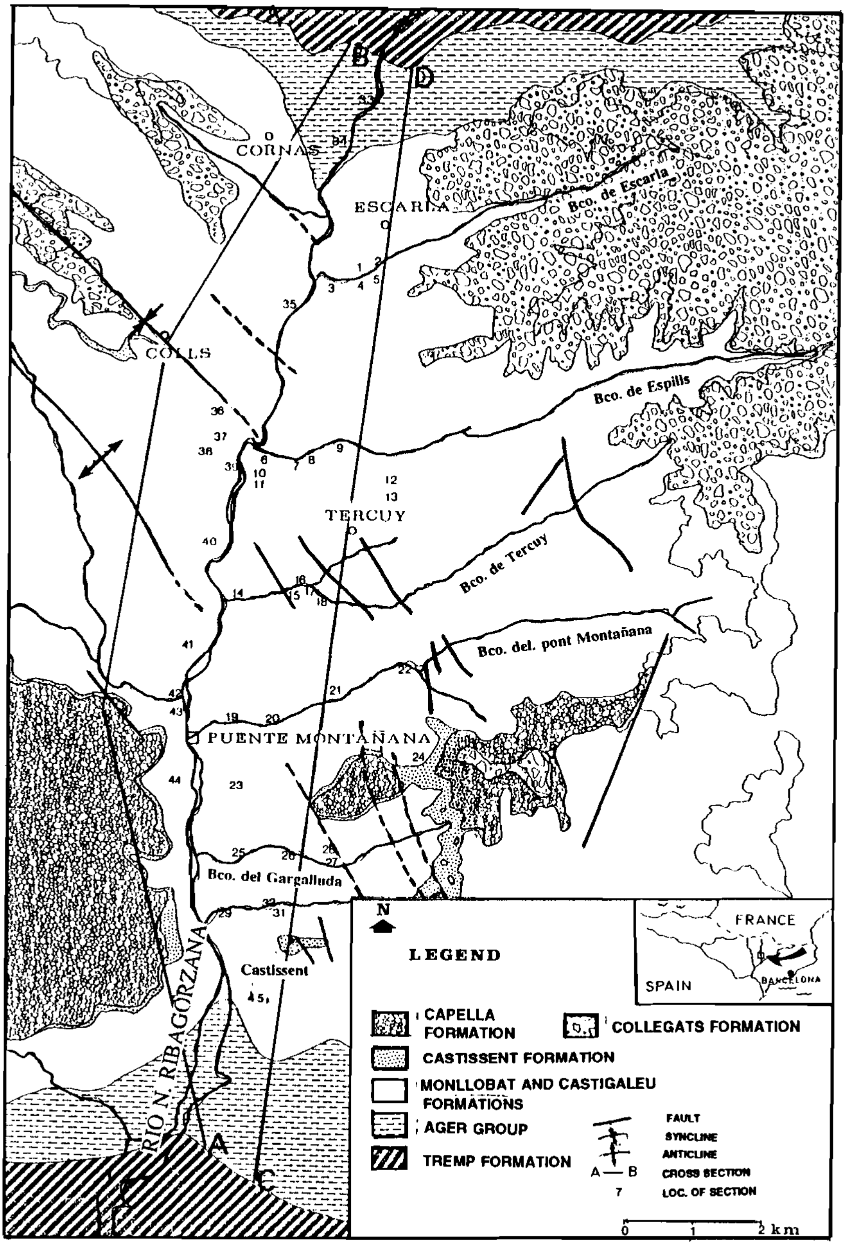


Figure 3. *Simplified geologic map of the studied area in the valley of the Rio Noguera Ribagorzana. The locations of the measured sections and the cross-sections are indicated.*

Both formations overly the Ager Group, possibly with local low-angle unconformities (Nijman & Nio, 1975). The upper boundary of the Monllobat and Castigaleu Formations is formed by the white sandstones of the Castissent Formation or, where the latter pinches out, by the alluvial and shallow marine sediments of the Capella and Perrarúa Formations (Nijman & Nio, 1975).

Structural deformation of the Montañana Group is minimal. The basin has been very slightly folded into an open, westward plunging asymmetrical syncline.

The main factors controlling the structural setting in the research area are:

- The Montsec thrust in the south, and a system of thrust sheets and antiformal stacks in the Nogueras zone in the north (Fig. 1).
- Local faults and folds (Fig. 3).
- Possible N-S aligned strike-slip fault movements along deep seated faults which may have occurred during sedimentation of the Montañana Group. One of these faults is supposed to have been active in the Rio Noguera Ribagorzana valley area and will be discussed below. Segmentation of the Tremp-Graus Basin through N-S strike-slip faults has earlier been suggested by Atkinson (1983).

SEDIMENTARY FACIES OF THE MONTAÑANA GROUP

Sedimentation of the Montañana Group was the result of a westward prograding deltaic sequence, by input of coarse-grained sediments from the SE, and by north-derived alluvial fan sedimentation. In a longitudinal E-W section the Montañana Group has a clastic wedge geometry, typical for deltaic bodies, while in N-S cross-section the basin fill reflects a synsedimentary active, synclinal structure (Nijman & Puigdefàbregas, 1989).

East of the Rio Noguera Ribagorzana Valley, an unknown portion of the Upper Montañana Group has been removed by erosion, which may have occurred since the Oligocene. Further to the north thick conglomerates of the Upper Eocene-Oligocene Campodarbe Group rest with a distinct unconformity on the underlying Montañana sediments. For this reason the true thickness and nature of the once overlying sediments are difficult to ascertain.

Within the Montañana Group three types of facies were recognized (e.g., Nijman & Nio, 1975):

1 - Coarse sandstones and conglomerates belonging to alluvial fan systems with S to SW-ward transport directions.
2 - Fluvial channel sandstones in a complex of varicoloured muds (paleosols) with transport directions dominantly towards the northwest to west.
3 - Heterolithic deposits consisting of channel, lens- and sheet-shaped sandstone units, with westward to northwestward transport directions, and grey mudstone intervals in between. These mudstones contain the remains of a marine fauna comprising oysters, gastropods and nummulitic foraminifera.

In terms of physiographic units of a deltaic environment, facies type 1 represents continental alluvial fan deposits, interfingering with facies types 2 and 3. Only where the coarse-grained units are thick and extensive, they can be mapped as separate lithological units. Still, many interfingering contacts occur. Facies type 2 represents the alluvial and upper deltaic plain. The upper deltaic plain is the part of the alluvial floodplain, down-stream of the bifurcation of the supplying river, but without marine influence (cf. Rainwater, 1966; Oomkens, 1974), or the lower floodplain of Allen (1970). Facies type 3 is the lower deltaic plain (i.e., the marine influenced inshore part of the deltaic plain) and the delta front platform (Allen, 1970) with coastal and fluvio-marine environments (Kruit, 1955; Oomkens, 1967). The above facies types occur adjacent to each other.

The Castigaleu Formation (Lower Montañana), which is exposed in the northern, southern, and western part of the research area, consists of fine- and coarse-grained sediments. The fine-grained sediments are characterized by massive, blue-grey mudstones with marine fossils, such as oysters. Also thin beds with oyster fragments are intercalated in the fine-grained sediments. The light-grey, coarse-grained deposits have locally been bioturbated strongly. The boundary with the Monllobat Formation is formed by red-coloured and mottled continental mudstone intervals (Puigdefàbregas & Van Vliet, 1978). The Monllobat Formation occurs partly laterally of the Castigaleu Formation and partly covers it. The relative volumes of coarse clastic sediments also increases to the north and east. In the north, the Monllobat Formation is unconformably overlain by the younger red-coloured Collegats Formation (Fig. 3).

LITHOFACIES

The lithofacies in the studied area can be subdivided into two major classes (Table 1, page 126): fine-grained clastic sediments, generally calcareous silty marls (F), and coarse-grained clastic sediments consisting of sandstones (SA) and conglomerates (CO). Medium- to coarse-grained lithofacies occur in stacked clusters within distinct, correlatable intervals (Figs. 4, 5, 6 and 7).

Fine-grained alluvial deposits

The central area is dominated by fine-grained deposits (Figs. 7 and 12). The fine-grained deposits consist of mudstones and siltstones. Mudstones either have a dark-grey or a light reddish colour.

The dark-coloured mudstones and silty to sandy marls (Fsi/Fsa) often contain oyster debris. Dark-coloured blue-grey mudstone intervals in the Castigaleu Formation are usually laterally continuous and can be correlated in the field for kilometres. They comprise 70-80% of the total volume of the Formation. They tend to be coarser (Fsa) towards the north. The marl intervals are generally separated by medium- to coarse-grained clastic sediments. The dark colour, indicative of reducing conditions, and the marine fossils indicate that the sediments have been covered by water most of the time, and represent transitional fluvio-marine environments.

A light reddish colour of mudstones in certain intervals indicates exposure to oxidizing conditions for longer periods. In addition to the reddish colour, colour mottling and caliche glaebules indicate pedogenic processes in paleosols. In general, oxidation during subaerial exposure and consequent paleosol development resulted in red to ochre and yellow-brown colours. Vertical mottling is typical for the brown and ochre-coloured mudstones. Blue and red mottled intervals reach thicknesses of about one metre. The metre-thick mottled horizons indicate low groundwater levels allowing the oxidation of the soil. The colour pattern and caliche intercalations are typical of pseudogley-type soils (cf. Buurman, 1980). The caliches in the fine-grained floodplain sediments point to a marked evaporation surplus.

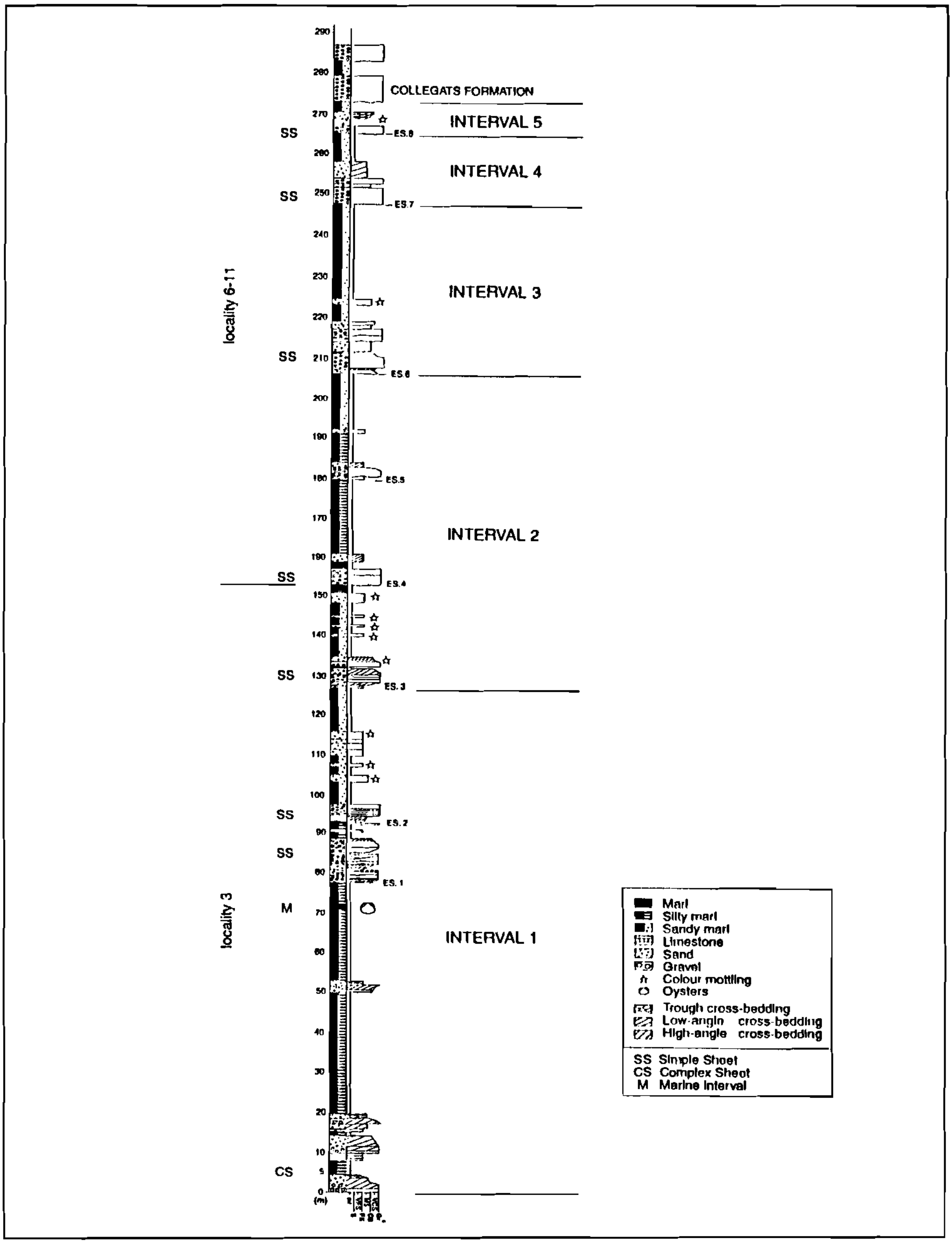


Figure 4. *Composite sections representative for the sedimentary succession of the Lower Montañana Group in the northern area. Several laterally extensive alluvial sandstone bodies are indicated (ES.1 through ES.8). The localities of the measured logs are shown in Figure 3.*

