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Depositional Sequence Response to Foreland Deformation in the Upper Cretaceous of the Southern Pyrenees, Spain[1]

Lluís Ardèvol,[2] Juan Klimowitz,[3] Jesús Malagón,[3] and Peter J. C. Nagtegaal[4]

ABSTRACT

During the initial stage of the evolution of the South Pyrenean foreland basin in the Late Cretaceous, foreland shortening resulted in the development of a major thrust sheet consisting of three main fold-thrust uplifts. As a consequence, the basin was segmented into uplift-bounded depocenters (minibasins) that were contemporaneously filled with basin-fill successions of prograding siliciclastics.

A sequence-stratigraphic study has led to detailed regional correlations showing turbidite, deltaic, and fluvial facies transitions (Vallcarga, Aren, and Tremp formations, respectively). The deltaic deposits are grouped into four depositional sequences based on recognition and definition of sequence boundaries and systems tracts.

The basinward migration of the depocenters was associated with uplifting and cyclic sedimentation. Turbiditic deposits accumulated on footwall synclines and onlapped onto growing anticlines during phases of active thrusting. Deltaic wedges prograded over deep-water sediments. Highstand shelves overlapped preexisting structural highs suggesting relative tectonic quiescence. The chronology of the shelves, based on planktonic foraminifera, provides new insight into the timing of deformation.

Three hydrocarbon plays are found within lowstand systems tracts: delta front, canyon fill, and slope channel fill. These may prove helpful as models for more prolific hydrocarbon-bearing clastic basins. The depositional model may serve as an analog for intraslope minibasins on Atlantic margins.

INTRODUCTION

In recent years the concepts of sequence stratigraphy, originally developed from passive margins, have been increasingly applied and tested in foreland basins (Van Wagoner, 1995). In this paper, we provide an example from a foredeep basin and demonstrate the influence of active thrusting on depositional sequence development in the Campanian–Maastrichtian Aren Sandstone. The superb and relatively undeformed exposures of this formation along the south-central Pyrenees (Figure 1) have been the subject of several sequence-stratigraphic studies (e.g., Simó and Puigdefàbregas, 1985; Simó, 1986, 1989; Díaz-Molina, 1987; Mutti and Sgavetti, 1987; Sgavetti, 1992, 1994; Ardèvol et al., 1993; Deramond et al., 1993; Rosell, 1994; Arbués et al., 1996); however, there are differences in interpretation among these workers, mainly caused by discrepancies in correlation schemes. The difficulties in correlating sections are mostly related to rapid and frequent lateral facies changes and to synsedimentary erosion and faulting. Detailed field mapping, correlation of sequence boundaries and systems tracts, and a new study of planktonic foraminiferal evolution have resulted in a considerable revision of the chronostratigraphy of the Campanian–Maastrichtian strata.

Hydrocarbon exploration in the area during the 1960s and 1970s met with very limited success. No oil has been found in the Upper Cretaceous rocks, only gas shows. Given the generally poor quality of the available seismic data, this lack of success can be attributed largely to problems of stratigraphic definition. There are also problems of reservoir quality resulting from widespread calcite cementation. Fresh analysis of the sequence stratigraphy

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[2]GeoPlay, Madrazo 32, Suite 5-3, 08006 Barcelona, Spain; e-mail: geoplay@catalunya.com

[3]Gessal, Toronga 21, 28043 Madrid, Spain; e-mail: gessal@idecnet.com

[4]Petroleum Consultant, Postbus 92, 7700 AB Dedemsvaart, The Netherlands: e-mail: pelie@xs4all.nl

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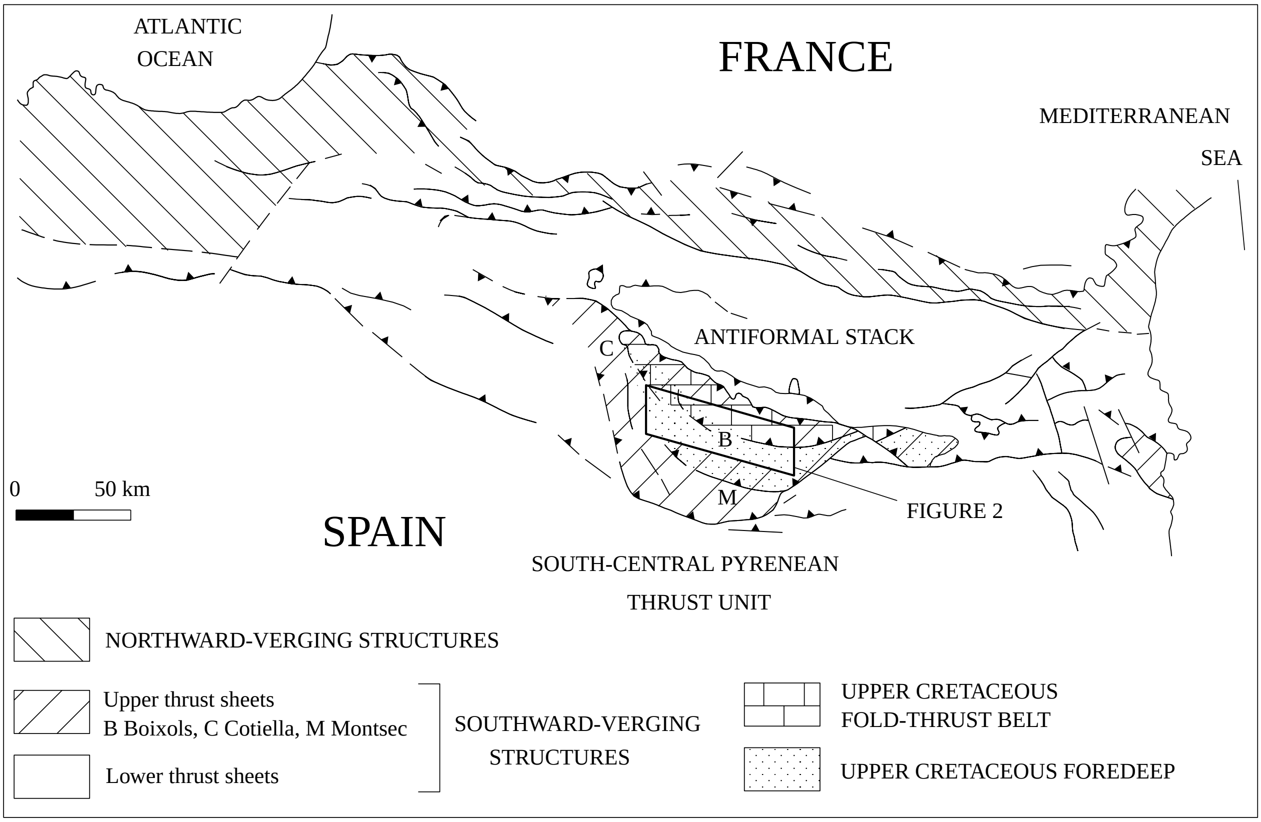

Figure 1—Generalized geologic map of the Pyrenees showing major structural units, modified from Berástegui et al. (1993). The box shows the study area (Figure 2).

may improve the interpretation of existing seismic data and be useful in planning the acquisition of new seismic data.

We also discuss the practical implications of the depositional model for the development of exploration play concepts elsewhere. Potential deltaic, submarine canyon, and slope reservoir systems are characterized by the integration of facies analysis, size and geometry of sand bodies, stacking patterns, and lateral/vertical distribution within a depositional-sequence framework.

GEOLOGIC SETTING

The South-Central Pyrenees

The Pyrenees are an Alpine fold-thrust belt extending east to west along northern Spain and southern France (Figure 1). The orogen formed during the Late Cretaceous–early Miocene as the result of south-to-north continental collision of the Iberian plate with the European plate. A cross section through the central Pyrenees has a fanlike geometry with an axial antiformal stack of Hercynian basement rocks flanked by both northward- and southward-directed thrust units (Muñoz, 1992) (Figure 1). We discuss strata within the south-central Pyrenean thrust unit (Séguret, 1972), a piggy-back imbricate sequence of three east-west–trending thrust sheets that detached above Triassic evaporites and were thrust over autochthonous Tertiary rocks (Cámara and Klimowitz, 1985; Vergés and Muñoz, 1990).

The Bóixols Thrust Sheet

The oldest thrust sheet, the Bóixols, is dominated by east-west–trending folds and consists of a 5000-m-thick carbonate section with strata that range in age from Triassic to Santonian (Berástegui et al., 1990). Field mapping and seismic analysis indicate that the Bóixols thrust sheet is formed by several imbricate thrust splays that are largely buried and become younger toward the foreland; these thrusts splays are the Bóixols sensu stricto, Riu, and Turbón (Figures 2–5).

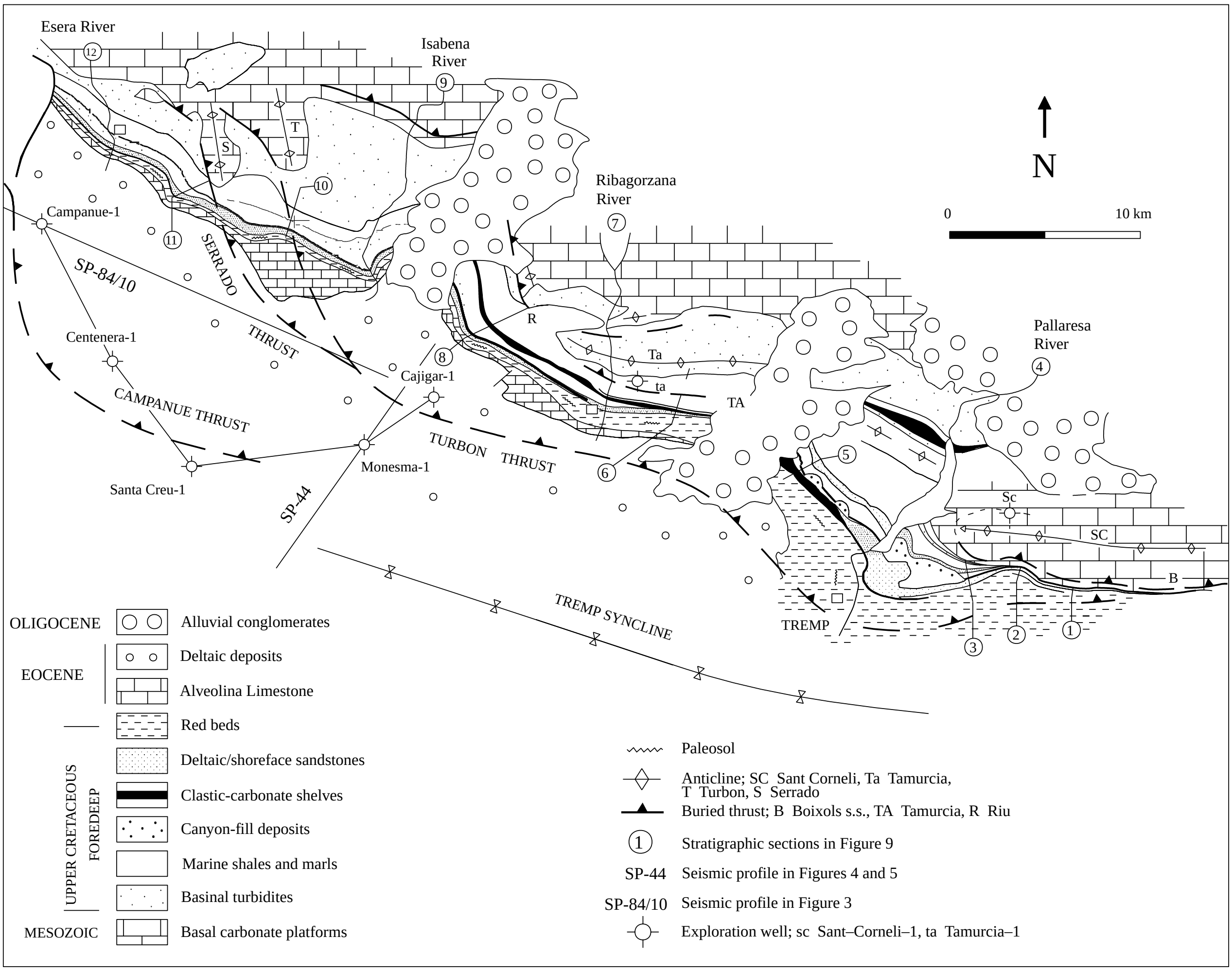

Figure 2—Simplified geologic map of the Upper Cretaceous foredeep of the south-central Pyrenees along the northern limb of the Tremp syncline. s.s.=sensu stricto.

