LITHOSTRATIGRAPHY AND SEQUENCE STRATIGRAPHY OF THE UPPER THANETIAN TO MIDDLE ILERDIAN STRATA OF THE CAMPO SECTION (SOUTHERN PYRENEES, SPAIN): REVISION AND NEW DATA

A. Payros[1], V. Pujalte[1], J.I. Baceta[2], G. Bernaola[1], X. Orue-Etxebarria[1], E. Apellaniz[1], F. Caballero[1] and Carles Ferrandez[3]

[1] *Departamento de Estratigrafía y Paleontología, Facultad de Ciencias, Universidad del País Vasco, Apartado 644, E-48080 Bilbao.*
[2] *Departamento de Geología, Facultad de Ciencias Experimentales, Universidad de Huelva, La Rábida, E-21819 Huelva.*
[3] *Departament d'Estratigrafia i Paleontologia, Facultat de Geologia, Universitat de Barcelona, Zona Universitària de Pedralbes, E-08071 Barcelona.*

Abstract: The Campo section in the Central Southern Pyrenees has long been recognized as an important reference section for early Paleogene biostratigraphy of middle latitude successions. Yet, a comparison of the stratigraphic and sequence stratigraphic schemes so far proposed readily shows that an agreement about these basic aspects is still lacking. In this paper we introduce some refinements on the stratigraphy of the upper Thanetian-middle Ilerdian interval of the Campo section, and discuss the evolution of the area deducted from the new data. Further, we propose a new sequence stratigraphic interpretation of the same key interval. We show that the studied succession is made up of two different types of sequences, some of them induced by relative sea-level changes, the other forced by pulsating tectonism.

Key words: Thanetian, Ilerdian, lithostratigraphy, sequence stratigraphy, Pyrenees, Campo section.

Resumen: La sección de Campo, en la parte central de los Pirineos meridionales, está reconocida como una importante sección de referencia para la bioestratigrafía de las sucesiones del Paleógeno inferior en latitudes intermedias. Sin embargo, la comparación de los esquemas estratigráficos y secuenciales propuestos hasta la fecha demuestra que aún no se ha alcanzado un acuerdo sobre estos aspectos básicos. En este trabajo se detalla la estratigrafía del intervalo Tanetiense inferior-Ilerdiense medio de la sección de Campo y se discute la evolución del área en base a nuevos datos. Además, se propone una nueva interpretación secuencial del mismo intervalo. La sucesión estudiada está compuesta por dos tipos de secuencias, unas inducidas por variaciones relativas del nivel del mar, y las otras provocadas por un tectonismo pulsante.

Palabras clave: Tanetiense, Ilerdiense, litoestratigrafía, estratigrafía secuencial, Pirineos, sección de Campo.

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The Campo section has long been considered a key reference section for two main reasons: first, it allows the intercalibration of early Paleogene biostratigraphic scales of fossils from shallow and deep-water marine settings (i.e., larger foraminifers and calcareous plankton); second, it is a link between boreal and Tethyan paleobiogeographies. Both facts were aptly summarized by Schaub (1973), who wrote: «the Campo section is rather unique because here various groups of microfossils that are used to zone the Paleogene of western Europe and the Mediterranean area occur together in one continuous outcrop». Consequently, during the 60's and 70's the section was the object of numerous biostratigraphic studies, mostly focused on larger foraminifers (Hottinger and Schaub, 1960, Schaub, 1966, 1973), but also including planktic foraminifers (Hillebrandt, 1965), calcareous nannofossils (Wilcoxon, 1973; Kapellos and Schaub, 1973, 1975), dinoflagellates (Caro, 1973) and ostracods (Ducasse, 1972; Tambareau and Villate, 1974). Further, Schaub (1969) proposed the Campo section as the parastratotype of the Ilerdian stage, an early Paleogene stage based on larger foraminifers, still widely used to zone shallow-water carbonate successions in the Tethys realm (Hottinger and Drobne, 1998, and references therein). More recently Samsó *et al.* (1990), Tosquella *et al.*