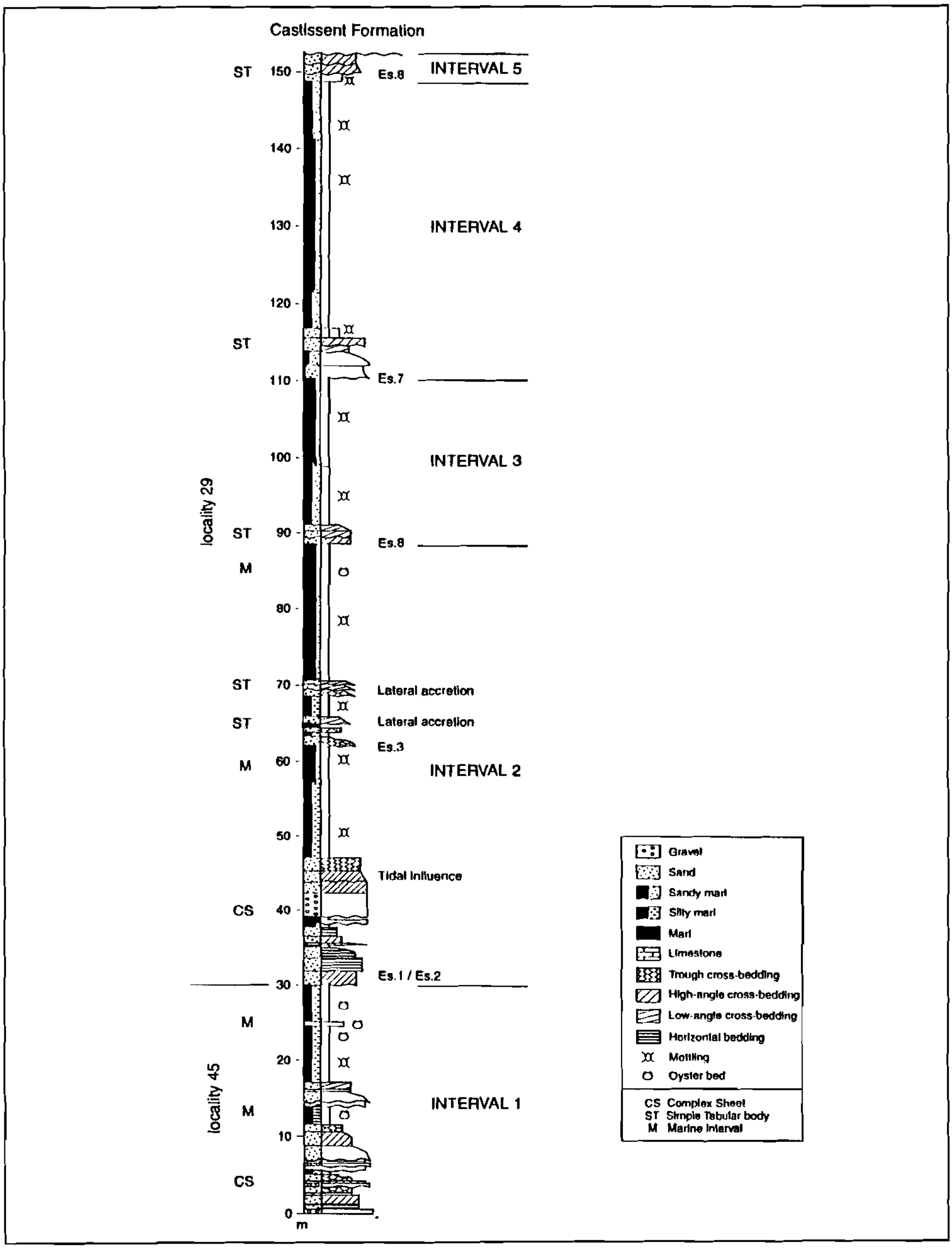


__Figure 5.__ Composite sections representative for the sedimentary succession of the Lower Montañana Group in the southern area. Several laterally extensive alluvial sandstone bodies are indicated (ES.1 through ES.8). The localities of the measured logs are shown in Figure 3.

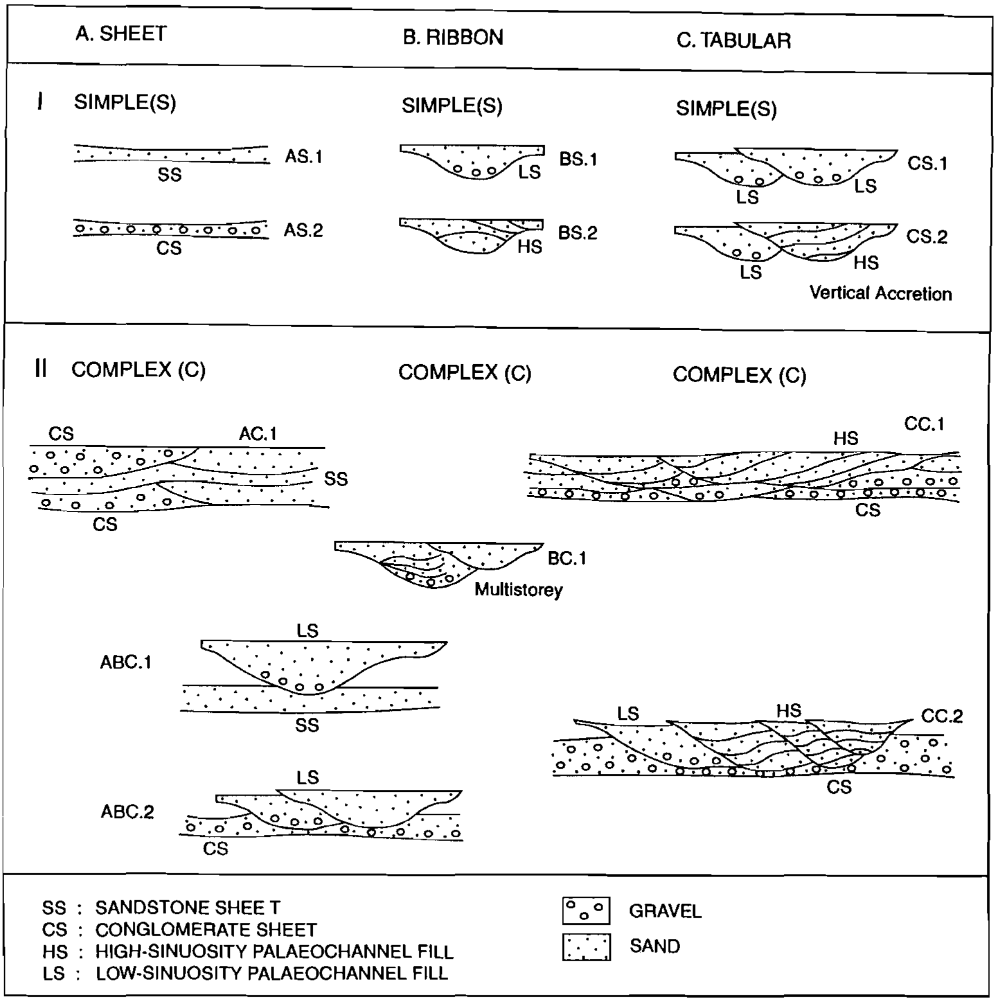
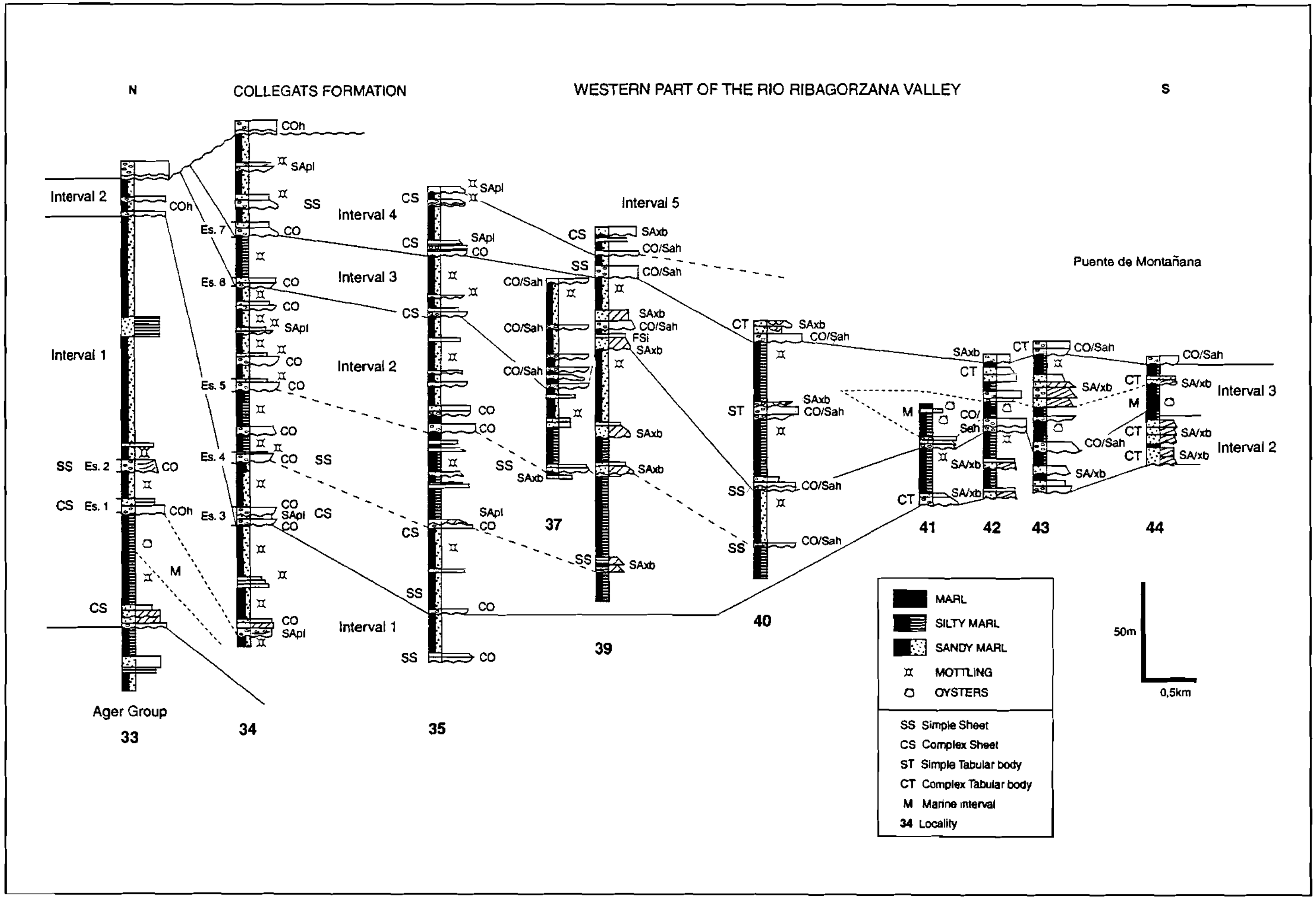
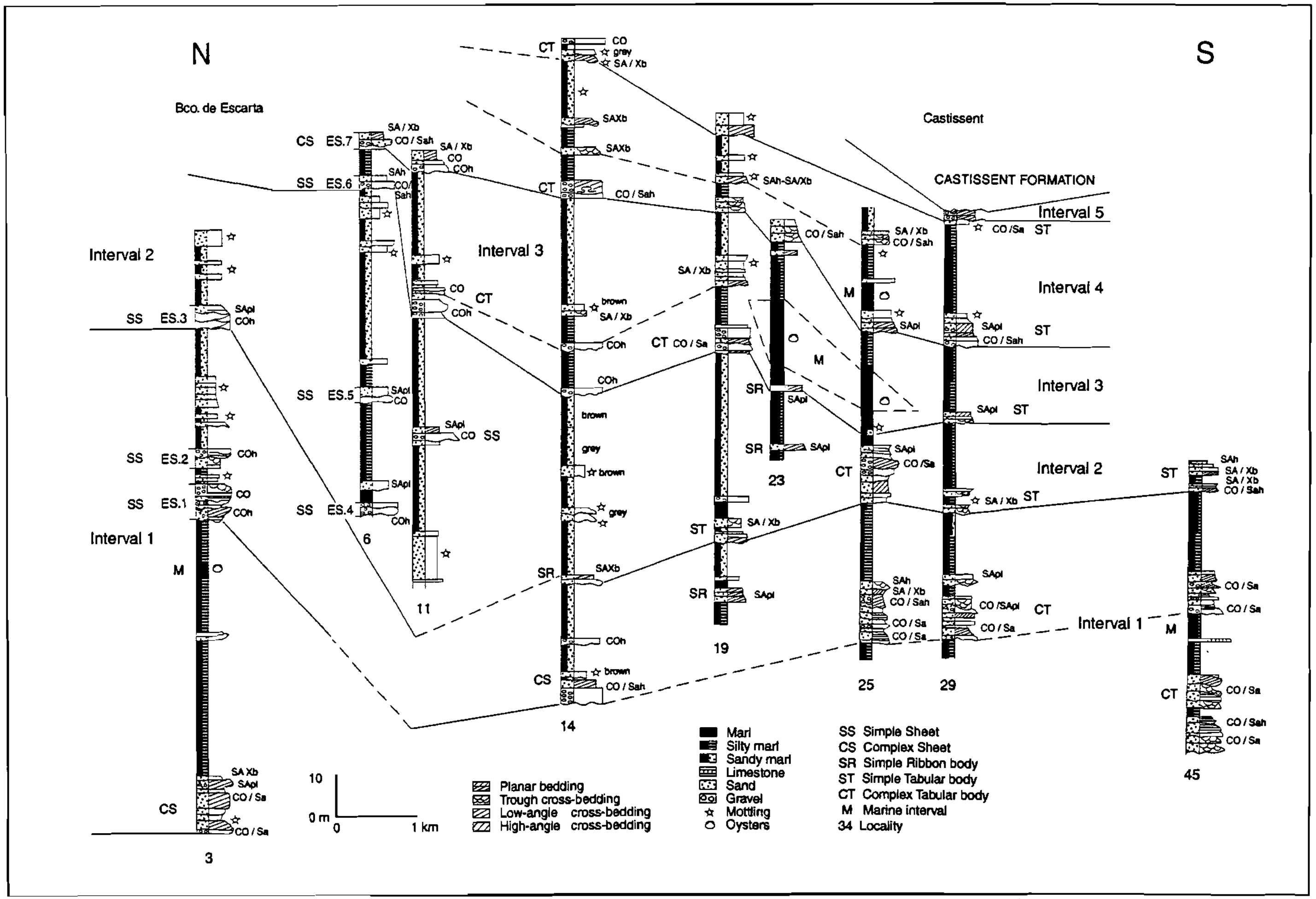