The Bóixols thrust sensu stricto consists of a fold-thrust uplift (terminology of Berg, 1962) that can be traced for 32 km from east to west before it dies out into the Sant Corneli anticline (Garrido-Megías and Ríos, 1972) (Figure 2). This fold is interpreted as a fault-propagation growth anticline at the leading edge of the thrust (Deramond et al., 1993; Bond and McClay, 1995).

The Riu thrust crops out as a north-south–trending oblique ramp anticline, on the southern limb of which Cenomanian marls overlie Campanian turbidites (Figures 2, 5). To the east, the Riu thrust may be linked to the Tamurcia structure (Figure 2).

The oblique ramp of the Turbón thrust is defined on the surface by two north-south–trending thrust uplifts, the Turbón and Serrado anticlines (Figure 2). As seen on seismic records, both structures branch into a fold-thrust uplift in the Cajigar area that can be traced eastward until it merges with the Bóixols thrust sensu stricto (Figures 2, 3, 5). In the seismic profile of Figure 4, approximately 2.6 km of total displacement is inferred along the Turbón thrust.

The Campanué thrust is interpreted as the subsurface expression of the largely Eocene Cotiella thrust sheet, developed to the west of the study area (Figures 2, 3). To the southeast this structure is linked to the Montsec thrust sheet (Garrido-Megías and Ríos, 1972) (Figures 1, 5).

Thrust splays of the Bóixols thrust sheet are coeval with Campanian–Maastrichtian deposits (Simó and Puigdefàbregas, 1985; Eichenseer, 1988; Deramond et al., 1993). These deposits dip to the south along the northern limb of the Tremp syncline and are overlain by only mildly deformed Paleocene–lower

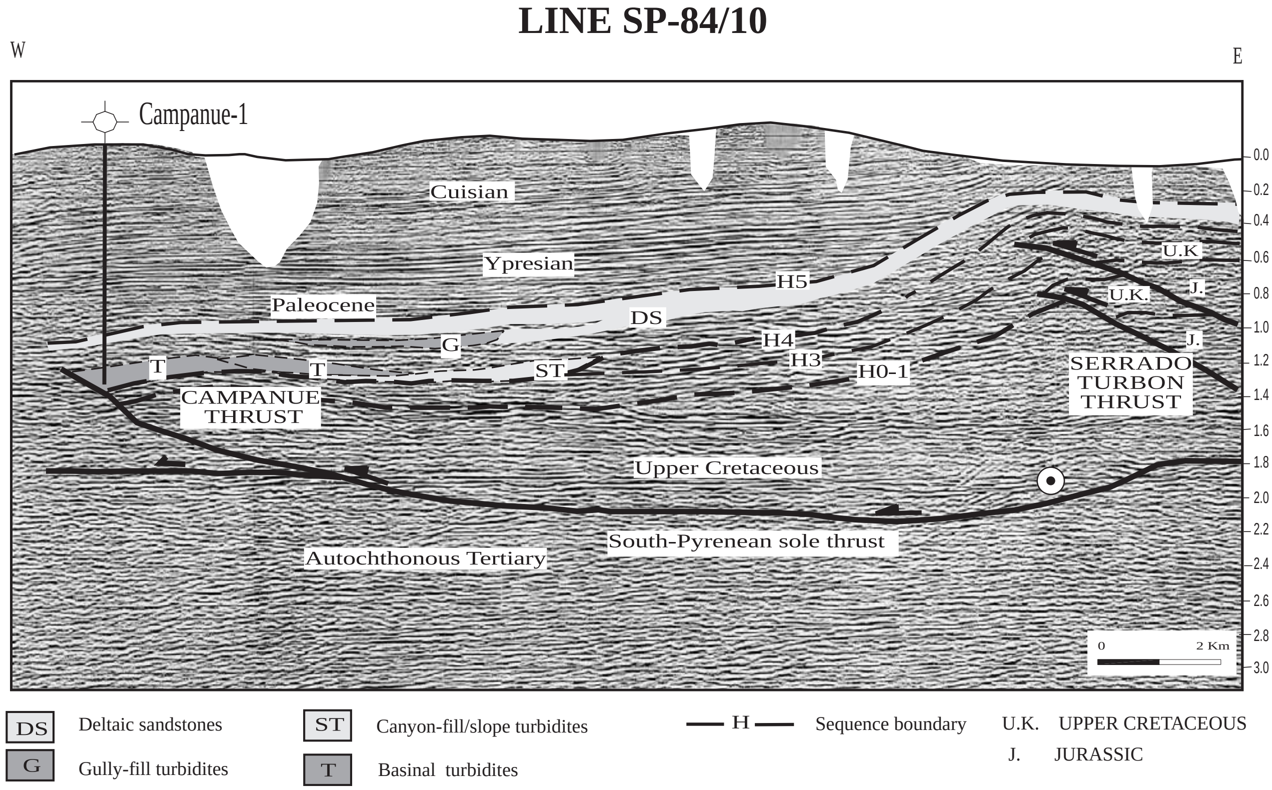

Figure 3—Interpreted migrated seismic reflection profile (SP-84/10 in Figure 2). Time is two-way traveltime. The profile has been calibrated against the Campanué 1 and Cajigar 1 wells, the latter being located 2.6 km east of the end of the line (Figures 2 and 7). The profile is aligned along paleodip from east to west. Numbers refer to the sequence boundaries discussed in the text.

Eocene shelf limestones. Late orogenic upper Eocene–Oligocene conglomerates were deposited unconformably on all of these rocks (Figure 2).

The Upper Cretaceous Foredeep

Southward movement on the Bóixols thrust sensu stricto was a result of Alpine inversion of a Lower Cretaceous extensional fault (Berástegui et al., 1990; Bond and McClay, 1995). The abrupt decrease in the thickness of the Jurassic and Lower Cretaceous south of the Riu and Turbón structures (Figure 5) implies that previous extensional faults were also tectonically inverted (J. G. Senz, 1998, personal communication).

South of the Bóixols thrust sheet, an elongate foredeep basin that deepened westward was created (Deramond et al., 1993) (Figure 1). The present-day dimensions of the foredeep fill are more than 100 km in length and 15 km in width. The basin was filled with westward-prograding deltaic-to-turbidite systems that range in age from Santonian to Maastrichtian. These were supplied by fluvial systems from the east, whereas alluvial fan supply came from the barely emergent orogen in the north (Arbués et al., 1996). Coeval shelf calcarenites and sandstones were deposited on the southern foredeep margin (Montsec uplift) (Figure 5). Simó (1993) interpreted surfaces of erosion and karstification as associated with tilting of a forebulge in this area.

STRATIGRAPHIC AND DEPOSITIONAL FRAMEWORK

The rocks in the south-central Pyrenean foredeep studied in this paper are exposed along the northern limb of the Tremp syncline and cover a distance of 75 km across the Pallaresa and Esera valleys (Figure 2). They form a 3400-m-thick section composed of three major, laterally transitional, depositional assemblages: basinal turbidites, deltaic deposits, and continental red beds (Figure 6). The overall section shallows upward, resulting ultimately in the complete filling of the basin. Characteristic lithofacies within

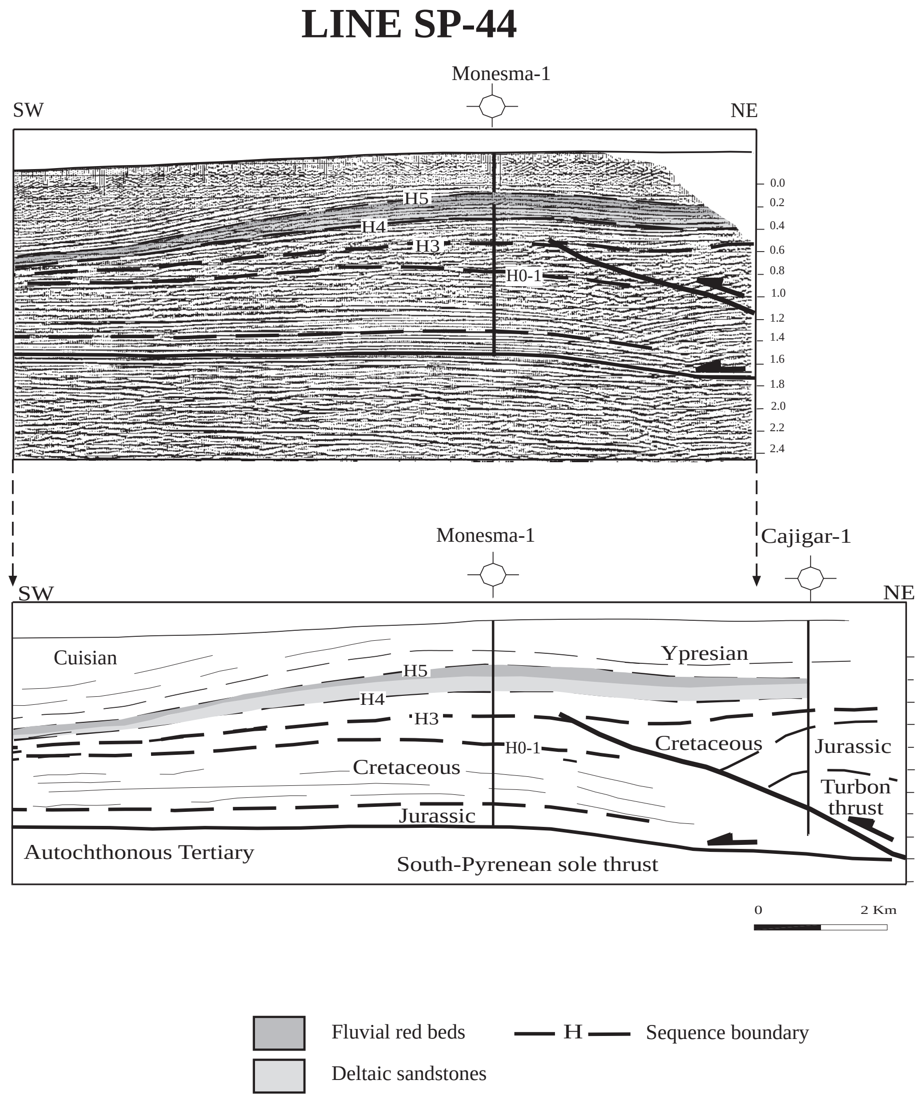

Figure 4—Interpreted migrated seismic reflection profile (SP-44 in Figure 2) and line-drawing section. Time is two-way traveltime. The profile is aligned along paleostrike. Numbers refer to the sequence boundaries discussed in the text.

each of these assemblages are briefly described in the following sections.

Basinal Turbidites

The basement of the foredeep consists mostly of Santonian carbonate platforms (Figures 5, 6). Above these strata the lower depositional assemblage forms a discontinuous fringe of turbiditic deposits up to 1500 m thick (Mascarell Member of the Vallcarga Formation) (Mey et al., 1968; van Hoorn, 1970; Nagtegaal, 1972). Turbidite sandstones (van Hoorn, 1970) are thin bedded and fine grained with minor thicker and coarser beds, and commonly display well-developed Bouma sequences. Channelized, thick-bedded, coarse-grained turbidites with abundant mudstone clasts are found in the upper part of the unit. Cohesive debris flows, slumps, and limestone breccias are locally common.