Figure 6. *Classification scheme of the various types of alluvial channel fills. Channel fill bodies are subdivided according to their geometry in sheet (S), ribbon (R) and tabular (T) bodies. In addition, they are further subdivided in simple bodies (S), referring to a single alluvial system, and in complex bodies (C). Complex bodies are built by more then one alluvial system, and resulted from lateral as well as vertical accretion.*

Figure 7. *See pages 95 and 96. Correlation chart of the measured logs along the western part (A) and the eastern part (B) of the Rio Noguera Ribagorzana valley showing the type of alluvial bodies. Legend for description of lithology is given in Table 1 (page 126). Localities of measured logs are shown in Figure 3.*


N
COLLEGATS FORMATION
WESTERN PART OF THE RIO RIBAGORZANA VALLEY
S
Interval 1
Interval 2
Interval 3
Interval 4
Interval 5
Puente de Montañana
Ager Group
33
34
35
37
39
40
41
42
43
44
COh
SApl
CO
CO/Sah
SAxb
SA/xb
FSi
Es. 1
Es. 2
Es. 3
Es. 4
Es. 5
Es. 6
Es. 7
SS
CS
ST
CT
M
MARL
SILTY MARL
SANDY MARL
MOTTLING
OYSTERS
SS Simple Sheet
CS Complex Sheet
ST Simple Tabular body
CT Complex Tabular body
M Marine interval
34 Locality
50m
0,5km


N
S
Bco. de Escarta
Castissent
CASTISSENT FORMATION
Interval 1
Interval 2
Interval 3
Interval 4
Interval 5
3
6
11
14
19
23
25
29
45
10
0 m
0
1 km
Planar bedding
Trough cross-bedding
Low-angle cross-bedding
High-angle cross-bedding
Marl
Silty marl
Sandy marl
Limestone
Sand
Gravel
Mottling
Oysters
SS Simple Sheet
CS Complex Sheet
SR Simple Ribbon body
ST Simple Tabular body
CT Complex Tabular body
M Marine interval
34 Locality

Changes in colour within individual profiles and across the basin indicate differences in environmental conditions during paleosol development. In general, the colour of soils is also influenced by the amount of organic carbon and the amount, the crystal size and the mineralogy of iron oxides or hydroxides. Red soils indicate relatively warm oxidizing conditions, and yellow or yellowish brown soils cooler oxidizing conditions (e.g., Kraus, 1992). Both are typical for low ground water levels. Greyish marls with ochre or reddish mottles reflect reducing soil conditions which are commonly related to relatively high ground water levels. Different soil types have been observed to grade into each other depending on the location on the delta plain or on the local morphology. In the upstream sector of the fluvial system, the floodbasin mudstones have more pronounced red colours with grey and violet mottles (Nijman & Puigdefàbregas, 1989). In the downstream sector where the fluvial system becomes distributive, calcretes disappear and colours become duller and eventually dull to dark grey.

In the reddish mudstone horizons, sheet-type channel fill complexes, pedogenically altered claystones, siltstones, and fine to very fine sandstones occur. These sediments are floodplain and overbank deposits. In several places in the central area small crevasse channel fills are present within the fine-grained deposits.

Coarse-grained alluvial and fluvial sediment bodies

The coarse-grained sediments of the lower Montañana Group have been studied by several authors (e.g., Van Eden, 1970; Nijman & Nio, 1975; Puigdefàbregas & Van Vliet, 1978; Van der Meulen, 1978, 1982, 1983, 1986, 1989; Atkinson, 1983; Nijman and Puigdefàbregas, 1989). Atkinson (1983) distinguished fluviatile channel fill sandstones and coarse conglomeratic channel fills. Nijman & Puigdefàbregas (1989) distinguished the channel fill bodies of the Montañana Group into ribbons, lateral accretion bodies and lenticular bedded bodies. Van der Meulen (1989) further subdivided the coarse-grained units of the Monllobat Formation into: 1) kilometre-wide massive sheet conglomerates, 2) relatively small sheets of conglomeratic sandstones, 3) winged, symmetrical channel fills (conglomerates or sandstones), 4) foreset lobes (conglomerates or sandstones and siltstones), 5) meander lobes (sandstones-siltstones) and 6) decimetre-thick coarse-grained sandstone sheets.

SEDIMENTARY FACIES

In the research area, low and high-sinuosity channel fills, conglomerate and sandstone sheets can be distinguished (Figure 6). The different coarse clastic bodies, of which correlations are shown in Figures 7a and 7b), can be divided into single bodies and complex bodies. The single bodies consist of ribbon-like low-sinuosity channel fills or sandstone sheets (BS), tabular-like high-sinuosity channel fills (CS), and conglomerate sheets (AS). The complex, composite or multi-storey bodies comprise combinations of several low-sinuosity channel deposits cutting into each other (BC), amalgamating low and high-sinuosity channel fills (CC), and combinations of conglomerate sheets, sandstone sheets (CS) and low-sinuosity channel deposits (ABC). Figure 6 gives an overview and an illustration of the coarse-grained facies types. The individual channel fills within a complex body are called a storey (cf. Friend et al., 1979).

Below, the alluvial architecture, comprising the three dimensional distribution of facies with the emphasis on coarse-grained channel bodies, will be described and discussed.

Sheet-like alluvial bodies

Coarse clastic sediment bodies with a laterally extensive or blanket-like form and a width/depth ratio of more than 15 are called sheet bodies (cf. Friend et al., 1979; Atkinson, 1983). Sheet-like bodies are mainly the products of unchannelized alluvial deposition. Usually sheet-like deposits result from deposition within a single major, very broad depression with a width/depth ratio greater than 15 (e.g., Miall, 1985). Such sheet-like channel fills may have practically imperceptible channel margins, with a slope of a few degrees only (Miall, 1985). They can also be produced by a relatively rapid channel migration or channel switching, resulting in a lateral amalgamation of channel fills which together form a sheet-like deposit.

Sheet deposits have been divided into two types: simple sheets (AS.) and complex sheets (AC). The simple sheets are sandstone sheets (AS.1) or conglomerate sheets (AS.2). Combinations of a sandstone and a conglomerate sheet (AC.1), a low-sinuosity channel fill incised into a sandstone sheet (ABC.1) were recognized. Simple conglomerate sheets (AS.2) were found to be incised by vertically stacked ribbon-like, low-sinuosity channel fills (ABC.2).

Proximal alluvial fan deposits

Sheet-like conglomeratic bodies with a width of hundreds of metres generally have an erosional base with locally deep scouring. Occasionally they show some large-scale high-angle tabular cross-bedding. Generally they consist of massive disorganized conglomerates (COh) in their most proximal parts, gradually changing into more organized conglomerates (CO) showing vague planar bedding with distinct pebble imbrications. The upper part often consists of fine to medium-grained planar laminated sands (SApl) or homogenous sandy marls (Fsa). The vertical change in lithofacies is more or less abrupt. The thicknesses of these sediment bodies vary between 1.5 and 9 m. The pebble size may reach up to 25 cm. Apart from a fining-upward tendency, trends in the vertical sequence of sedimentary structures are not distinct.

These conglomeratic bodies (Fig. 8) are interpreted as high-viscosity debris flow deposits and or sheet flood deposits on proximal, alluvial fan areas. These deposits indicate a relatively high relief and an abundant sediment supply.

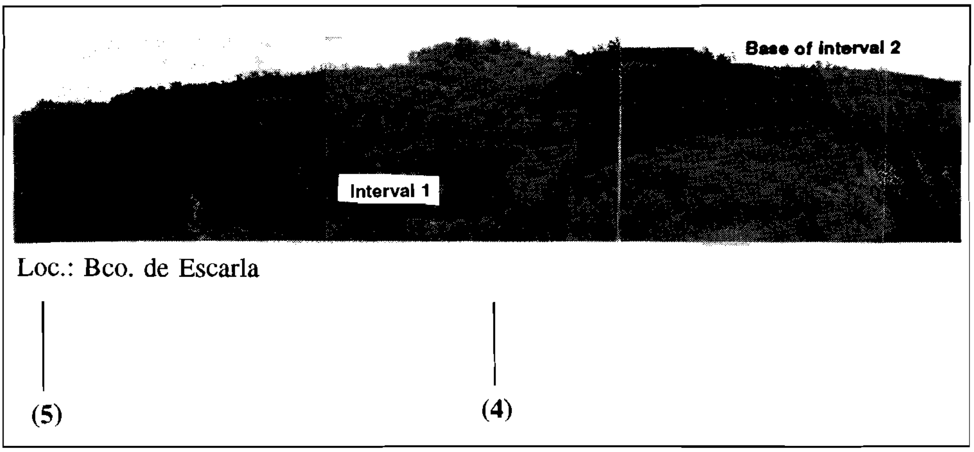


Figure 8. Photograph of the lithofacies typical for the proximal alluvial fan facies in outcrops along the Barranco de Escarla. Some major sheet-like coarse-grained alluvial bodies are visible.

Distal alluvial fan deposits

Sheet-like conglomerate and sandstone bodies consist of a basal layer of disorganized pebbly sandstone (CO/sah), followed by planar bedded and/or tabular or trough cross-bedded medium to coarse-grained sandstones (SAxb). Thicknesses vary between 1.5 and 6 m. The lateral extension may be hundreds of metres. The conglomerates grade upwards into siltstones and mudstones with plant remains. These bodies are interpreted as deposits formed on the distal parts of alluvial fans.

Ribbon-like alluvial bodies: low-sinuosity channel deposits

Ribbon-like sandstone bodies (B) have channels with width/depth ratios of less than 15 (cf. Friend et al., 1979). They can be divided into simple ribbons (BS) and complex ribbons (BC). Simple ribbons (BS) are interpreted as the result of deposition in low-sinuosity channels with a fixed channel position (BS.1) and intermediate-sinuosity channels (BS.2) with some lateral accretion. Complex ribbons are the result of unstable and laterally shifting channels producing laterally stacked channel fills by multiple erosion of former channel fills (BC.1).

In general, the low-sinuosity channel fills in the area consist of greyish-brown, moderately sorted, medium to coarse-grained sandstones with conglomeratic basal parts. These channel deposits have a distinctly erosional base. The basal part of the channel successions consists of a disorganized conglomeratic interval and brown, very coarse sandstone (CO/sah) (Table 1, page 126) with sometimes indications of primary sedimentary structures, such as low-angle cross bedding (CO/sa). A distinct fining-upward can be observed into planar bedded medium-coarse sands (SApl), followed by trough cross-bedded medium sands (SAxb). Gravel intervals show low to high-angle cross-bedding, even-bedding and even-lamination. The thickness of the channel fills usually is between 1.5 and 10 metres. At some places, (Fig. 12) a thickness of more than 15 metres is reached. The average thickness of the ribbon-type sandstone bodies observed is about 6 m. The upper parts of these channel fills are characterized by moderate to strong internal erosional surfaces. These deposits are interpreted as low-sinuosity channel fills formed in the distal parts of the alluvial fans along the northern margin of the basin.

Tabular alluvial bodies: high-sinuosity channel deposits

Tabular-shaped sandstone bodies in the Lower Montañana Group along the Rio Noguera Ribagorzana always show evidence of lateral accretion in the form of well-developed lateral accretion surfaces. They are the result of lateral accretion during long-term shifting of high-sinuosity channels.

Lateral accretion sets form at a high angle to the down-channel palaeocurrent direction and commonly exhibit a grain-size decrease upwards. Simple (CS) and complex tabular sandstone bodies (CC) have been distinguished. Simple tabular bodies consist of combinations of vertically and laterally stacked low-sinuosity channel fills (CS.1), or low- and high-sinuosity channel bodies (CS.2). Complex tabular sandstone bodies consist of combinations of sandstone or, rarely conglomerate and sandstone sheets, which are covered by high-sinuosity channel deposits. This combination is called the CC.1 type. The CC.2 type is characterized by laterally stacked low- and high-sinuosity channel deposits incised within a conglomerate sheet at the base.

Good exposures of the high-sinuosity channel fills are found in the Barranco del Pont de Montañana and Barranco de Tercuy (Figs. 9 and 10). The scoured surface at the base of the channel fills is locally overlain by patches of gravel forming a lag deposit. Lateral accretion surfaces as a result of lateral point bar migration are well developed in sandstone bodies which are intercalated in a muddy succession. The channel sandstone bodies generally show a well developed fining-upward trend. A basal organized conglomerate (CO/sa) (see Table 1, page 126) is overlain by trough cross-bedded and low-angle medium-grained sandstone (SA/xb). The top of the channel fills consists of mottled homogenous fine-grained sandstones often with calcrete horizons (SAh). On average the bodies are 3-5 metres thick and several tens of metres wide and show a palaeocurrent direction to the west.

LATERAL AND VERTICAL TRENDS

The studied succession consists of a cyclic repetition of coarse and fine-grained deposits. Within the sequence in the Castigaleu and the Monllobat Formations, five laterally consistent intervals with genetically related lithofacies have been recognized (Fig. 11). Each of these five intervals represents a depositional cycle with a comparable vertical and lateral facies development.

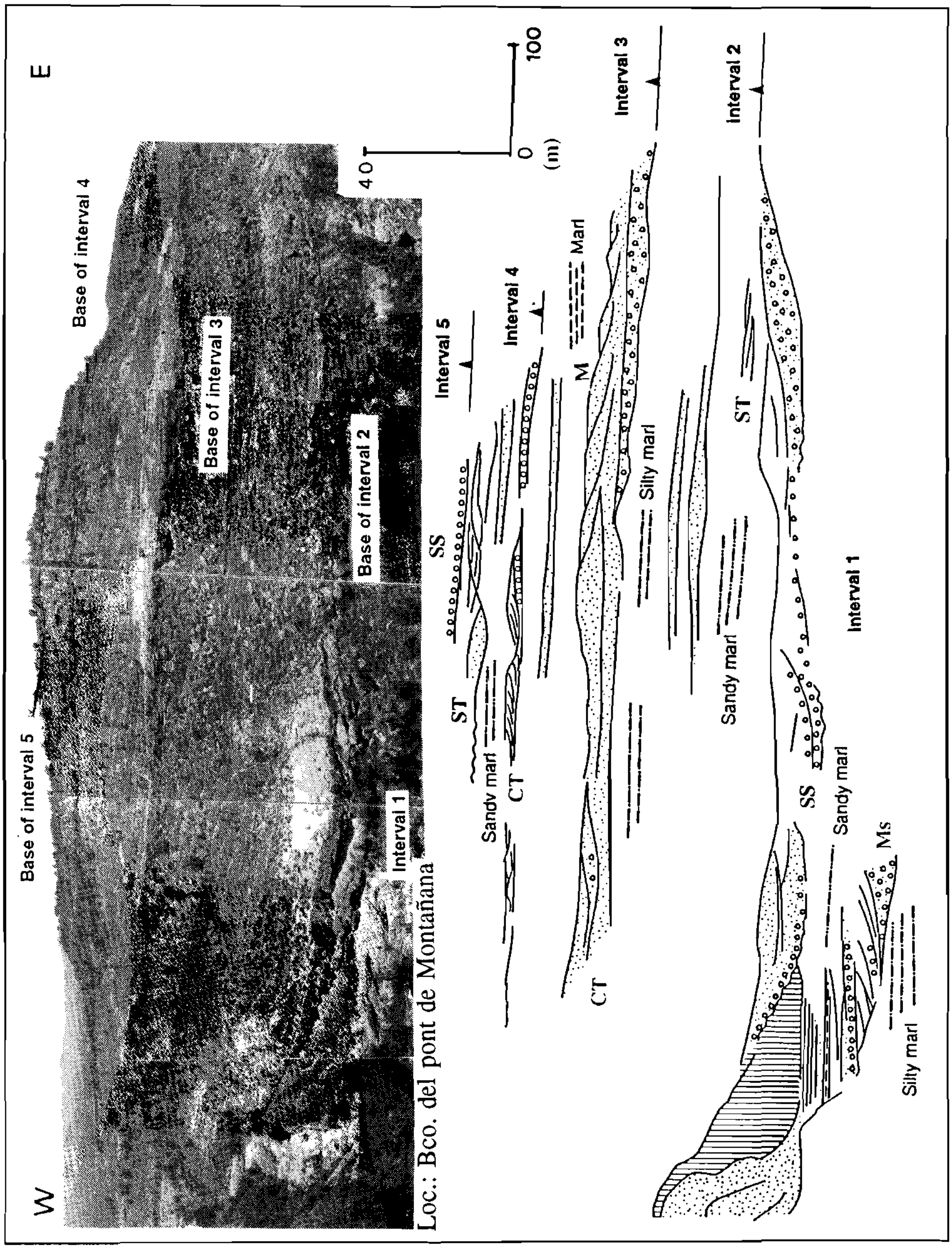


Figure 9. Photograph of the lithofacies in the central part of the studied area showing the distal alluvial fan facies in outcrops along the Barranco del pont de Montañana.

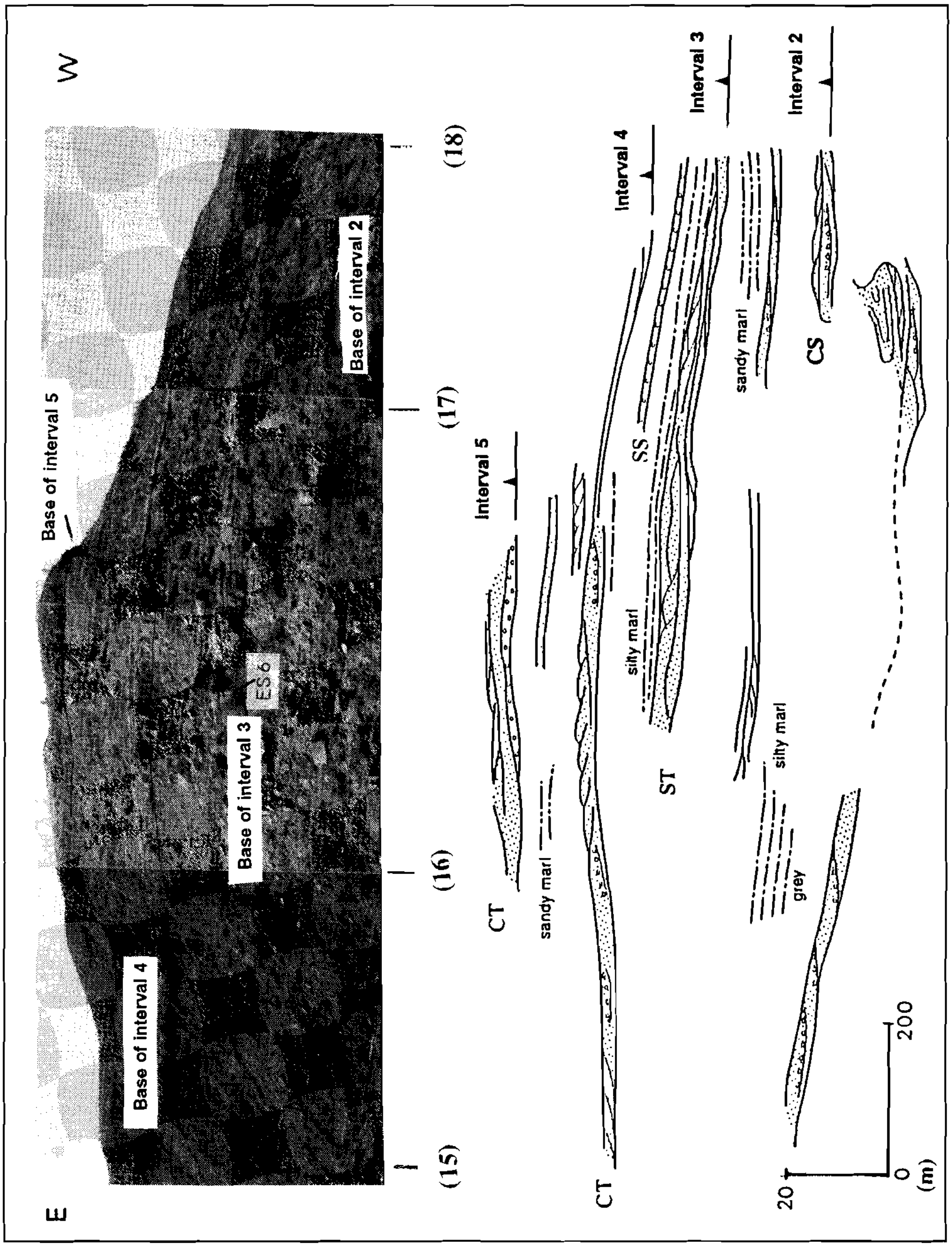


Figure 10. Photograph showing the major alluvial sediment bodies in outcrops along the Barranco de Tercuy (section localities are given in Figure 3).

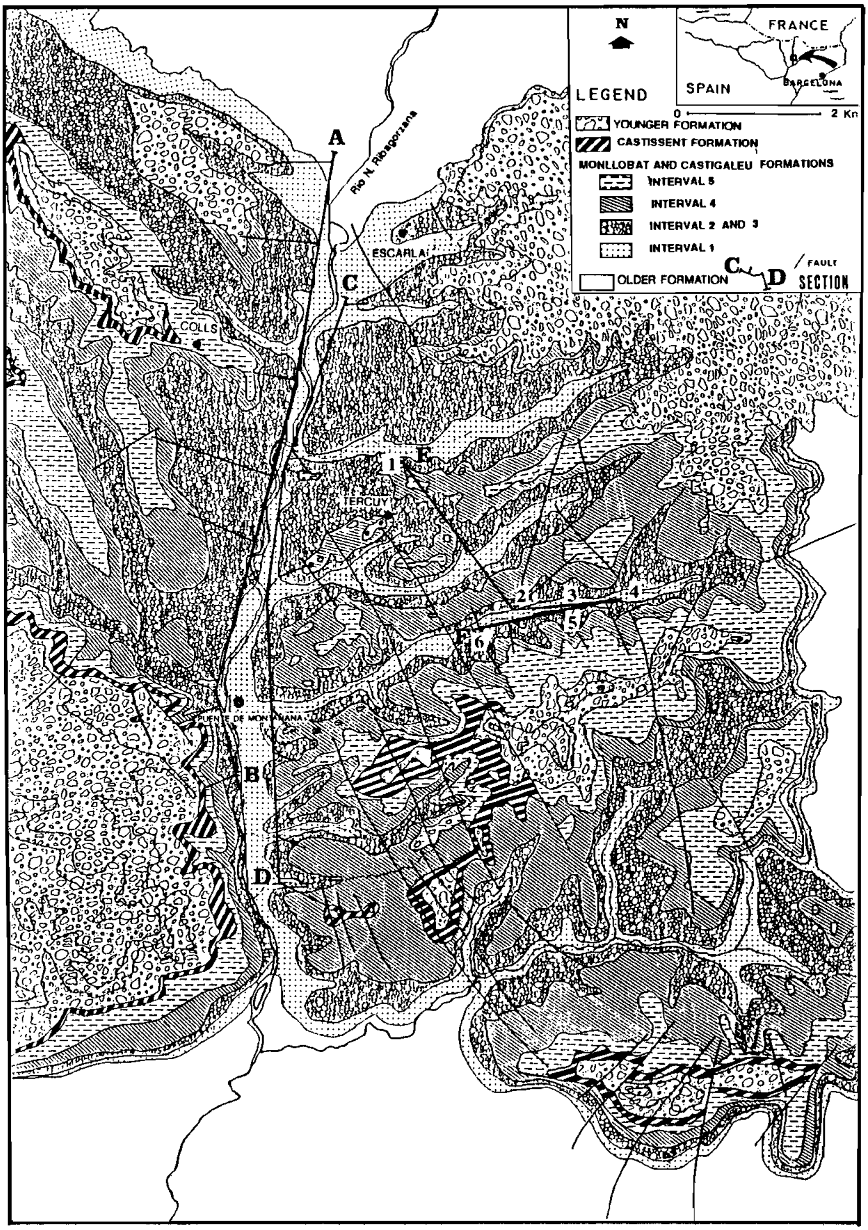


Figure 11. Geological map of the studied area showing the five intervals of the Lower Montañana Group. Locations of cross-sections (Figure 14) and of correlation chart of Figure 13 are indicated.

The lower boundary of each interval is formed by extensive conglomerate sheets. Several of these extensive sheets have been used for correlation between measured sections (Figs. 12 and 13). They have been numbered ES.1 through ES.8. These conglomerate sheets usually cover marine mudstone deposits in the southern area. Towards the top of an interval the channel sandstone bodies decrease in size and frequency, whereas floodplain mudstones become more abundant. From the sections it appears that the interval boundaries are erosional surfaces because sediment of underlying intervals has been eroded (Fig. 7). The intervals are thus commonly bounded at the top and at the base by slight unconformities or erosional surfaces (Fig. 7). For the whole succession, from base to top a decreasing marine influence is observed.

Towards the south, sheet-like coarse grained alluvial bodies disappear and grade into ribbon-type and finally into tabular channel bodies with evidence of lateral accretion on pointbars. Proximal-distal changes on the alluvial fan and plain are evident: on the proximal alluvial fan deposition occurred in unchannelized streams or laterally stacked braided channels, which distally gave way to simple straight channel and high-sinuosity meandering channel deposition. In the more distal parts of the area, channels may show tidal influences. Tidal influence has not been found in the proximal fan deposits dominated by conglomerate and sandstone sheet bodies. The distal alluvial deposits occupying the basin centre may even interfinger with shallow-marine deposits. The interfingering provides a sensitive record of the local fluctuations in base level.

The observed repetitive vertical changes in alluvial architecture potentially contain information about tectonic activity, climate and relative sea level. The lateral differences of the alluvial facies within each interval in the north-south direction are expected to be mainly controlled by the tectonic relief. This will be discussed below.

Composite sections representative of the character of the series in the northern and the southern part of the area are given in Figures 4 and 5. The sections show that large amounts of fine-grained sediment occur in between of the coarse sandy and conglomeratic units, and that the relative amount of the fine-grained sediments increases towards the south.