Deltaic Deposits

The middle depositional assemblage comprises a diversity of lithofacies, from slope carbonates and prodelta shales (Puimanyons olistostrome and Salàs marls members of the Vallcarga Formation, respectively) to shoreface and deltaic sandstones (Aren Sandstone Formation) (Mey et al., 1968; Nagtegaal, 1972). These deposits form an overall shallowing-upward succession up to 1500 m thick.

The Aren Sandstone is composed mostly of quartz grains with scarce potassium feldspar, chert,

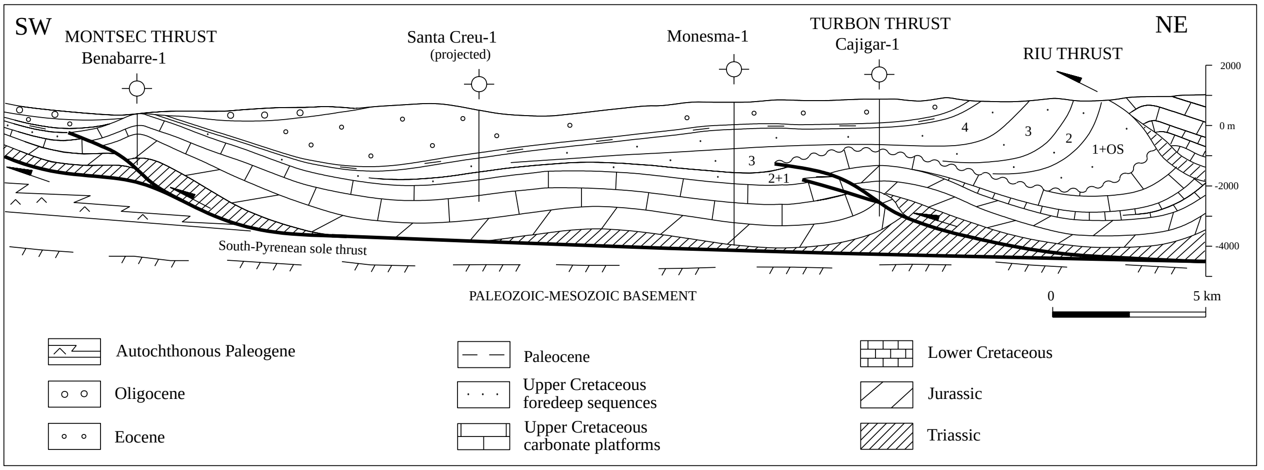

Figure 5—Structural cross section through the South-Central Pyrenean thrust unit along the seismic profile SP-44 (Figure 4). The section is transverse to the axis of the Upper Cretaceous foredeep. OS denotes older sequences and numbers refer to the depositional sequences Aren 1 to Aren 4. The Benabarre 1 well was drilled farther south on the Montsec thrust (Figure 1).

quartzite, mica, and fossil debris (Nagtegaal et al., 1983). In outcrop, most sandstones are pervasively cemented by calcite and have porosities of 5% or less. Fracture porosity is preserved in the subsurface, however, because the Aren Sandstone constitutes the main aquifer along the Tremp syncline (C. Roca, 1996, personal communication).

Continental Red Beds

The upper depositional assemblage is the lower part of the Tremp Formation (Mey et al., 1968). This formation is also known as the Garumnian facies. The assemblage, which is up to 400 m thick and thins basinward, is composed of coastal-plain variegated mudstones with intercalations of fluvial channels and lacustrine limestones (Krauss, 1990). The faunal content is restricted mainly to dinosaur remains. At the base, gray claystones with brackish water fauna and coal lenses are locally found. The sandstones and scarce conglomerates are mostly made up of Cretaceous limestone fragments cemented with calcite.

SUBSURFACE DATA

A number of hydrocarbon exploration wells were drilled in the south-central Pyrenees during the 1960s and 1970s in the hope that large antiformal structures, interpreted from poor-quality seismic reflection data, would provide structural traps. Primary targets were the Santonian carbonate platforms because of their good fracture porosity. Secondary targets were the fractured and karstified Dogger-Malm dolomites because they were potential analogs to the Aquitaine-type reservoirs in southern France (A. Garrido-Megías, 1998, personal communication). Upper Cretaceous and Eocene clastic formations were also of interest.

Five wells have penetrated into the Upper Cretaceous foredeep, but they are restricted to the western sector of the Tremp syncline (Figures 2, 7; Table 1). The Centenera 1 well is of particular interest because it discovered a noncommercial gas accumulation. During one drill-stem test, a 6-m-thick interval produced at an average rate of 58,000 m^3/day from fractured sandstones with a porosity of up to 10%.

A new seismic grid with 5–10 km spacing was acquired in the mid-1980s, but no more wells were drilled in the Tremp syncline. Figures 3 and 4 illustrate seismic sections from the western sector of the syncline, aligned along paleodip and paleostrike, respectively.

DEPOSITIONAL FACIES

Seven sandstone depositional facies are distinguished within the Upper Cretaceous foredeep strata. From proximal to distal, these facies are described in the following paragraphs.

Fluvial Channel

This facies association comprises a succession of erosively based fining-upward sandstone lenses interbedded with red shales (Figure 8A). Individual

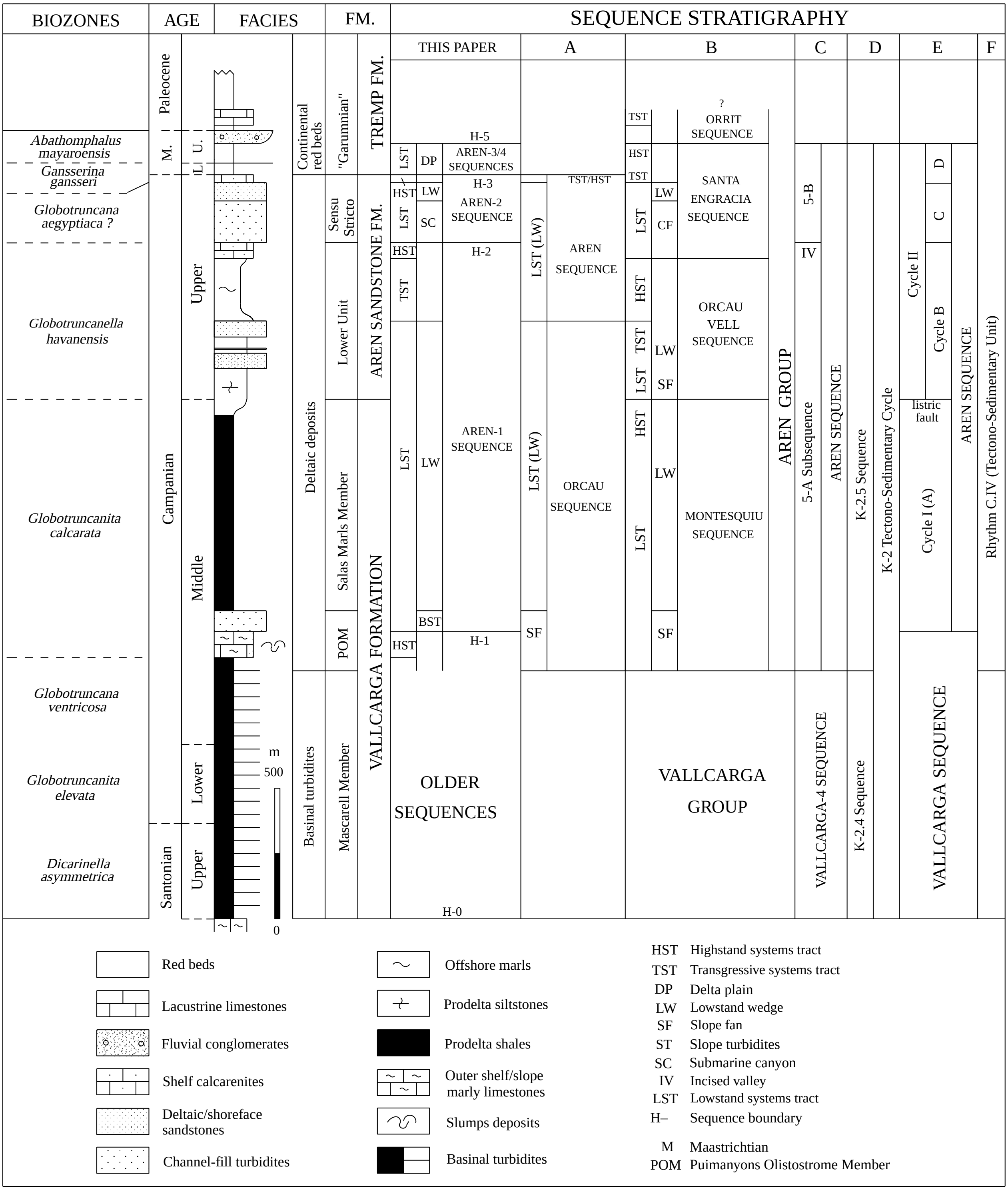

Figure 6—Biostratigraphy, lithostratigraphy (Mey et al., 1968; Nagtegaal, 1972), and sequence stratigraphy of the Aren Sandstone. The present analysis contrasted with previous work. Under sequence stratigraphy: (A) Sgavetti (1994, 1992) and Mutti and Sgavetti (1987); (B) Deramond et al. (1993) and Fondecave-Wallez et al. (1990); (C) Simó (1989); (D) Puigdefàbregas and Souquet (1986); (E) Simó (1986) and Simó and Puigdefàbregas (1985); (F) Garrido-Megías and Ríos (1972). The planktonic biochronozones have been established from the data of Caus et al. (1981), Elser (1982), Caus and Gómez-Garrido (1989), López-Olmedo and Ardèvol (in press), Senz (in press), and X. Orue-Etxeberria and J. I. Canudo (1998, personal communications). They are based on the biozonation of Robaszynski and Caron (1995) and Kennedy et al. (1995). *Gansserina gansseri* and *Abathomphalus mayaroensis* biozones are extrapolated from basinal settings.

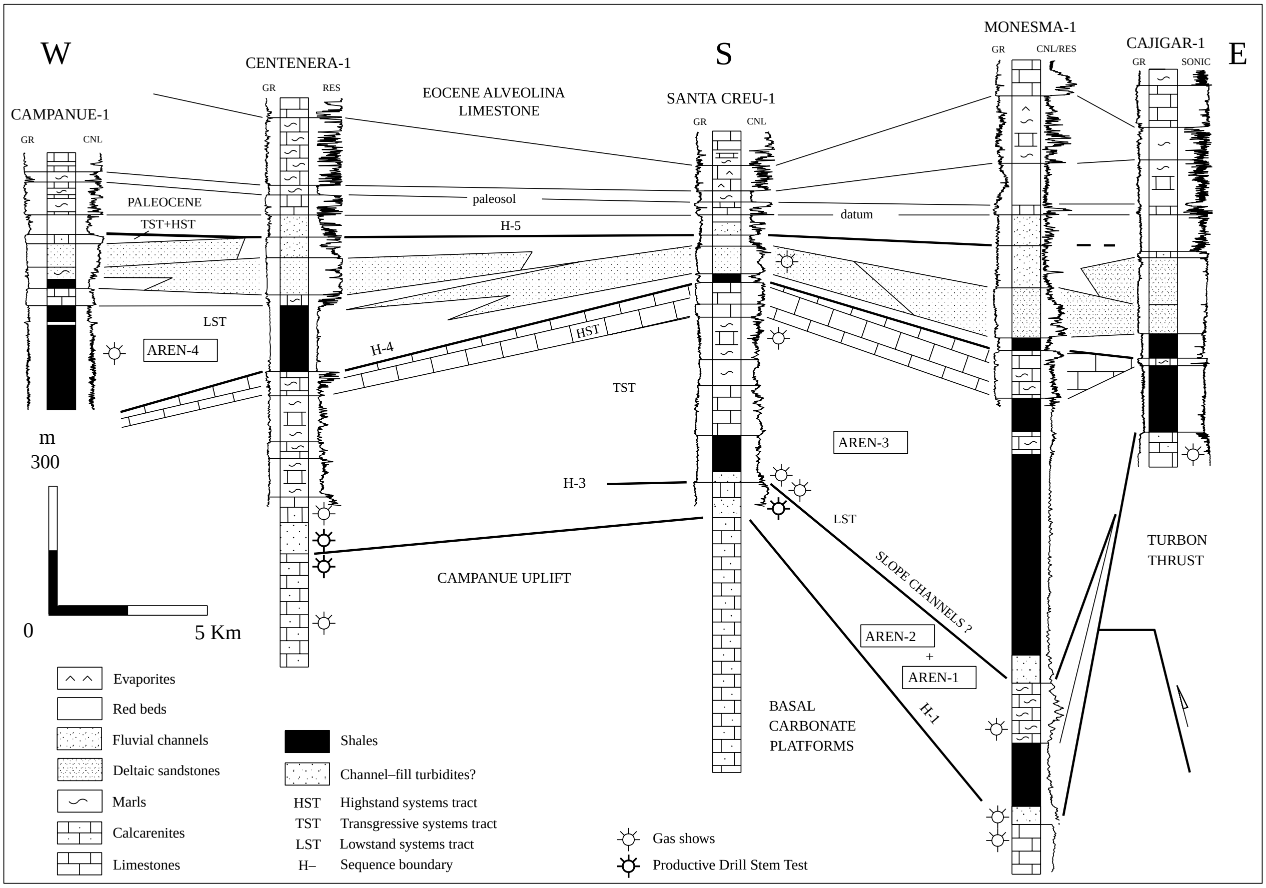

Figure 7—Correlation of well logs of representative penetrations into the Upper Cretaceous foredeep showing interpreted facies and depositional sequences. The wells are located in the Tremp syncline (Figure 2).

units range from 1 to 12 m in thickness and extend over several hundred meters to several kilometers. In outcrop, they commonly consist of two intervals. The lower interval includes matrix-supported cross-stratified conglomerates overlain by coarse- to fine-grained cross-bedded sandstones. The upper interval consists of fine sandstones to siltstones with a high degree of root bioturbation. Both intervals are laterally transitional, resembling a sigmoidal bar as described by Mutti et al. (1996). Individual bars are amalgamated or separated by thin mudstone intercalations.

Both the facies characteristics and lateral correlation with deltaic sandstones in a downdip direction suggest that the sand lenses were deposited as fluvial channels in a delta-plain environment; however, the channels and deltaic sandstones are not always physically in contact at outcrop. The development of sigmoidal bars suggests deposition by flood-related river systems in the sense of Mutti et al. (1996).

Delta-Front Mouth Bar

Quartz-rich, coarse sandstones in this depositional facies stack to form a series of prograding sand wedges separated by turbidites, bioturbated sandstones, or calcarenites. The sand wedges have erosive bottoms and planar tops, are up to tens of meters thick, and have several kilometers of downdip lateral continuity abruptly terminated by a surface of erosion. Internally, the sand wedges comprise a small number of amalgamated fining-upward units (Figure 8B), each beginning with massive, crudely graded, pebbly sandstones, overlain by coarse- to fine-grained sandstones with large- and medium-scale cross-bedding.

The fining-upward sandstone units are interpreted as deltaic mouth bars. Their vertical trend and poor sorting suggest the bars probably resulted from catastrophic flooding (Mutti et al., 1996) during which fine-grained sediments were washed away from the delta front. Such sand-rich deltas are

Table 1. Results of Exploration Wells and Penetrations into the Upper Cretaceous Foredeep*

Well	Total Well Depth (m)	Shows	Sequence	Lithology	Depth† (m)	Depositional Environment
Cajigar 1	3881.5	Gas	BCP	Limestone	1750	Shelf
Monesma 1	4750	Gas	Aren 1, 2	Marly limestone	2250–2265	Slope
Monesma 1	4750	Gas	Aren 1, 2	Sandy limestone	2425	Channel fill?
Monesma 1	4750	Gas	BCP	Limestone	2750	Shelf
St. Creu 1	3032	Gas	Aren 4	Sandstone	1596–1655	Delta front
St. Creu 1	3032	Asphalt				
St. Creu 1	3032	Gas	Aren 3	Sandstone	2112–2139	Channel fill?
St. Creu 1	3032	Gas	Aren 1, 2	Sandy limestone	2139–2173	Slope?
St. Creu 1	3032	Gas**	Aren 1, 2	Sandstone	2173–2220	Channel fill?
Centenera 1	4355	Gas**	Aren 2, 3	Sandstone	2400–2464	Channel fill
Centenera 1	4355	Gas**	BCP	Sandy limestone	2464–2700	Shelf
Campanué 1	3296	Gas	Aren 4	Sandy limestone	3115–3296	Prodelta

*BCP = basal carbonate platform; Aren 1 to Aren 4 refer to the depositional sequences discussed in the text.
**Productive drill-stem test.
†Depth of the productive interval.

interpreted elsewhere in the subsurface mostly as estuarine channel-fills and mouth sands.

Mixed Shelf

This facies association consists of medium-bedded bioclastic grainstones with up to 20% terrigenous sediment. They reach a maximum thickness of 75 m and show low-angle, large-scale seaward-dipping cross-stratification (Figure 8C). Petrographic studies (M. E. Arribas, 1998, personal communication) identified fragments of red algae, echinoids, rudists, bryozoans, and foraminifers such as *Orbitoides*. The terrigenous grains consist of quartz, feldspar, volcanics, chert, and mica coated by recrystallized sparry cement. The sedimentary structures are dominated by anisotropic hummocky cross-stratification in the sense of Arnott and Southard (1990).

Depositionally updip, the bioclastic grainstones onlap and gradually thin out into red beds. Downdip, they grade to a rhythmic alternation of marls and nodular bioclastic packstones and wackestones that contain foraminifers, spicules, nannoplankton, and ostracods. The bedding is also progressively contorted by slumping.

Similar facies associations of limited extent also occur. They show aggradational or progradational stacking patterns, downlapping and shaling out rapidly into offshore marls.

The virtually basin-wide calcarenites are interpreted as wave-dominated carbonate shelves with clastic influence. They grade downdip to outer shelf and slope environments. The hummocky cross-stratification is thought to have been produced by combined flows generated in coastal waters by enhanced wind and wave energy (Arnott and Southard, 1990). The laterally restricted calcarenite beds are interpreted as either beach deposits or, when landward-equivalent lagoonal facies are recognized, as barrier islands.

Channelized Turbidites

Channelized turbidite sandstones are commonly found in three distinctive depositional associations: gully-fill sands, canyon-fill sands, and slope channel-fill sands.

Gully-Fill Sands

Downcurrent, the delta-front sand wedges are replaced by turbidite sandstones via a surface of erosion. The beds are mostly amalgamated and fine grained with abundant mudstone clasts. They fill sharp-based sand bodies that measure several meters to tens of meters thick. Some of them stack to form a sediment pile up to 100 m thick and 3 km long (Figure 8D).

The erosional surfaces are interpreted as gully incisions produced by slope failure or growth faulting. The sand bodies infilling the gullies are interpreted as turbidite elements showing mostly channel features. These gully systems are comparable in terms of size, internal facies, and stratigraphic position to the oil-producing Miocene slope deposits of the San Joaquin basin, California, as described by Hewlett and Jordan (1994).

Canyon-Fill Sands

This facies fills large, deep incisions that are cut into mixed shelves and older deposits and associated with faults in the substratum. The infill is dominantly shaly but includes a number of channelized sand lenses. These lenses are several meters or tens of meters thick and several hundreds of meters wide, and show thinning- and fining-upward trends (Figure 8E). The sandstones are very thick to medium bedded, highly irregular, with erosive bases. The dominant lithology consists of very coarse to medium sands with abundant mudstone clasts near the bases. Cross-bedding may occur. Rare shaly intervals are associated with overbank thin-bedded turbidites. Wedges of calcarenitic blocks occur in some channels, decreasing in size and increasing in roundness away from the erosive margins.

The sand lenses are interpreted as turbidite channels. The large-scale erosional features are interpreted as submarine canyons, as deduced from their canyonlike geometry and from the occurrence of turbidite channels and resedimented blocks eroded from the canyon walls. The location of the canyons was controlled by synsedimentary faulting.

Analogous fault-controlled submarine canyons of the same age have been described from the oil-producing Campos basin in Brazil (Bruhn and Walker, 1995). Other examples may include the Lagoa Parda field, also in Brazil (Cosmo et al., 1991), and the Chicontepec field in Mexico (Busch, 1992).

Slope Channel-Fill Sands

This depositional facies consists of channels interbedded with mudstone (Figure 8F) and amalgamated channels (Figure 8G).

The channels interbedded with mudstone form accumulations up to 500 m thick. The channel sands measure a few meters in thickness and pinch out over several hundreds or thousands of meters. Most of the channels include thin-bedded fine-grained turbidites, although toward the base coarser and larger channels occur, as well as large-scale carbonate slumps. Turbidites are occasionally amalgamated and may show small mudstone clasts. Thin-bedded turbidites (levee facies) and slump deposits are observed in muddier intervals.

The amalgamated channels stack to form fining- or coarsening-upward sand bodies encased in shale up to 25 m thick, and can be traced longitudinally over several kilometers. They are filled with an amalgamation of structureless or convoluted, coarse- to medium-grained turbidites with abundant mudstone clasts at the base of minor channel fills.

These turbidite channels are related updip to submarine canyons, as inferred from the stratigraphic correlations (Figure 9). The amalgamated channels are interpreted to have been deposited at the canyon mouth in slope settings. Comparable facies associations have been described from intraslope basins (e.g., Pratson and Ryan, 1994) and submarine fans of the North Sea (e.g., Den Hartog et al., 1993; Shanmugam et al., 1995).

The channels interbedded with mudstone are interpreted to have been deposited on margins or levees. Again, analogous facies associations have been reported from North Sea submarine fans (e.g., Den Hartog et al., 1993) and the Port Acres oil field in the United States (Jackson, 1991).

Unchannelized Turbidites

Thick accumulations of basinal sands underlie, and develop downcurrent from, the described turbidite channels. These systems are not discussed in this paper.

SEQUENCE STRATIGRAPHY

Previous Work

The pioneering work of Garrido-Megías and Ríos (1972) interpreted both the middle and upper depositional assemblages (deltaic deposits and continental red beds) as one single, unconformity bounded, tectonic-sedimentary unit (Figure 6). This larger unit was later identified as one depositional sequence with its upper boundary in the middle part of the red beds (e.g., Puigdefàbregas and Souquet, 1986; Simó, 1989). Deramond et al. (1993) assigned this unit to the Aren Group and divided it into four depositional sequences, similar to those presented in this paper; however, Figure 6 shows differences in correlation schemes and the placing of sequence boundaries. By contrast, Mutti and Sgavetti (1987) and Sgavetti (1992, 1994) divided the deltaic deposits into two sequences and placed a sequence boundary between deltaic deposits and continental red beds.