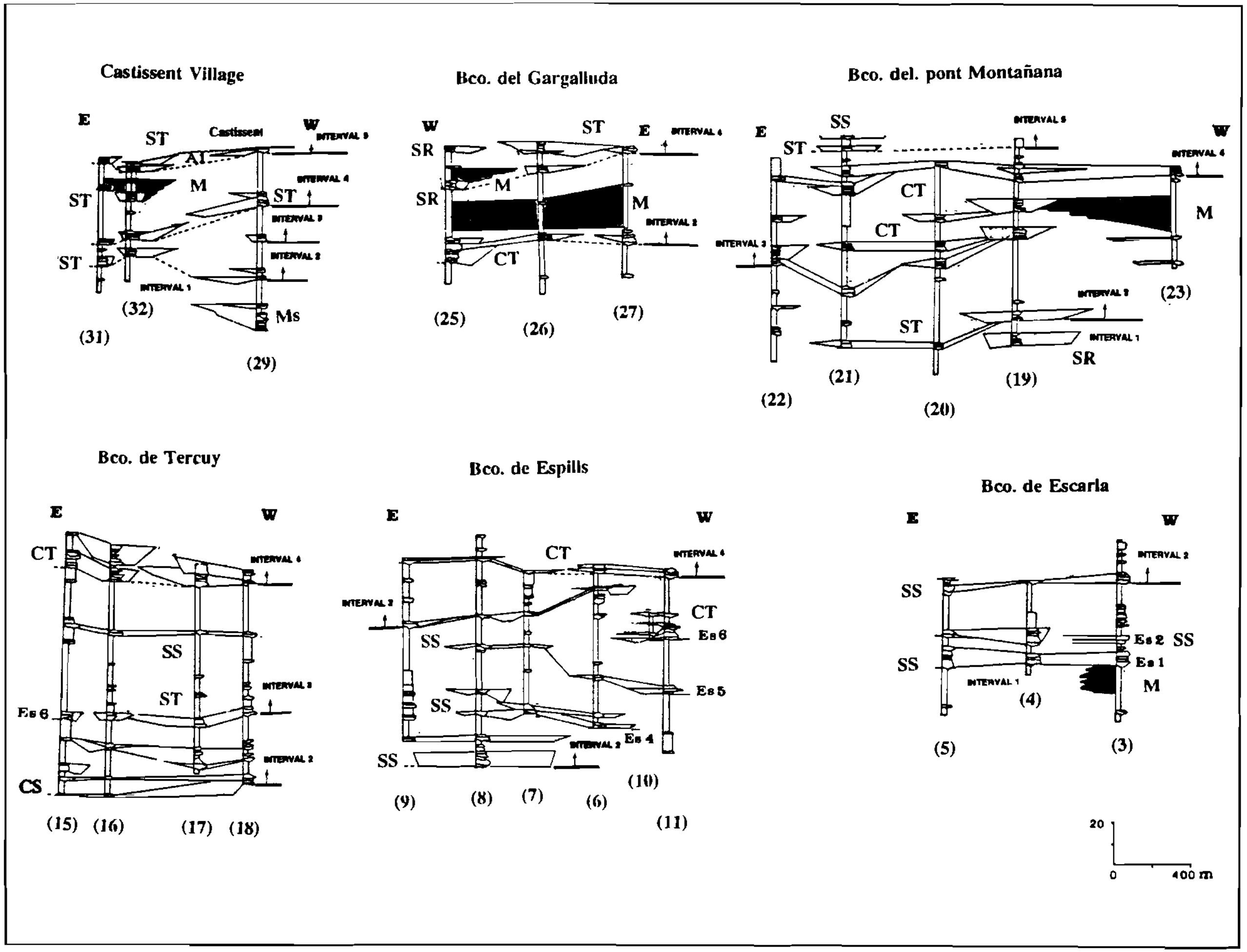


Figure 12. *Examples of measured logs and correlations of coarse-grained alluvial bodies in outcrops along several Barrancos. Localities are given in Figure 11.*

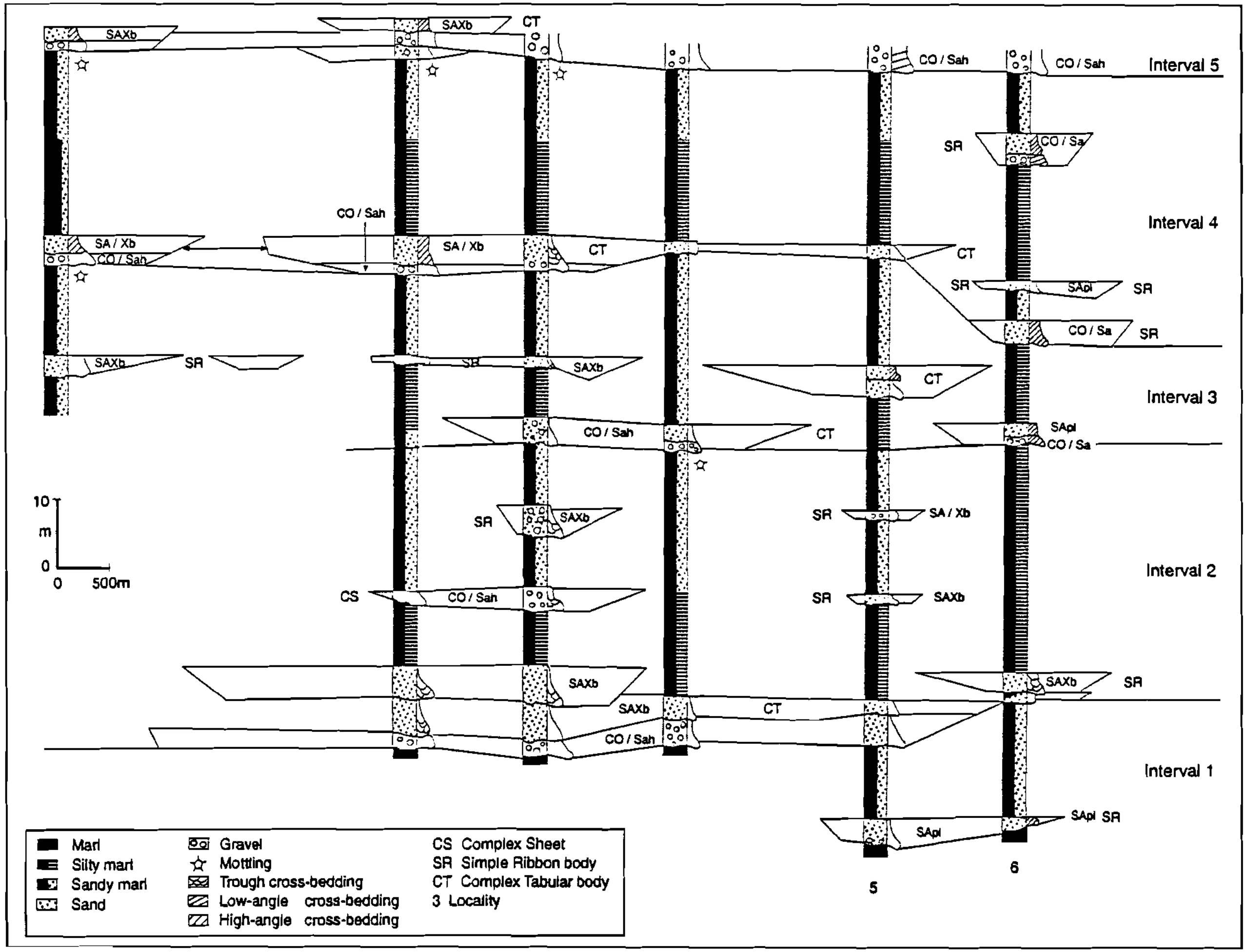


Figure 13. *Examples of measured logs and correlations of coarse-grained alluvial bodies in outcrops along several Barrancos. Localities are given in Figure 11.*

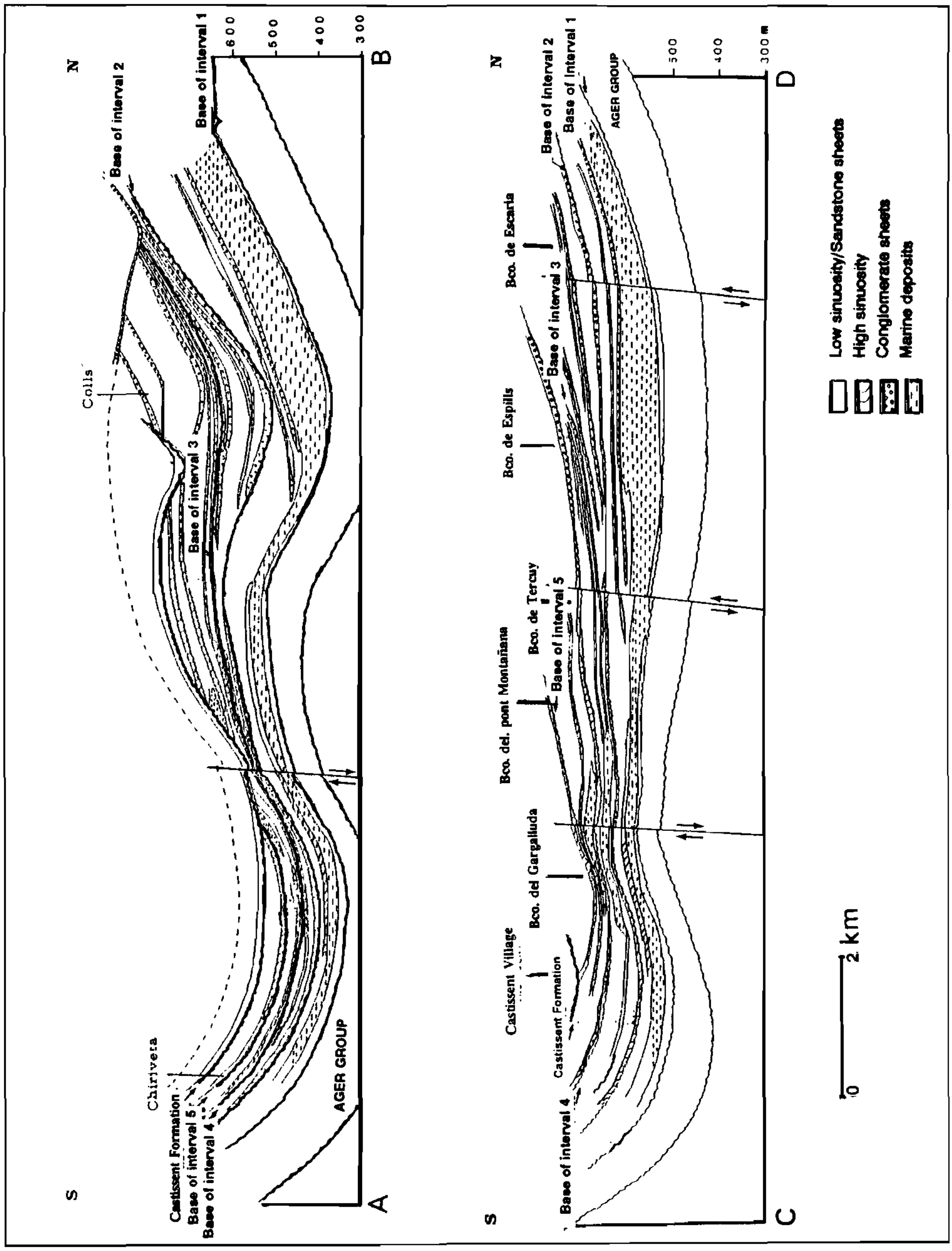


Figure 14. *(A) Cross-section along eastern side of the Rio Noguera Ribagorzana valley. The location of the cross-section is given in Figure 11. (B) Cross-section along western side of the Rio Noguera Ribagorzana valley. The location of the cross-section is given in Figure 11.*

Two N-S cross sections, east and west of the Rio Noguera Ribagorzana are shown in Figure 14. The profiles clearly illustrate that coarse-grained sediments were supplied through alluvial fan feeding systems from the N-NE and through an axial river system from the SE. The centre of the area was dominated by deposition of floodplain muds as well as marine muds (cf. Fig. 7). Measured thicknesses and isopach maps for the five successive intervals as well as for the overlying Castissent Formation are shown in Figures 15 and 16, respectively. They are based on the measured sections and interpolations between them.

The five intervals are described below.

Interval 1

The lower part of this interval consists of simple conglomerate sheets (AS.2) in the north, complex bodies (ABC.1) in the central area comprising sandstone sheets which are incised by low-sinuosity ribbon fills, which locally show tidal influence. In the south, multistorey tabular channel fills (CC.1) occur, which also show tidal influence. On top of these channel fills, fine-grained marine deposits are present. The marine sediments are mainly mudstones with abundant bivalve shells and some intercalated thin marine limestone layers.

The upper part of interval 1 consists of sheet-like deposits in the north and complex tabular bodies (CC.2) typical for high-sinuosity channels in the central and southern area.

Interval 2

The base of interval 2 is formed by a laterally extensive conglomerate sheet (AS.2) with an erosional base (ES.3). Laterally of this conglomerate sheet, tabular bodies (CC.2) occur in the central and western areas. These tabular bodies are composed of low and high-sinuosity channel fills, the latter with well-developed pointbar deposits. In the south, a small simple tabular body (CS.2) occurs. On top of the extensive conglomerate sheet (ES.3), a thin interval with mudstones as well as a sheet-like complex of conglomerate and sandstone bodies (AC.1) (ES.4) were found in the north. To the west these sheet-like bodies have low-sinuosity channel fills incised at the top.