Methods

Field mapping and the correlation of multiple stratigraphic sections form the basis for the sequence-stratigraphic framework described in this paper (Figure 9). Five basin-wide, carbonate-rich levels, intercalated within thick siliciclastic strata, provide key correlation markers for defining four depositional sequences named Aren 1 to Aren 4 (Figure 10). The biostratigraphic framework, based on planktonic foraminifera, is linked to the biozonation of Robaszynski and Caron (1995) and the time scale of Gradstein et al. (1995) (Figure 6). The sequence-stratigraphic terms used to describe the Aren Sandstone succession are adopted from Van

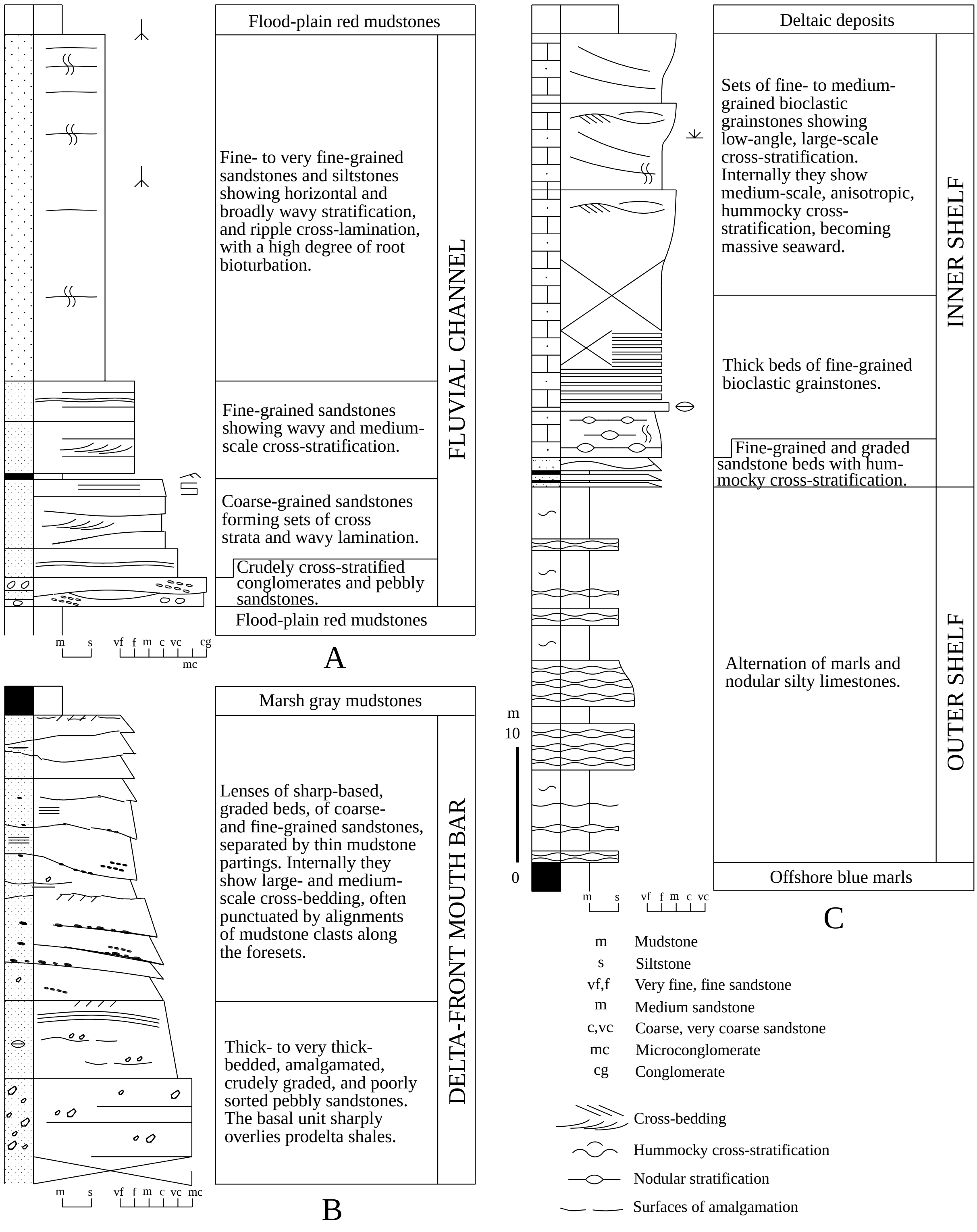

Figure 8—Representative stratigraphic logs of sandstone depositional facies distinguished within the Upper Cretaceous foredeep. Location of logs is indicated by corresponding letters in Figure 9. (A) Fluvial channel, section 2; (B) delta-front mouth bar, section 4; (C) mixed shelf, section 5; (D) gully-fill turbidites, section 8; (E) canyon-fill turbidite channel, section 4; (F) slope turbidite channel, section 9; and (G) slope amalgamated turbidite channels, section 10.

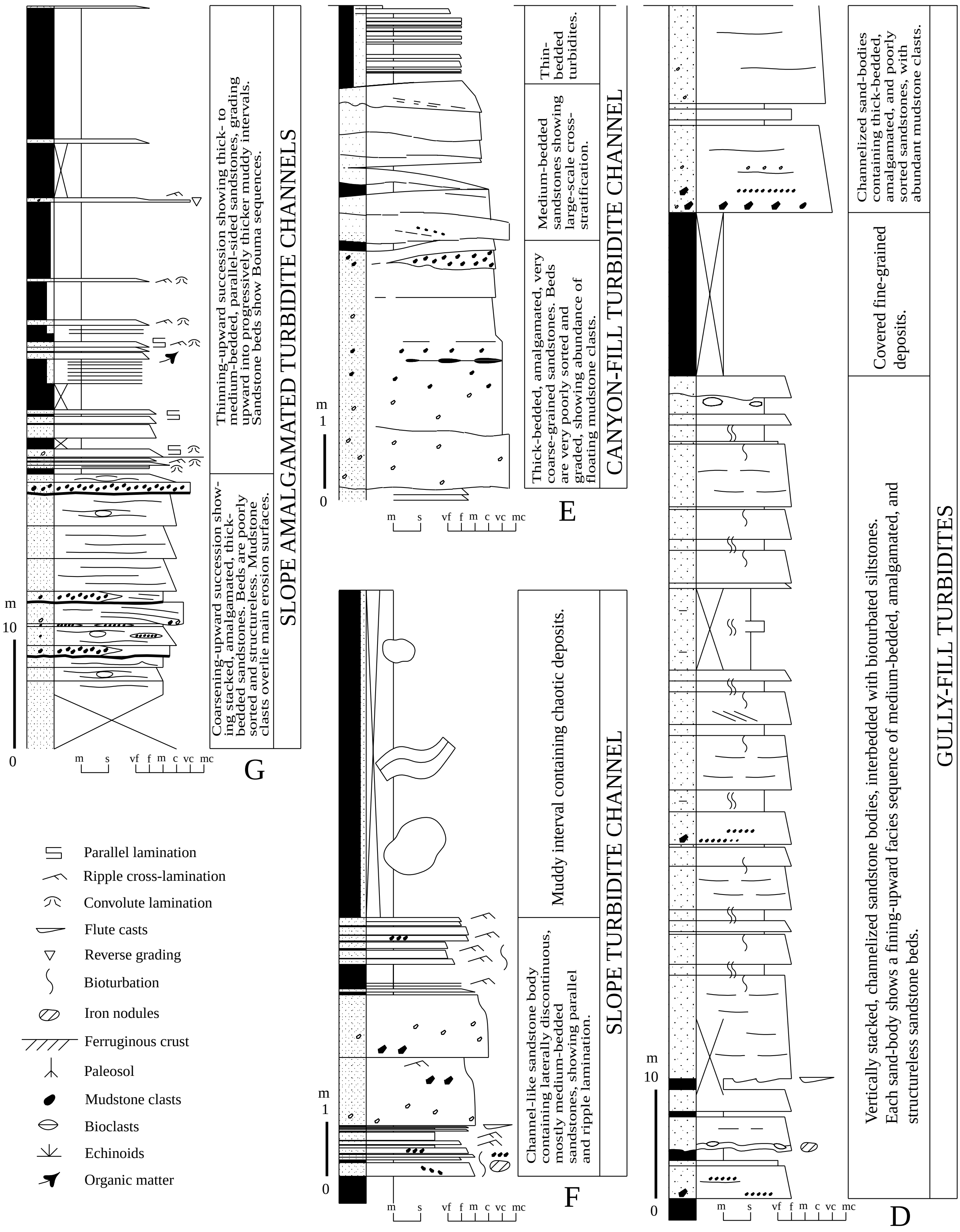

Figure 8—Continued.

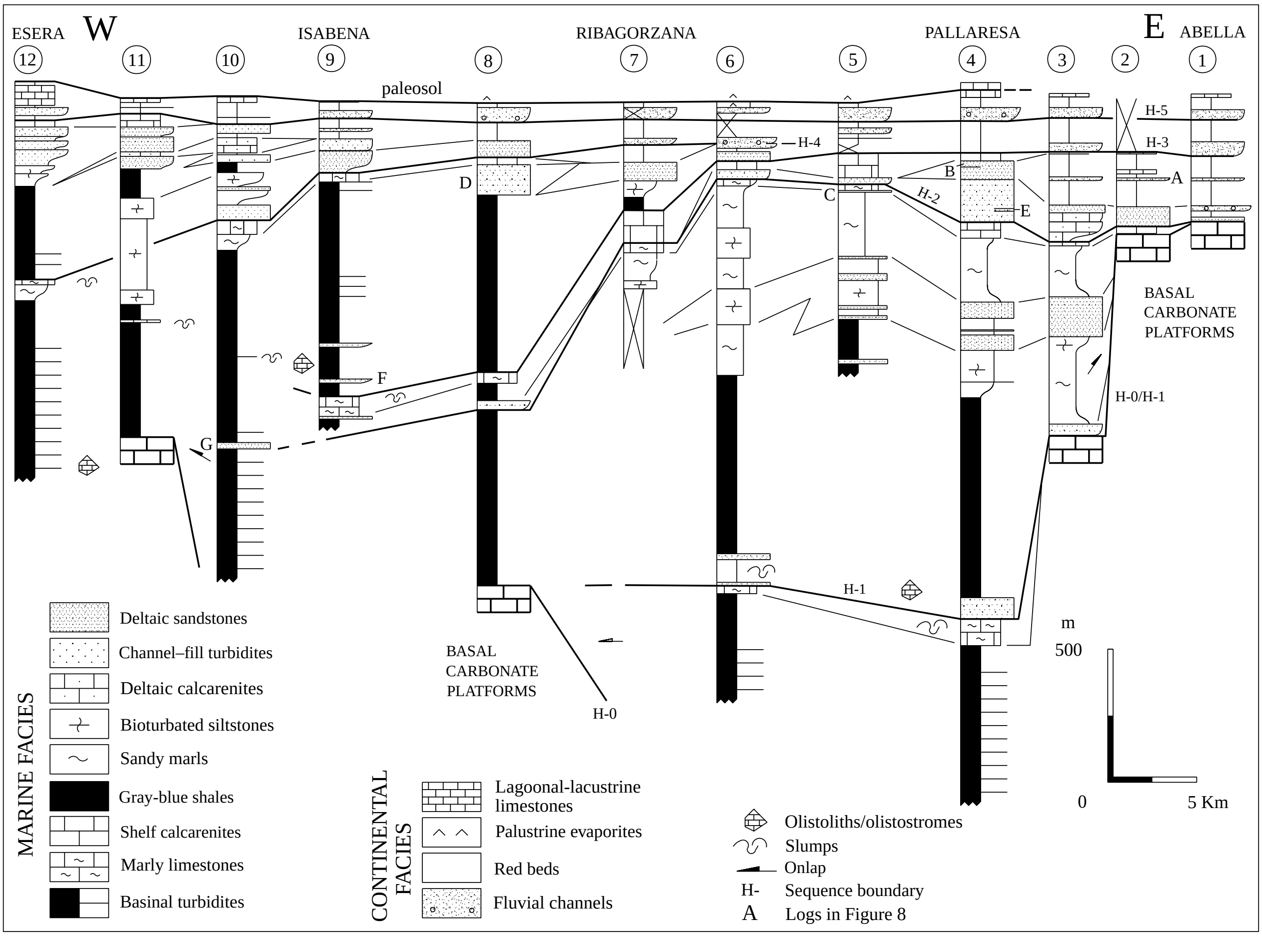

Figure 9—Schematic cross section showing the correlation of main representative stratigraphic sections of the Upper Cretaceous foredeep along the northern limb of the Tremp syncline. Depositional facies, facies transitions, and interpreted sequence boundaries are represented. Locations of sections are shown in Figure 2. The correlation diagram is supported by field mapping at scale 1:25,000.

Wagoner et al. (1988) and Posamentier and Vail (1988).

Aren 1 Sequence

The lowermost basinal turbidites infilling the Upper Cretaceous foredeep above the horizon H-0 (older sequences in Figure 10) are not discussed in this paper (for further information refer to van Hoorn, 1970; Simó, 1993; Deramond et al., 1993). On the footwall syncline of the Bóixols thrust sensu stricto these rocks are overlain by deltaic deposits of the Aren 1 sequence, some 1500 m thick in outcrop (Figure 10). The antiformal shape of the sequence is interpreted to be related to synsedimentary compressional deformation (Mutti and Sgavetti, 1987).

Sequence Boundary

The sequence boundary of the Aren 1 sequence is located at a surface of erosion on slope limestones (highstand systems tract or HST of the preceding sequence), blanketed by a pebble lag (horizon H-1 in Figures 9, 10). On well logs the horizon H-1 is identified with the first appearance of sandstone above limestone (Figure 7). Discontinuous seismic reflections are interpreted as an unconformity with truncation of previous strata associated with surfaces H-0 and H-1 in seismic profiles (Figures 3, 4).

Systems Tracts

At the base of the sequence, two channel-fill sand bodies form a slope turbidite system (Figure 10). These are overlain by a thick section of prodelta shales and tide-influenced sandstones and siltstones

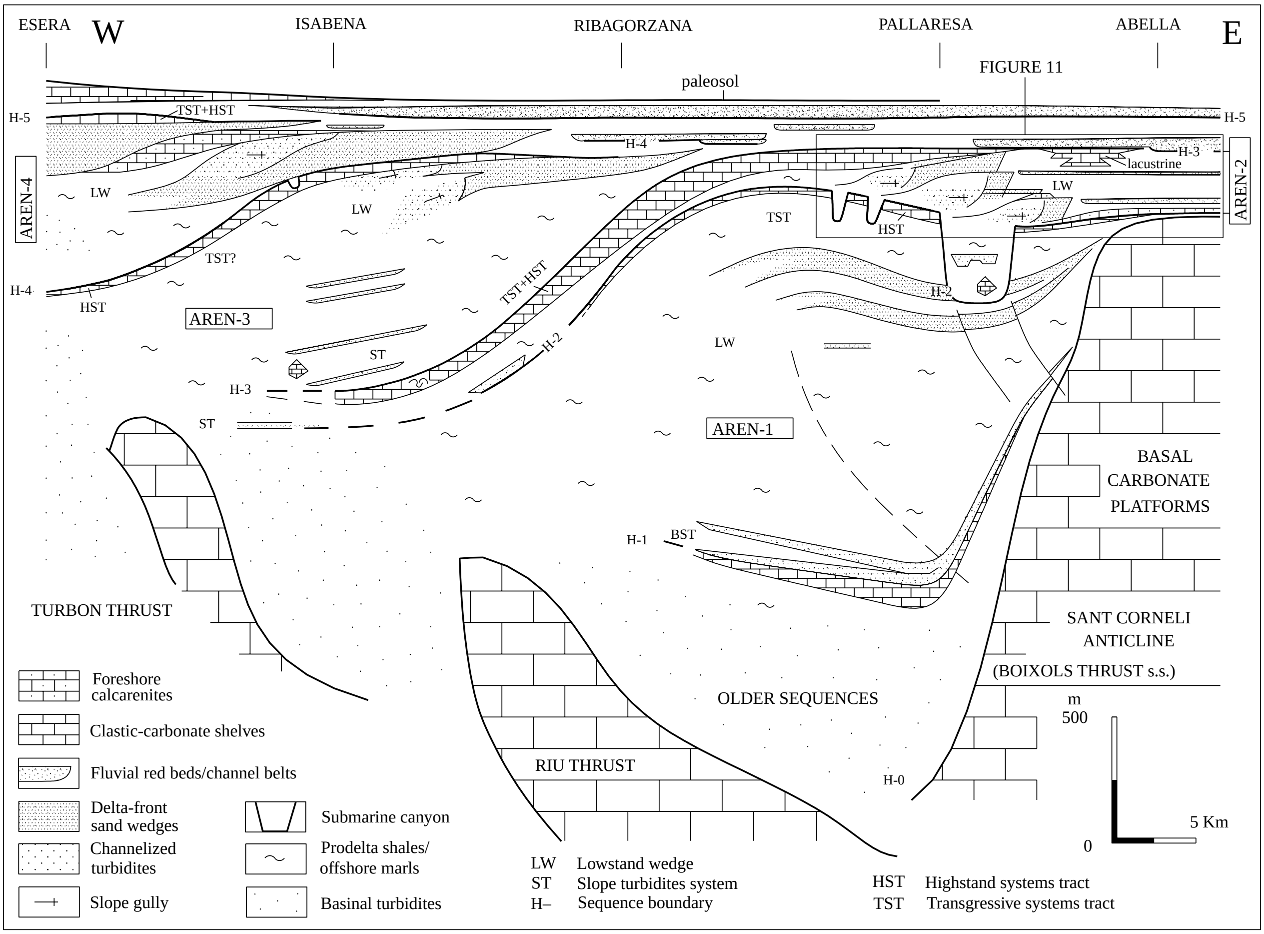

Figure 10—Stratigraphic cross section of the Upper Cretaceous foredeep along paleodip. Depositional systems and their relationships to synsedimentary anticlines are shown. Note that these anticlines are not the in the same vertical plane as the foredeep deposits, which lie south of the structures (Figure 2). s.s.=sensu stricto.

interpreted as a prograding lowstand wedge. The wedge, capped by a hardground surface, is overlain by transgressive marls (transgressive systems tract or TST) grading upward to a 50-m-thick package of shelf calcarenites (HST), with echinoids and coral/rudist patch-reefs (Gallemí et al., 1983; Liebau, 1984). The LST (lowstand systems tract) and TST onlap with progressive unconformity onto the hanging-wall anticline of the Bóixols thrust sensu stricto, which is unconformably overlapped by the HST (Figure 10).

Chronostratigraphy

The uppermost turbidite rocks underlying the Aren 1 sequence are middle Campanian in age, as indicated by the occurrence of planktonic foraminifera from the *Globotruncana ventricosa* zone (Caus and Gómez-Garrido, 1989). The slope limestones underlying the Aren 1 sequence and the prodelta shales contain foraminifera from the *Globotruncanita calcarata* zone (Caus et al., 1981; López-Olmedo and Ardèvol, in press). Basinward, the Aren 1 deltaic sandstones have yielded fauna from the *Globotruncanella havanensis* zone (X. Orue-Etxeberria, 1998, personal communication). In conclusion, the age of the Aren 1 sequence is mid- to late-Campanian and spans approximately the interval 76 to 74 Ma (Figure 6).

Aren 2 Sequence

The fluvio-deltaic deposits of the Aren 2 sequence filled up the remaining accommodation space on the footwall of the Bóixols thrust sensu stricto and reached a thickness of 600 m in outcrop (Figure 10).

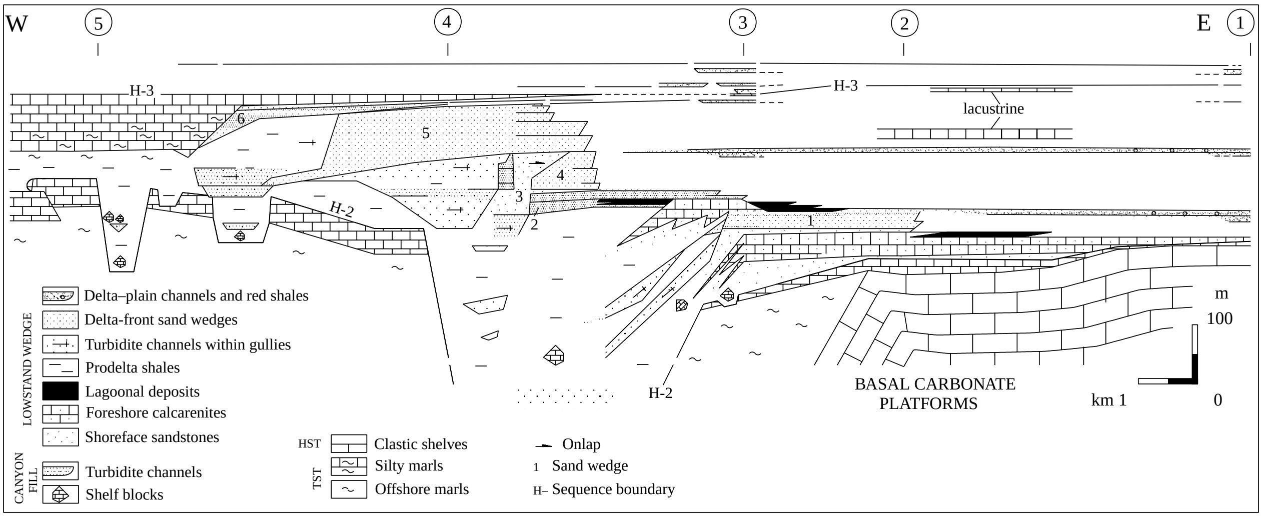

Figure 11—Stratigraphic cross section of the Aren-2 sequence in its fluvio-deltaic setting along paleodip. The basal shelf (preceding HST) is cut by three transverse canyon incisions (sequence boundary H-2). Overlying the canyon-fill deposits, the lowstand wedge consists of six prograding deltaic sand wedges interstratified with transgressive turbidites and shoreface deposits (see Figure 12). The location of the cross section in the stratigraphic framework of the Upper Cretaceous foredeep is shown in Figure 10.

Sequence Boundary

The sequence boundary of the Aren 2 sequence is a karst surface on inner shelf strata of the preceding Aren 1 highstand (horizon H-2 in Figures 9–11). The karst surface is laterally correlative with a series of north-south–trending deep incisions, interpreted as submarine canyons (J. Rosell, 1985, personal communication). The largest canyon, 300 m deep and 5 km wide, was controlled by two systems of syndepositional reverse faults with a net displacement of 30 m. Its eastern margin has been interpreted either as a listric fault (e.g., Simó, 1986) or as a surface of deep erosion crosscutting basinward (e.g., Sgavetti, 1994).

Systems Tracts

In a downdip direction, slope turbidite channels can be correlated with the canyons (Figure 10). Updip, a 250-m-thick deltaic complex (lowstand wedge) reveals six prograding delta-front sand wedges interstratified with transgressive deposits (Figure 11). These packages are interpreted to be high-frequency depositional sequences (Figure 12). The uppermost sand wedge is cut by a transgressive ravinement surface with a local pebble lag, overlain by backstepping shoreface sandstones (TST) and prograding shelf calcarenites up to 75 m thick (HST) (Figures 10, 11).