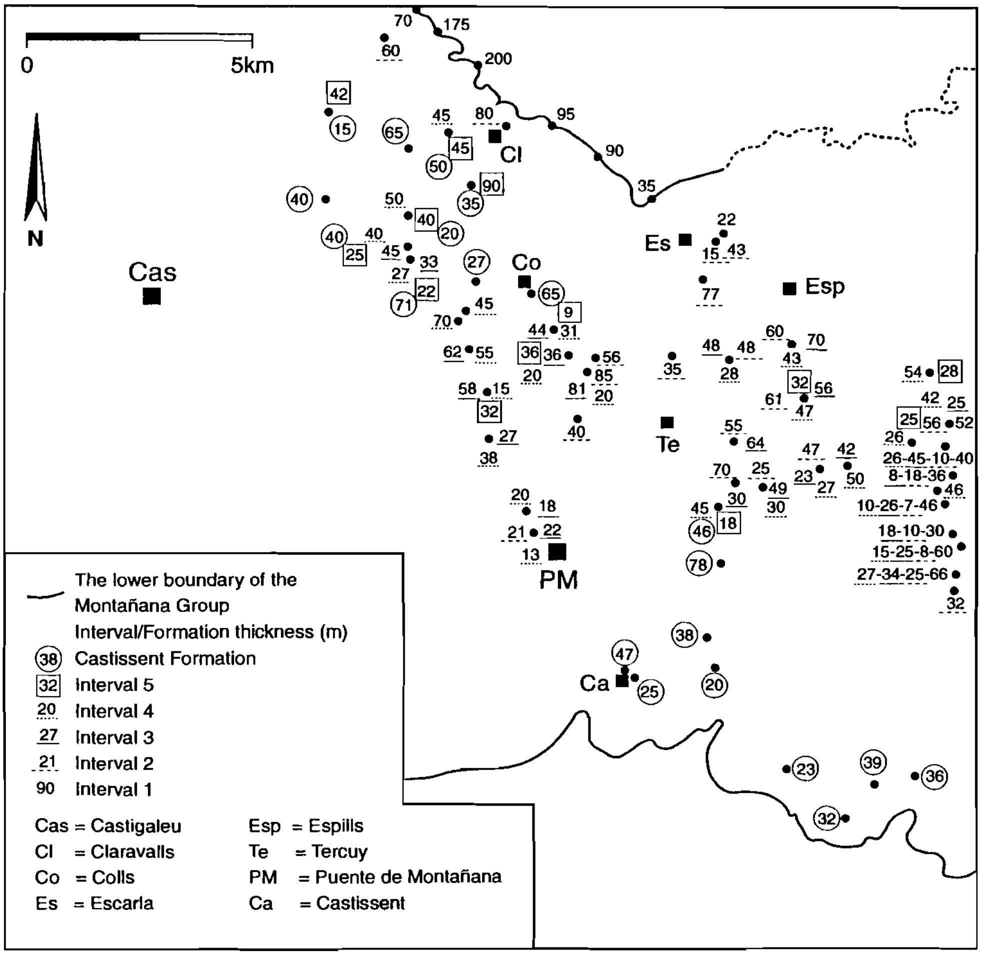


Figure 15. *Measured thicknesses of the various intervals in the Lower Montañana Group and the Castissent Formation.*

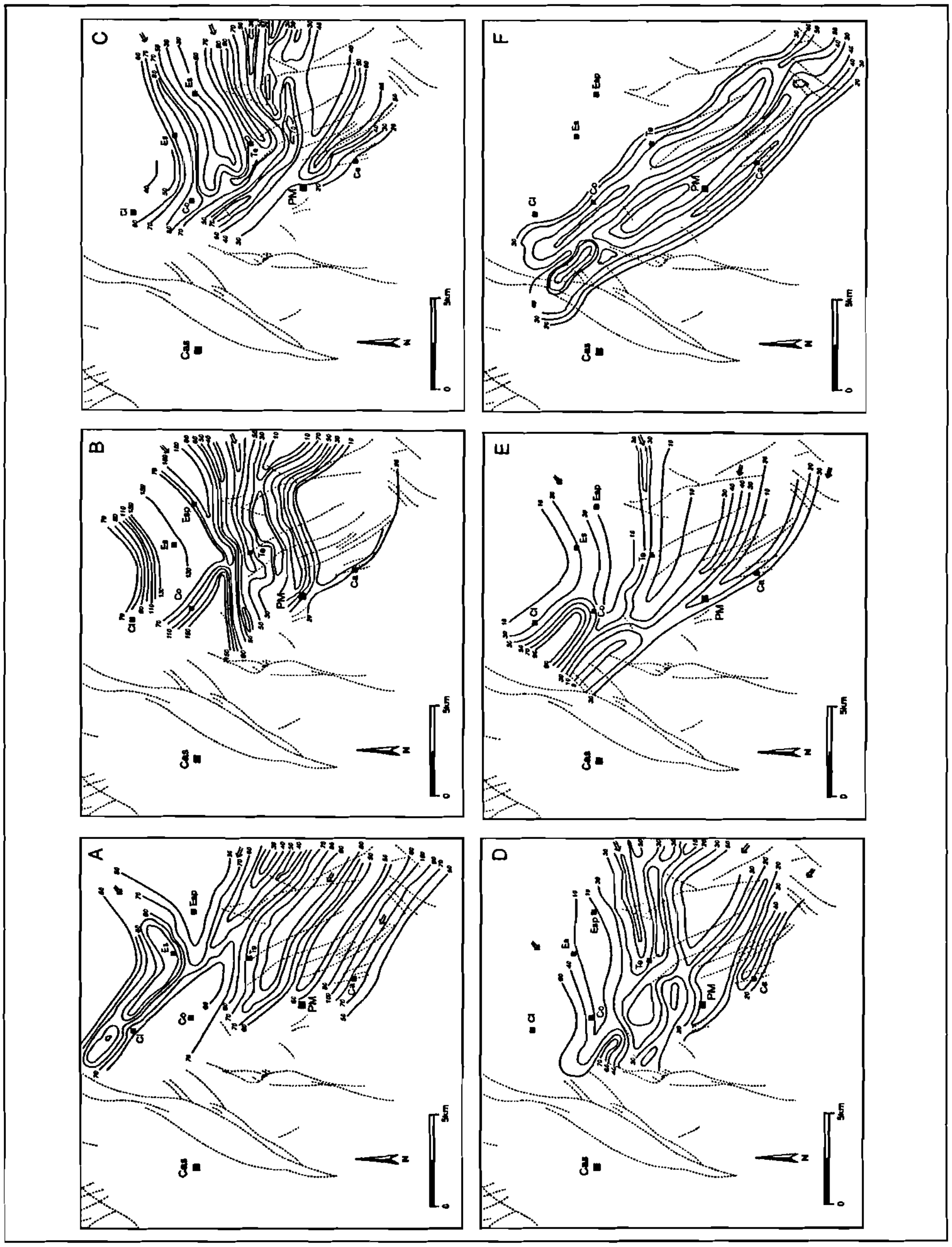


Figure 16. *Generalized isopach maps of the five alluvial intervals in the Lower Montañana Group. A: interval 1; B: interval 2; C: interval 3; D: interval 4: E: interval 5; F: Castissent Formation.*

The upper part of interval 2 is formed by a sheet-like conglomerate body (AS.2) in the north (ES.5) which develops into complex bodies (ABC.1) and low-sinuosity ribbon channels (BS.1) in the central area. In the south only mudstones were deposited.

Interval 3

The lower boundary of interval 3 is a sheet-like body (ES.6) which laterally grades into tabular bodies with low and high-sinuosity channels. Small simple ribbon- and tabular-like channel fills (CS.1 and BC.1) form the basal part of interval 3 in the south. Some of them show tidal influence. Marine mudstones occur on top of these channel bodies along the Barranco de la Gargalluda and Barranco del Pont de Montañana in the southern part of the studied area. The upper part of this interval comprises complex bodies of basal conglomerate sheets and incised, stacked tabular channel fills (CC.2) in the north, whereas in the central area, complex sheet and ribbon bodies occur.

Interval 4

The lower part of this interval is formed by sheet-like deposits with widespread erosion (body ES.7). On top of these, high-sinuosity channel bodies are present. The middle part of this interval is characterized by tidally reworked sandstones and a thin layer of marine mudstones.

Interval 5

In the north, interval 5 consists of laterally extensive conglomerate sheets (AC.1). Complex tabular bodies (CC.1) are found in the south. These tabular bodies belong to the axial fluvial system, with sediment supply from the southeast.

Lateral trends

From the measured logs and the cross-sections a clear change in alluvial bodies appears. In the proximal area sheet-like coarse-grained bodies dominate, especially at the base of each interval, whereas in the distal areas tabular bodies and associated fine-grained floodplain deposits occur. This reflects a change from braided to meandering alluvial processes. In addition the depocentre shifted towards the south (Figs. 7, 14 and 16). Marine fine-grained intercalations are found in the distal parts of the alluvial plain of some intervals. In successively younger intervals, these marine intercalations occur increasingly more southward.

SEDIMENT SUPPLY AND PALAEOCHANNEL TRENDS

Current directions have been measured on pebble imbrications and cross-bedding. Figure 17 shows the rose diagrams of the palaeocurrent measurements in the different intervals. Figure 18 shows the geographic distribution of palaeocurrent directions for each interval.

Two sediment supply systems existed in the research area. The first system comprises northern derived coarse-grained material deposited on alluvial fan systems. In the proximal area this resulted in the deposition of sheet-like alluvial systems. Towards the south, in the more distal part of the alluvial fan system, simple ribbon-type channel fills dominate.

The second clastic sediment source was from the southeast along the more or less axial fluvial system. Both supply systems joined and interfered in the central part of the research area and from that point they go straight to the west to northwest.

In the northern area, the thickness of the sedimentary column is greater than in the south (Figs. 14 and 16). In the central area between Barranco de Espills and Barranco del Pont de Montañana shallow marine deposits were formed in coastal plains.

From the alluvial facies and palaeocurrent patterns it is evident that the northern fed alluvial system was the dominant depositional system in this area.

Based on the thickness patterns (Fig. 16) and palaeocurrent directions (Figs. 17 and 18) of each interval four major palaeovalleys can be distinguished. The four palaeovalley systems were differently active during deposition of each interval. The main sediment supply along palaeochannel 1 was from the SE and pertains to the axial fluvial system. This palaeochannel is located between Castissent village and Barranco de la Gargalluda. Palaeochannels 2, 3 and 4 belong to the alluvial fan systems with sediment supply from the NE.