Chronostratigraphy

A late Campanian age may be assigned to the Aren 2 sequence because the upper part of the preceding sequence is early late Campanian, and the planktonic foraminifera *Gansserina gansseri* zone occurs basinward in the Aren 2 HST (Elser, 1982; X. Orue-Etxeberria, 1998, personal communication) (Figure 6). The sequence spans approximately the interval 74 to 72 Ma.

Aren 3 Sequence

The deltaic deposits of the Aren 3 sequence accumulated in a westward depocenter on the footwall of the Riu thrust and reached a maximum thickness of 750 m in outcrop (Figure 10).

Sequence Boundary

The sequence boundary of the Aren 3 sequence is a 1-m-thick paleosol on inner shelf strata of the preceding Aren 2 highstand, which shows boring, iron mineralization, and accumulations of terrestrial snails (horizon H-3 in Figures 9–11). The soil is sharply overlain by coastal-plain deposits. On the outer shelf to slope the boundary is an erosive unconformity accompanied by a shift of facies to prodelta shales.

Systems Tracts

In outcrop, canyon incisions were not recognized at the base of the Aren 3 sequence, but a 500-m-thick slope turbidite system accumulated downdip

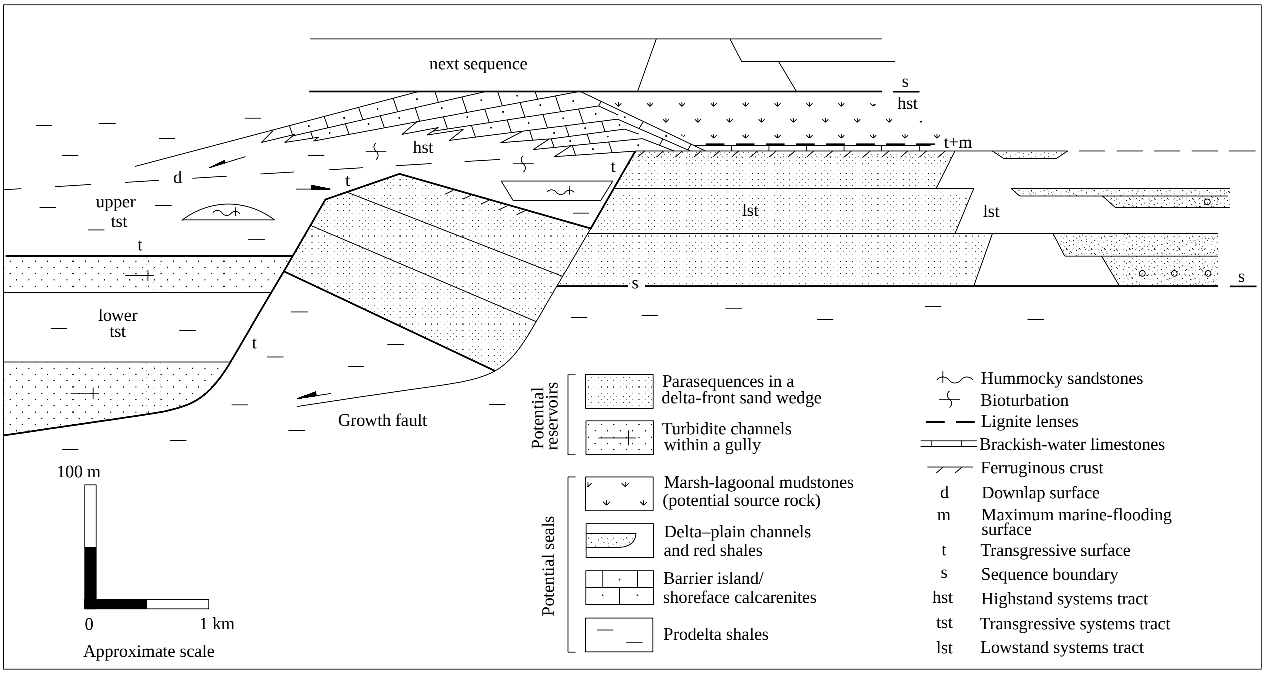

Figure 12—Fourth-order depositional sequence model. See text for explanation. The fourth-order lowstand systems tract is represented by the fluvio-deltaic sand wedge eroded downdip by a gully incision that resulted from growth faulting. Turbidite gully fills are interpreted to be deposited during early rise of relative sea level [lower part of the fourth-order transgressive systems tract (TST)]. Shoreface sandstones and siltstones onlapping the transgressive surface built up the upper part of the fourth-order TST. Lagoonal marls and barrier island calcarenites were deposited during relative highstand (fourth-order highstand systems tract).

(Figure 10). The lowstand wedge is interpreted to be a shelf-edge delta expanding basinward, and is well exposed in the Ribagorzana valley (Figures 2, 10). These rocks have been interpreted as a regressive beach complex (Mutti et al., 1975) and forced-regressive shoreline (Posamentier et al., 1992). The HST is represented by shelf calcarenites in a prograding stacking pattern. These become outer shelf marly limestones in the Isábena valley, where they overlie basinal marls (TST) and underlie the Aren 4 lowstand sandstones (Figures 2, 9). The HST overlaps the hanging wall of the Turbón thrust (Figures 5, 7).

Chronostratigraphy

Planktonic foraminifera from the *Gansserina gansseri* and *Globotruncana contusa* zones appear in prodelta shales of the Aren 3 sequence (Elser, 1982; X. Orue-Etxeberria, 1998, personal communication). The HST marks the last appearance of fauna from the *G. gansseri* zone as defined by X. Orue-Etxeberria and J. I. Canudo (1998, personal communication), and the first appearance of fauna from the *Abathomphalus mayaroensis* zone as defined by Senz (in press). These determinations give an age of latest Campanian–early Maastrichtian for the Aren 3 sequence, approximately spanning the interval 72 to 69 Ma (Figure 6).

Aren 4 Sequence

The deltaic deposits of the Aren 4 sequence accumulated on the footwall of the Turbón thrust and reached a thickness of 500 m in outcrop (Figures 3, 10).

Sequence Boundary

The sequence boundary of the Aren 4 sequence is represented by sharp-based deltaic sands and turbidite channels (submarine canyon?) eroding into the preceding Aren 3 highstand (horizon H-4 in Figures 7, 9, 10). On the seismic profile of Figure 3, the H-4 sequence boundary is interpreted as a deep canyon incision truncating the underlying sequence.

Systems Tracts

On the seismic profile of Figure 3, the truncation surface is overlain by a series of downlapping sigmoidal reflections interpreted as canyon-fill to slope turbidites (ST). The mounded patterns are interpreted as basinal turbidites (T) because strata of this type occur in outcrop farther north (Figure 10). The progradational shingled reflections are interpreted as deltaic clinoforms (DS). In outcrop and well logs, two major delta-front sand wedges (lowstand wedge) are recognized in the Isábena and Esera sections (Figures 2, 7, 10). The lowstand wedge is overlain by a 50-m-thick wave-built calcarenite shelf, which includes the TST and HST. Landward, the shelf is eroded by a surface of karstification (horizon H-5 in Figures 7, 9, 10). The erosion surface can be related updip with the base of alluvial fan deposits (Talarn conglomerates).

Chronostratigraphy

The prodelta shales of the Aren 4 sequence contain planktonic foraminifera from the *Abathomphalus mayaroensis* zone (X. Orue-Etxeberria, 1998, personal communication). Above the horizon H-5, in the Isábena valley, fluvial channels contain dinosaur bones (N. López-Martínez, 1998, personal communication); consequently, the Aren 4 sequence is dated to late Maastrichtian and spans approximately the interval 69 to 66 Ma (Figure 6).

The deposits with dinosaur remains underlie lagoonal-lacustrine limestones of Danian age, topped by a basin-wide paleosol horizon (Eichenseer, 1988; van den Hurk, 1990) (Figures 7, 9, 10). This complex builds up a broadly tabular sequence (uppermost Maastrichtian–Danian) and displays continuous, parallel seismic reflections that indicate the end of the foredeep stage (Figures 3, 4).

DISCUSSION

Sequence Architecture

Each siliciclastic-carbonate package in the Aren Sandstone succession is interpreted to be a type 1 third-order depositional sequence (Figure 10), according to the criteria of Van Wagoner et al. (1988). Each sequence spans 2–3 m.y.

During relative sea level fall the inner shelf underwent karstification while canyons were incised on the outer shelf, feeding coeval slope and basinal turbidite systems. During lowstand, sand-rich deltas accumulated within the lowstand wedges. Some of the Canadian Holocene deltas described by Hart and Long (1996) provide modern analogs. Both deltaic systems are characterized by sharp-based littoral sands overlying prodelta sediments, slope failure with channel development, and deposition of base-of-slope turbidites.

TSTs are represented by offshore marls and backstepping lower shoreface sandstones wedging onto the shelf margin. The transgressive surface is a hardground or ravinement. During relative highstand, mixed clastic/carbonate shelves were deposited and grade downdip to chaotic slope marly limestones. Updip, brackish water and lacustrine limestones were deposited.

Fourth-Order Depositional Sequences

Third-order lowstand wedges consist of a succession of fluvial/deltaic sand wedges interstratified with finer grained sediments that record episodes of shoreline progradation followed by a transgression (see, for example, the Aren 2 sequence in Figure 11). They thus built up a set of higher order depositional sequences (Figure 12).

The fourth-order sequence boundary is a surface of fluvial/deltaic erosion or deposition on preceding lagoonal or marsh deposits. The fourth-order LST is represented by the delta-front sand wedge. Sand wedges are usually composed of two to five parasequences (mouth bars, Figure 8B).

The fourth-order transgressive surface is the gully incision of the LST, locally blanketed by lags of pebbles and fossil fragments. Updip, it grades to a ferruginous surface on top of the deltaic sands, which commonly contains dinosaur bones or egg shells (e.g., Sanz et al., 1995). Gully-fill turbidites, interpreted as the lower part of the fourth-order TST, are overlain by onlapping shoreface sediments, treated as the upper part of the fourth-order TST.

Above the ferruginous surface, the fourth-order maximum marine-flooding surface is a decimeter-thick limestone layer associated with lignite and coquina beds. The fourth-order HST is formed by a lagoon-barrier island system. HSTs are frequently absent owing to nondeposition or erosion.

Depositional Model

Third-order depositional sequences of the Upper Cretaceous foredeep filled uplift-bounded minibasins. This development followed the migration of the depocenter from east to west, interpreted to be related to the emplacement of thrust splays (Figure 13).

During phases of active uplift, thrusting, and associated flexural subsidence, accommodation space was created into which turbidite systems were deposited. These accumulated on the footwall syncline and onlapped the backlimb of the next growing structure, which acted as a sediment trap. Basin fills grade vertically from unchannelized to channel-fill sands. The more outlying depocenter was either starved of sediment or received distal turbiditic deposits. Deltaic wedges prograded over deep-water sediments, and terrestrial systems were

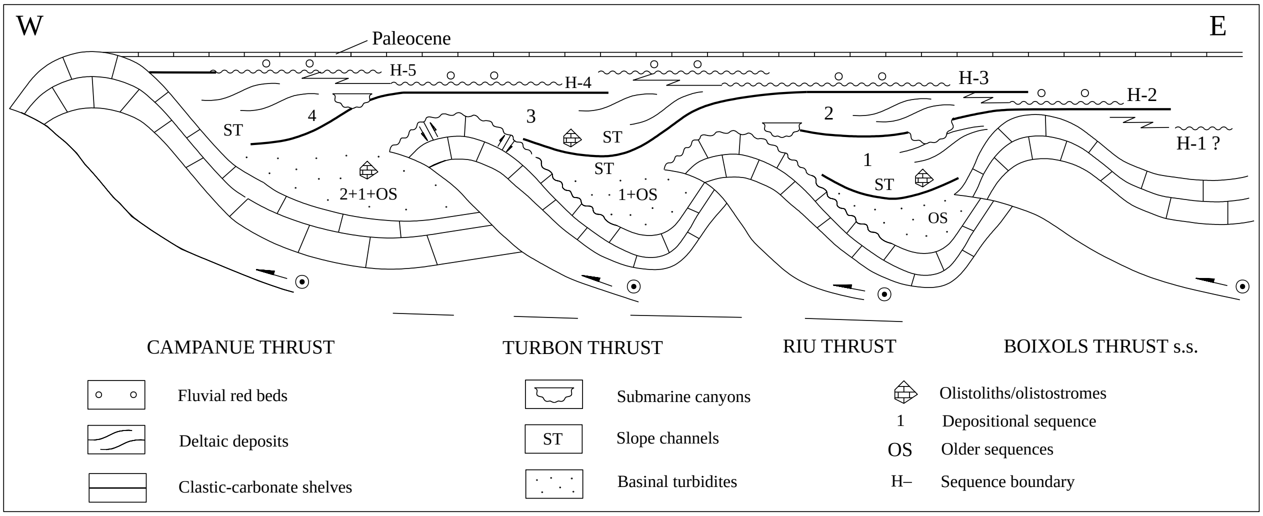

Figure 13—Drawing illustrating the depositional sequence response to thrust propagation. See text for explanation. Not to scale. Approximate horizontal distance 100 km; approximate maximum thickness of clastic deposits 3000 m. s.s.=sensu stricto.

deposited on thrust-top basins. The submarine canyons followed structural lows, with steep walls controlled by faults that were linked to the anticline crests. These faults also controlled the location of successive coastlines and shelf-breaks.

When uplift ceased or the rate of uplift/sedimentation reached equilibrium, the remaining accommodation space was filled with finer grained TSTs. As thrust activity waned, so subsidence decreased and mixed sediments prograded unconformably over structural highs, forming HSTs. When the footwall syncline was filled, the depocenter shifted to the forelimb of the next structure.

Comparable tectonic-depositional models have been proposed for the Italian Apennine foredeeps (i.e., Ricci Lucchi, 1986; Butler and Grasso, 1993); furthermore, the depositional model of the Upper Cretaceous foredeep may serve as an analog for the Tertiary slope minibasins of the Gulf of Mexico and the West African margin, where the growth of salt diapirs and sedimentation are closely linked (e.g., Weimer et al., 1998). Certain similarities in basin configuration and infill are also shared with the ponding-to-bypass basin model as defined by Prather et al. (1998).

Timing and Sequence of Uplifts

The transition from an extensional, thermally subsiding margin to foreland occurred in the late Santonian, as interpreted from a change from carbonate platform to widespread turbidite deposition. The growth of fold-thrust uplifts was coeval with the development of the older turbiditic sequences in the late Santonian–early Campanian. This is inferred from the onlapping of strata onto the backlimbs of the anticlines and from the overthrusting in the forelimbs (Figures 10, 13). The cessation of deformation progressed from northeast to southwest, as inferred from the depositional relationship between biostratigraphically dated sequences and growing anticlines (Figures 13, 14).

The slope limestones preceding the Aren 1 sequence constrain movement on the Bóixols thrust sensu stricto to the middle Campanian. Further uplift of its hanging-wall anticline during the middle–late Campanian is sustained by the progressive unconformity of the LST and TST of the Aren 1 sequence. The upper Campanian HST overlaps the anticline (Figures 13, 14). The Aren 2 sequence provides data on the timing of the Riu thrust. The upper Campanian LST is coeval with thrust uplift, whereas the HST overlaps the structure, dating the cessation of movement to the latest Campanian (Figures 5, 10). Cessation of uplift on the Turbón thrust is constrained by overlying lower Maastrichtian strata of the Aren 3 sequence. This is recorded in the thinning out and slight overlap of the Aren 3 sequence onto the hanging wall (Figures 3, 5, 7). The upper Maastrichtian Aren 4 sequence is only slightly affected by early uplift in the area of the Campanué thrust.

Foredeep deformation culminated in the Cretaceous, as documented by the continuity of beds in the uppermost Maastrichtian–Danian sequence. These deposits reflect a period of low relief and tectonic stability (Eichenseer, 1988). As suggested by seismic sections and as seen in outcrop farther west, these rocks were affected by the Campanué

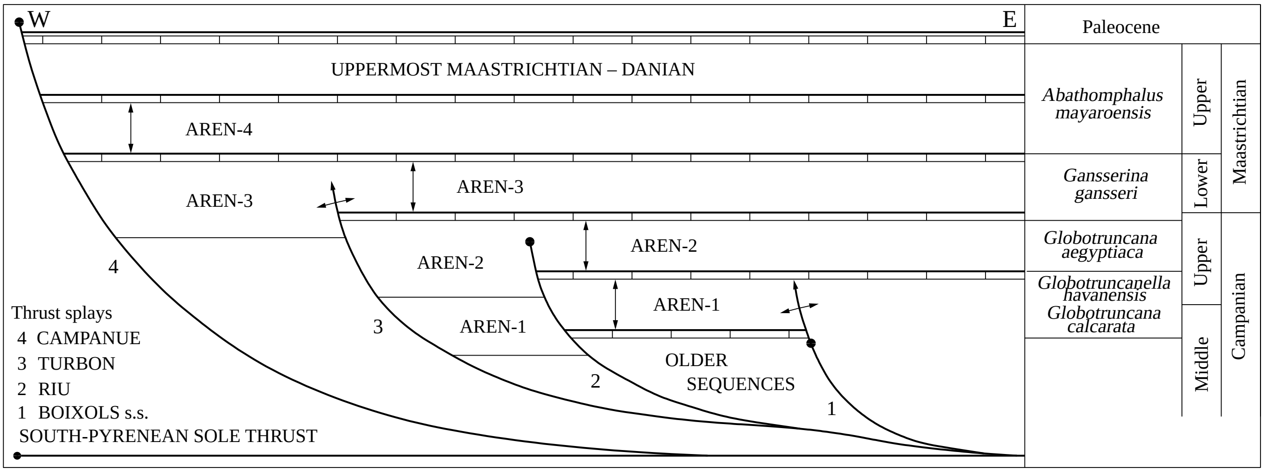

Figure 14—Timing and sequence of uplifts. The cessation of the deformation along the Bóixols thrust sheet of the Upper Cretaceous foredeep progressed from northeast to southwest. This is inferred from the depositional relationship between biostratigraphically dated sequences (Aren 1 to Aren 4) and growing anticlines (1 to 4). s.s.=sensu stricto.

thrust, which was emplaced in the late Paleocene–Eocene during later stages of the foreland's evolution (Figures 3, 13, 14).

POTENTIAL HYDROCARBON PLAYS

Three hydrocarbon sandstone plays are defined within LSTs of the Upper Cretaceous foredeep.

Delta-Front Play

Reservoir targets within the area's deltaic settings are delta-front sand wedges and turbidite-filled gullies (Figure 12).

The delta-front sand wedges of the Aren 1 sequence shale out in a downdip direction and are sealed by transgressive marls (Figure 10). The sealing potential of the red mudstones overlying the sand wedges of the other sequences may be reduced where there is increased silt and sand content. Potential hydrocarbon traps are synsedimentary anticlines (Aren 1 sequence, Figure 10), updip pinch-out into marsh deposits, and downdip erosive/structural closures where sand wedges are cut by gullies (Figure 12). Gully-fill turbidites, commonly in direct contact with delta-front sandstones, could have excellent reservoir properties. Base and top seals consist of prodelta shales.

Field mapping and stratigraphic data indicate that the delta-front sand wedges of each sequence communicate; however, within the different sequences these potential reservoirs are commonly compartmentalized.

Canyon-Fill Play

The submarine canyons are potential stratigraphic traps and their turbidite channels, mostly encased in shale, good reservoirs. Locally, thick accumulations of coarse-grained channel fills occur near the base of canyons and in canyon reentrants or bends; however, no channels larger than a few hundred meters have been found in outcrop.

Slope Channel-Fill Play

Thick-bedded, amalgamated slope channels, sealed by shale, could form excellent reservoirs. Examples occur in the Aren 1 and Aren 2 sequences (Figure 10). Gas tested from the Centenera 1 well is believed to have come from this facies (Figure 7, Table 1), which appears from seismic profiles to form an elongate belt (A. Garrido-Megías, 1998, personal communication). Turbidite channels interbedded with mudstone have excellent potential for sand-prone reservoirs, well sealed by shale. Net-to-gross ratios are up to 40%.

CONCLUSIONS

Based on the integration of field mapping, correlation of stratigraphic sections, and interpretation of seismic reflection data and subsurface well data, the following conclusions can be drawn from this study.

The orogenic front of the Upper Cretaceous foredeep of the south-central Pyrenees, known as the Bóixols thrust sheet, was formed by an imbricate sequence of three distinct thrust splays that become younger toward the foreland: Bóixols sensu stricto, Riu, and Turbón thrusts.

The stratigraphic evolution of the depocenters created by these growing anticlines is predictable on a depositional-sequence scale. During phases of uplift, thick LSTs (turbiditic deposits overlain by prograding deltaic wedges) accumulated on footwall synclines. During phases of relative tectonic quiescence, HSTs overlapped onto paleohighs. When the footwall syncline was filled, the depocenter shifted to the forelimb of the next structure.

Four distinct depositional sequences are recognized within the deltaic succession spanning the middle-late Campanian–late Maastrichtian. Relationships between sequences and growing anticlines imply that the cessation of deformation moved from east to west. Anticline growth took place in the late Santonian–early Campanian. Uplift of the Bóixols thrust sensu stricto ceased in the late Campanian, uplift of the Riu thrust in the latest Campanian, and uplift of the Turbón thrust in the early Maastrichtian. Foredeep deformation culminated in the latest Maastrichtian.

The present sequence-stratigraphic approach has led to the definition of three hydrocarbon plays within LSTs: delta front, canyon fill, and slope channel-fill. We believe that gas tested from 1960s vintage exploration wells was produced from slope turbidite channels. If these data are combined with the acquisition of new high-resolution seismic data, brighter prospects are believed to be in store for oil exploration in the southern Pyrenees.

This account of the evolution of the south Pyrenean foredeep should provide a useful analog for the petroleum exploration of prospective intraslope basins. In particular, the model illustrates how tectonics can affect facies distribution and how it can be used as a predictive tool to help locate sand-prone lithofacies.

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ABOUT THE AUTHORS

Lluís Ardèvol

Lluís Ardèvol is founder and director of GeoPlay (formerly Geoprep), a company specializing in field training for the petroleum industry since 1990. He received his M.S. degree in geology from the University of Barcelona, Spain, where he was assistant professor. Lluís has 17 years of experience in field geology and consultancy for oil, mining, and civil engineering companies. His research is in clastic facies analysis, sequence stratigraphy, and their application to reservoir characterization.

Juan Klimowitz

Juan Klimowitz is technical director of Gessal, a company specializing in subsurface exploration and interpretation. He received his M.S. degree in geology in 1982 from the University of Madrid, Spain, and has worked with Prakla-Seismos, the Spanish Geological Survey, and Repsol. Juan specializes in structural interpretation and has published several papers on thrust tectonics in the Pyrenees.

Jesús Malagón

Jesús Malagón is a geophysicist with Gessal. He received his M.S. degree in geology in 1980 from the University of Granada, Spain. He has worked for the Spanish Geological Survey, being mostly involved in field geology and subsurface research in the Pyrenees. His research interests include tectonics and basin analysis.

Peter J.C. Nagtegaal

Peter J.C. Nagtegaal is a consultant petroleum geologist. He received his Ph.D. in geology from the University of Leiden, The Netherlands, and has published a number of papers on the Pyrenees and the North Sea. With Shell Oil Company he has held the positions of head of exploration research and exploration manager in Malaysia and Colombia. His research is in Pyrenean geology and its petroleum applications.