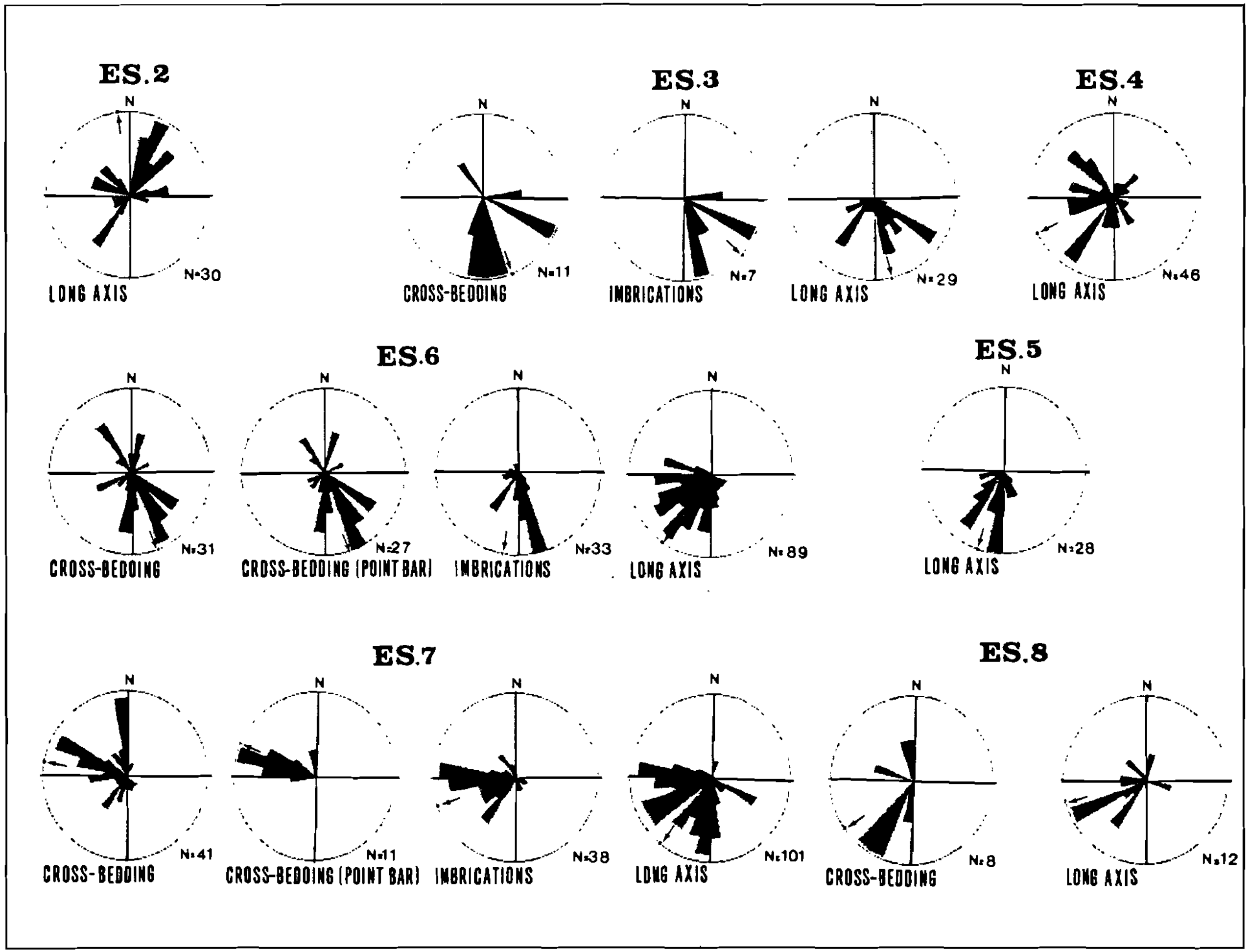
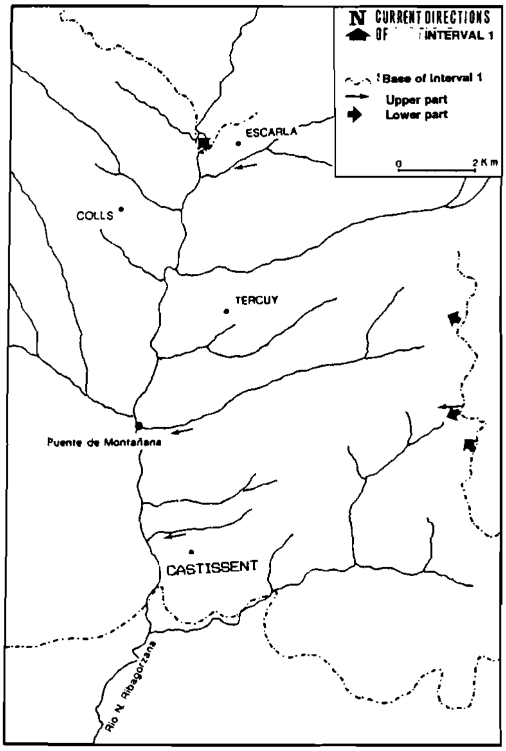
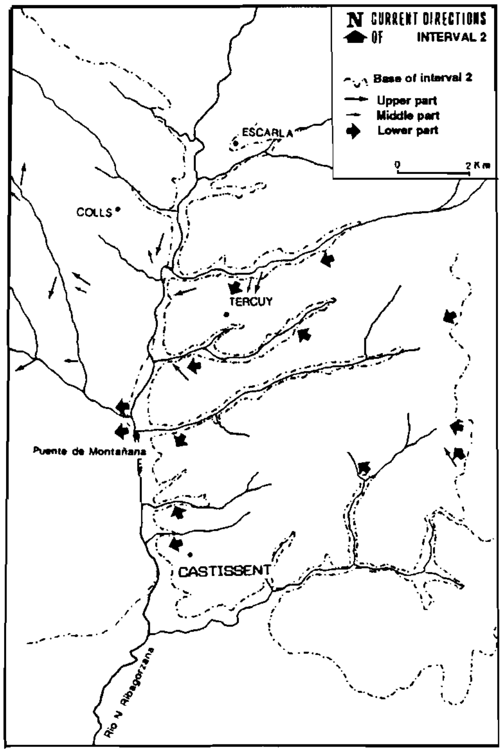
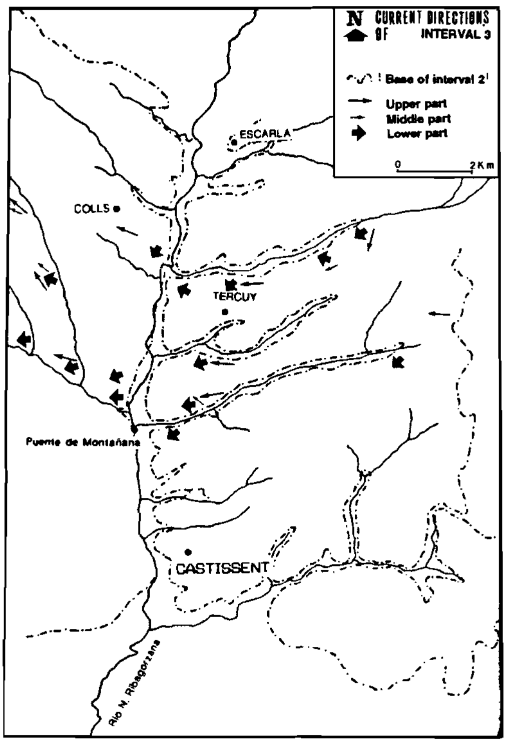
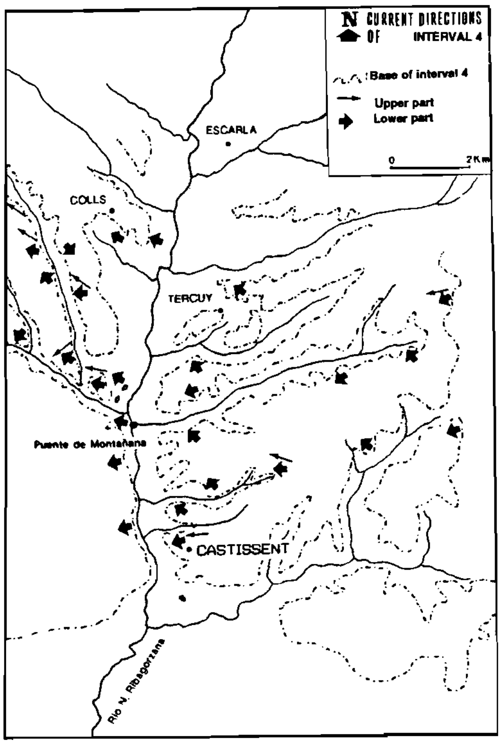


Figure 17. *Palaeocurrent rose-diagrams of some of the major sediment bodies ES.2 through ES.8.*


N
CURRENT DIRECTIONS
OF INTERVAL 1
Base of interval 1
Upper part
Lower part
0
2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana


N
CURRENT DIRECTIONS
OF INTERVAL 2
Base of interval 2
Upper part
Middle part
Lower part
0
2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana


N
CURRENT DIRECTIONS
OF INTERVAL 3
Base of interval 2
Upper part
Middle part
Lower part
0
2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana


N
CURRENT DIRECTIONS
OF INTERVAL 4
Base of interval 4
Upper part
Lower part
0
2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana

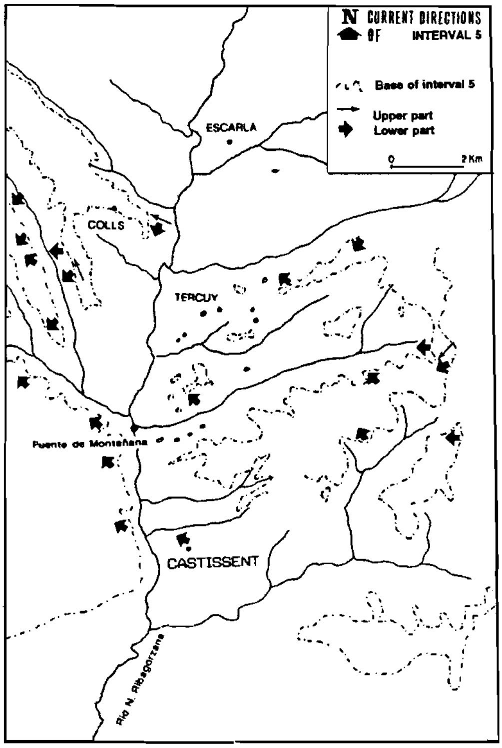
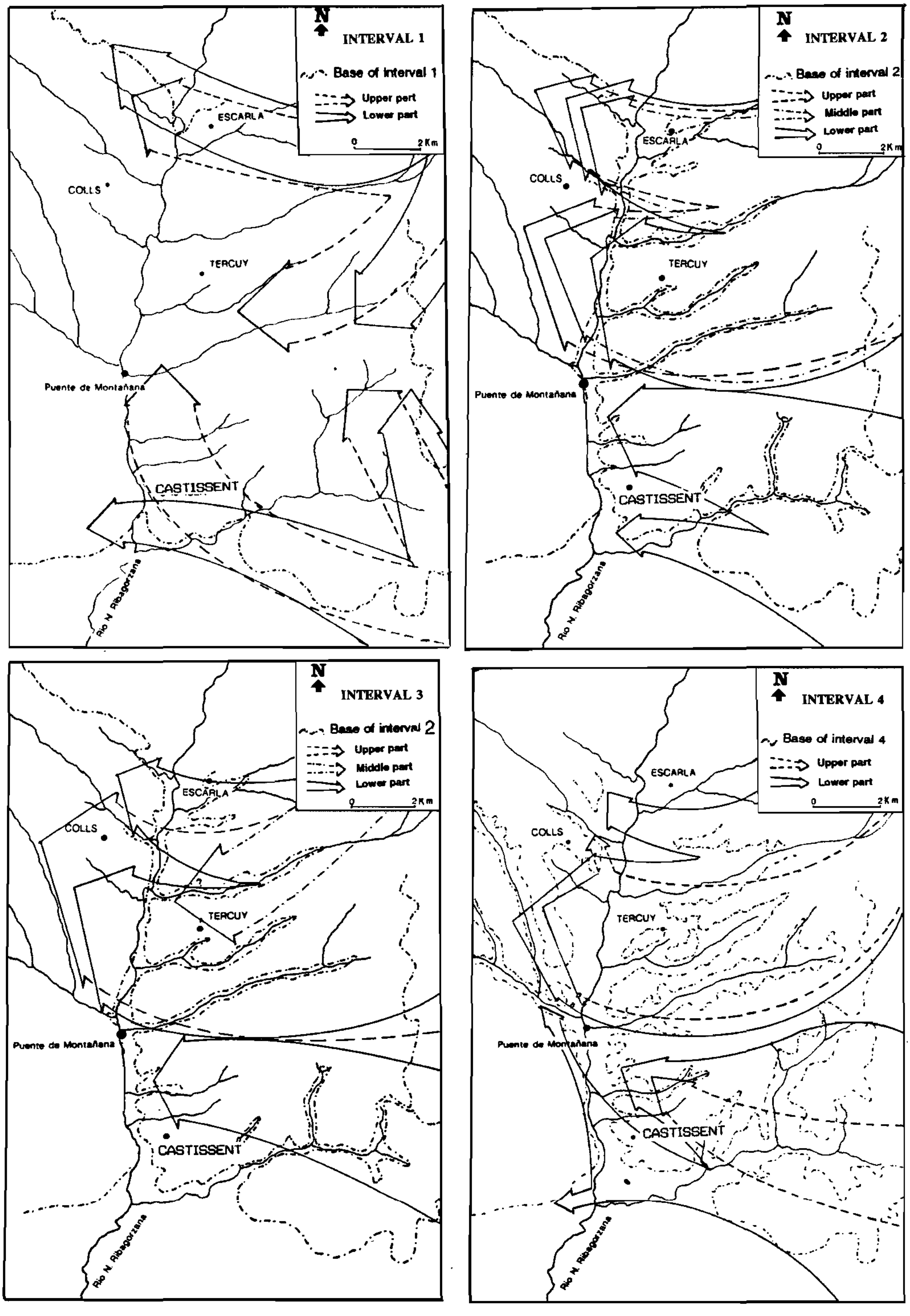


Figure 18. See pages 115 and 116. Geographic palaeocurrent pattern of the five intervals of the Lower Montañana Group.

The orientation of the local structures agrees with the general trends of the palaeochannels (Fig. 19). This suggests that the palaeochannel trends were influenced by the tectonic structures in this area. Especially in the upper part of the Lower Montañana Group, lateral movements along an inferred strike-slip fault along the Rio Noguera Ribagorzana valley is thought to have influenced the fluvial channel directions. Palaeochannels 1, 2 and 3 there changed direction across this assumed strike-slip fault along the Rio Noguera Ribagorzana valley.


N
INTERVAL 1
Base of interval 1
Upper part
Lower part
0 2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana
N
INTERVAL 2
Base of interval 2
Upper part
Middle part
Lower part
0 2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana
N
INTERVAL 3
Base of interval 2
Upper part
Middle part
Lower part
0 2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana
N
INTERVAL 4
Base of interval 4
Upper part
Lower part
0 2 Km
ESCARLA
COLLS
TERCUY
Puente de Montañana
CASTISSENT
Rio N. Ribagorzana

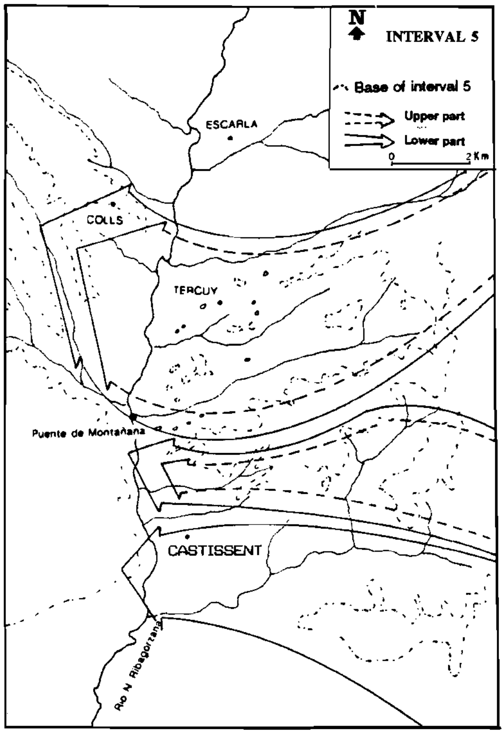


Figure 19*. See pages 117 and 118. Generalized interpreted palaeochannel trends in the five intervals of the Lower Montañana Group.*

DISCUSSION

Cycles

A recurring feature in the succession is the occurrence of fining upwards intervals consisting at the basis of a sheet-like conglomeratic/coarse sandy unit of braided fluvial origin, and at the top of a finer grained sandy unit composed of laterally extensive lateral-accreted tabular bodies with thick floodplain deposits. During deposition of each interval fluvial processes thus changed systematically from braided fluvial systems to high-sinuosity fluvial systems (Fig. 7). Such successions reflect a general decrease of the grain size of the sediment and a concurrent increasingly more regular fluvial discharge. Each alluvial interval, moreover represents a small-scale progradational cycle of alluvial fan activity along the northern margin of the basin.

The large-scale cycles in the Montañana Group have been ascribed to periodic enhanced tectonic activity and coupled temporary increased clastic supply (Atkinson, 1983; Van der Meulen, 1989).

The stratigraphic succession in a fossil sedimentary basin is the long-term response of the depositional system to prolonged subsidence. Recent developments of sequence stratigraphical concepts have resulted in major changes in the approach to the analysis of regional facies associations. The current sequence stratigraphical approach, which stresses the importance and the effects of global sea-level changes, is applied by many authors. This approach is based on the Exxon sea-level curve (Haq, Hardenbol & Vail, 1987, 1988; Vail, Mitchum & Thompson, 1977). Third-order cycles represent the highest frequency sea-level events portrayed on the Exxon curve. Such eustatic sea-level changes are often considered to be the principle factor producing cyclic repetitive stratigraphic sequences. Indeed they form often the dominant control on sedimentary processes and sedimentation patterns along passive margins (Vail et al., 1977; Posamentier & Vail, 1988). Sea-level change was perhaps the first factor recognized as having the capacity to produce repetitive stratigraphic sequence, and numerous authors still consider eustasy as the dominant control on sedimentary processes and sedimentation patterns. However, tectonics and climate, although having a predictable effect on clastic deposition, are frequently ignored in discussions about the origin of rhythmic clastic sedimentary successions.

Tectonics

In many cases, the geometry and timing of alluvial clastic wedges and associated coastal plain deposits can be ascribed to tectonic events in the source area(s) (Miall, 1992). In his opinion, the alternation of regressions and transgressions can well be driven by variations in sediment supply and depositional slope, and thus be the result of changes in the depositional regime.

In the north and south of the studied area, the sandstones and conglomerate layers frequently are amalgamated, whereas in the central area alluvial channel bodies occur isolated within floodplain fines. The axial fluvial system, related to palaeochannel 1 (Fig. 19), is restricted to the southern part of the basin, pointing to the continuous existence of an alluvial palaeoslope towards the south. Two cross sections from the western and eastern Rio Noguera Ribagorzana valley (Fig. 14) show the differences in sediment accumulation. Subsidence in the north was much greater than in the south. The thickness pattern, i.e., the thinning of intervals towards the south, and the erosional surfaces indeed indicate major subsidence in the northern part of the basin. This suggests syntectonic activity of the thrust systems, which formed the basin boundaries, likely related to growth of the orogenic wedge (cf. Muñoz, 1991).

The differential accumulation of sediments along the Rio Noguera Ribagorzana during the lower Eocene reflects, and is obviously related to the structural development of the basin. From the thickness patterns and the facies distribution, it can be inferred that tectonic induced subsidence, rather than a rise of the sea level, was the main factor creating the sediment accommodation space. Therefore it also must have largely controlled the alluvial architecture. Tectonic activity is reflected in movements along the Boixols thrust system in the north and along the Montsec thrust in the south (i.e., the Montsec anticline), where anticlinal doming was active (Muñoz, 1991). The thrust movements controlled the general vertical development of the succession as well as the lateral facies patterns. Other structures, such as local faults and folds, constrained the position of the four palaeochannels. However, as will be discussed below, climatological and eustatic effects may have been superimposed on the tectonic signal.

The development of the basin and the basin fill in the research area was influenced by several tectonic features, which are summarized below.

- The Boixols thrust sheet was not active during deposition of the lower Montañana group in the Upper Ypresian, but caused subsidence by thrust wedge loading during the Lower Eocene; possibly it was reactivated during deposition of the Collegats Formation during the Oligocene (Puigdefàbregas et al., 1989).

- Activity along the Montsec thrust and the development of the Montsec anticline in the south constrained the position of palaeochannel 1 (Fig. 19), which developed when the rate of updoming decreased. This is confirmed by the Ypresian age of the Montsec thrust in the ECORS traverse, as recorded by the syntectonic Ypresian sediments at both sides of the frontal thrust ramp (Puigdefàbregas et al., 1989). The Montsec anticline was active during deposition of intervals 1 through 4. During development of interval 5 and the Castissent Formation, large tabular bodies were formed belonging to the axial fluvial system supplied from the southeast. This indicates that the Montsec anticline at that time did not form a prominent positive relief.
- local synsedimentary developing folds constraining the palaeochannel trends.
- lateral movements along the inferred strike-slip fault along the Rio Noguera Ribagorzana, which tended to offset gently developing growth folds, similar to those discussed by De Boer et al. (1991) along the Esera river, with a deflection of palaeochannels on both sides.

The style of the alluvial facies, i.e., the type of channels and of the alluvial deposits, is primarily the result of the palaeoslope on the alluvial fan and its lateral and temporal changes. As discussed above, this can be ascribed to tectonic control on the basin evolution including the shape of the basin and the subsidence. This does not, however, explain the rather regular cyclic nature of the progradational cycles of the succession.

Of course, periods of accelerated tectonic activity could have caused temporary increased alluvial fan progradation. Tectonic activity also would cause steepening of the depositional slope. During such periods increased supply of gravel and sand, eroded from the relatively high, rejuvenated relief in the source area to the north, would have caused fan progradation followed by retrogradation during successive quiet periods. However, up to now no mechanisms have been described in the literature which could have produced the rather regular patterns observed in the research area.

The profiles show a gradual migration of the depocentre towards the south, and the succession of the Lower Montañana Group as a whole is a progradational system. The stacking of successive alluvial intervals and the depocentre migration both point to a more or less steady, tectonically induced subsidence of the basin and a constant creation of sediment accommodation space.

Climate

Apart from the palaeoslope, the channel type also depends on the type of discharge (regular versus irregular) and the type and amount of clastic material being transported. On alluvial fans, the discharge tends to become more regular in the downstream direction. This favours channelization and eventually meandering alluvial systems. The effect of this enhances that of the decreasing slope of the alluvial fan on the alluvial channel type.

In proximal areas and in systems with relatively small drainage basins, changes in climatological conditions are directly noticeable in river discharge and thus in type of alluvial transport and style of channels. Climate thus controls sediment supply, which is obvious if recent weathering and erosion patterns in different climate zones are considered.

A systematic change in the type of alluvial channels could be the result of changes in sediment supply and the ratio between coarse-grained bedload and suspended load. Large amounts of conglomeratic bedload tend to produce braided-type alluvial channels. The sheet-like braided systems at the base of the cycles thus indicate a sudden release of large amounts of coarse-grained material. More suspended load and regular discharge thus would have favoured the development of meandering-type river deposits in the top of the cycles.

Climate has been recognized as an important and independent force in controlling weathering, erosion and sediment supply to alluvial environments. Climate can cause cyclic changes in facies (cf. Olsen, 1989, 1990 and 1993). A direct result of global changes of climate are eustatic changes in sea level due to the growth and decay of large continental ice sheets (Plint, 1991). However, during the Eocene such ice sheets did not or hardly existed. On a smaller scale, weathering and erosion in the source area as well as the transport of terrigenous sediment depend largely on precipitation in the drainage area (cf. Perlmutter & Matthews, 1989).

Differences in the amount of clastic supply can be caused by changes in climatological conditions in the sense of alternations of relatively wetter and dryer periods with consequent changes in discharge and weathering intensity. Changes in the amount of clastic supply related to climatological variability can be expected to occur on a regional scale rather than on the scale of one alluvial fan only. The internal structure of individual fans potentially can be strongly influenced by autocyclic factors such as avulsion.

However, former research indicates that the alluvial cycles in the area near the Rio Noguera Ribagorzana valley are regional features (Van der Meulen, 1989).

Weathering, runoff and the regularity of runoff, transport mechanism and type of clastic supply are thus a function of climatological conditions in both the source area and the alluvial basin. The climate directly controls weathering intensity through the rate of chemical weathering and through the rate of erosion (Suttner & Dutta, 1986; Grantham & Velbel, 1988; Johnsson & Stallard, 1989). The type of weathering and its intensity determine the amount and type of clastic material available for transport, i.e., bedload, suspended load and dissolved matter. The transport mechanisms, fluvial transport versus mass-transport mechanisms, also are directly related to the runoff and the regularity of the runoff.

Changes in climatological conditions between more humid and more arid conditions can evoke cyclic changes in the supply of coarse clastic material and the transport mechanisms. The last factor influences the slope of the alluvial fan and the lateral extension of the deposits. Wet-type fans display lower gradients because fluvial processes dominate, which have larger transport capabilities than mass-flow processes. The latter dominate on relative arid alluvial fans, which consequently have higher slope gradients. A more humid climate, or large drainage basins with an eventual regular discharge, favours fluvial dominated alluvial fans with gentle slopes and a larger areal extent. This implies that cyclic changes in climate may cause cyclic progradational patterns in alluvial fans.

Pulses in tectonic activity, periodic changes in climate and global sea-level changes may influence the architecture of fans in terms of progradational and retrogradational patterns, reflected as coarsening and fining upwards sequences. However, the scale on which the sedimentary succession reflects these factors is different. The geographic position of the major climate belts, defined by the solar radiation over the Earth in combination with the distribution of continents, oceans and major mountain belts, changes due to periodic variations in the orientation of the Earth's axis and the orbit around the Sun (see review by De Boer & Smith, 1994). This results in so-called Milankovitch cycles in climatological conditions and thus in sediment supply. These cycles may induce periodic changes in sedimentation patterns with periods of approximately 20,000, 40,000, 100,000, and 400,000 years. Such cycles would represent high frequency 4th or 5th order cycles which could be superimposed on higher order, longer term tectonic and/or eustatic cycles.

The above mentioned observations strongly suggests a cyclic change in climatological conditions, ranging between relatively more arid and relatively more humid epochs (De Boer et al., 1991). During relatively arid periods, clastic material is produced by weathering in the source areas, but cannot be transported into the basin at the same rate as it is produced. On the other hand, at the onset of relatively humid periods, large amounts of clastic material are available which can be rapidly eroded and transported into the basin. At the same time chemical weathering becomes increasingly important; more intense chemical weathering causes the production of finer grained, monomineralic material (Suttner & Dutta, 1986; Grantham & Velbel, 1988; Johnsson & Stallard, 1989). At the end of the humid periods, only sparse fine-grained bedload material and, more importantly, suspended material is available for transport and deposition (e.g., Weltje, 1994). This results in a gradual change from sheet-like braided systems into meandering rivers and their well-developed fine-grained overbank deposits (Figs. 9 and 10). The vertical aggradation rate therefore tends to decrease during one cycle, and eventually the sea may transgress onto the lower alluvial plain due to the fact that subsidence is stronger than sediment supply. Such a transgression thus merely would be the result of the rate of sediment supply and the vertical aggradation of the alluvial plain would be lower than the subsidence rate.

Sea-level fluctuations

Relative sea-level changes affect marine and coastal depositional environments, their influence being most obvious in shoreline and shallow marine areas.

In the studied area, a discrete shoreline occurred. The shoreline could migrate towards the east and north when (local) sediment supply decreased. It is likely that over longer time spans (third order cycles) (local) sea level has changed due to thrusting events in the northern and southern areas.

Posamentier & Vail (1988), Posamentier et al. (1988), and van Wagoner et al. (1988) elaborated the effects which global eustatic sea-level changes may have on the character of coastal successions along passive margins.

In the above described succession tectonics clearly influenced the position of the shore line on a third-order scale. The limited time control does not allow to define if and to which extent eustatic sea-level changes added to the observed third-order cycles.

CONCLUSION

In general, alluvial fans typically occur in tectonic active settings and tectonism is therefore expected to be the main factor controlling their occurrence, architecture and preservation. In the Lower Montañana Group, tectonic activity was a major control on the sediment supply by uplift through antiformal stacking in the source area, and through basin subsidence by thrust-sheet loading. On a large-scale, the Lower Montañana Group represents a coarsening-upwards cycle, representing major progradation of the alluvial fan system. Within the sequence, smaller scale cycles can be distinguished, which represent smaller scale progradation-retrogradation cycles.

The amount of sediment available from the source area and slope gradients were important factors influencing the development of the alluvial succession and its geometry in the Tremp-Graus basin. Apart from the tectonically controlled relief, the architecture of the succession and its observed cyclic nature on a forth or fifth order scale is ascribed to cyclic changes in the climate. Especially the type of discharge and the sediment availability through weathering were important for the eventual alluvial style and architecture. The local position of the sea level is assumed to have been the resultant of the interplay between tectonic subsidence and climatologically controlled sediment supply.

Table 1

Lithofacies	Description
SAh	Sandstone; fine- to coarse-grained; homogeneous; badly sorted; sometimes bimodal grain-size distribution
SApl	Sandstone; fine- to coarse-grained; planar bedded or low-angle cross-bedded; moderate to well sorted
SAxb	Sandstone; fine- to coarse-grained; tabular or trough cross-bedded; well sorted
SAcoh	Disorganized pebbly sandstone; fine- to coarse-grained sandstone
SAco	Organized pebbly sandstone; fine- to coarse-grained sandstone; horizontal bedded or low-angle cross-bedded
CO/sah	Disorganized massive conglomerate with ±25% sandstone matrix
COh	Disorganized massive conglomerate with <10% sandstone matrix
CO	Organized conglomerate with <25% sandstone matrix; planar- or cross-bedded
CO/sa	Organized conglomerate with ±25% sandstone matrix; planar- or cross-bedded
Fm	Marl (> 50% mud)
Fsi	Silty marl; usually homogeneous
Fsa	Fine-sandy marl
/sh	bivalve shell debris
/mot	pedogenic mottling

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