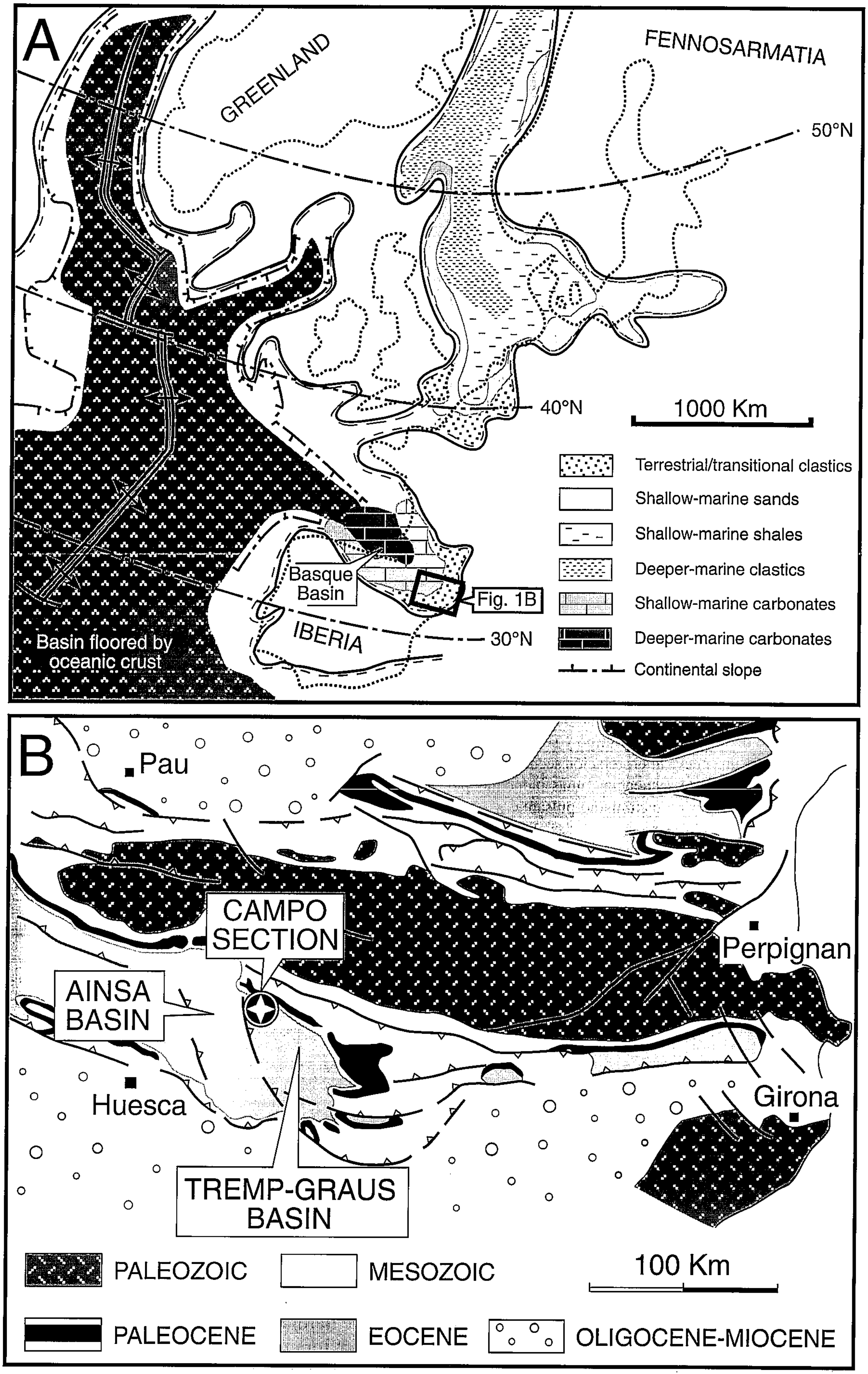

Figure 1.- A, Early Paleogene paleogeography of the North Atlantic (based on Ziegler, 1998), with location of the Campo area. B, Geological map of the Campo area.

(1990), Robador *et al.* (1991), and Tosquella (1995) revised the larger foraminifers of the section; Canudo (1990, 1991), Canudo *et al.* (1989), Arenillas and Molina (1995) and Molina *et al.* (1998, 2000) reexamined its planktic foraminifers. Finally, Molina *et al.* (1992) and Serra-Kiel *et al.* (1994) carried out integrated studies that included planktic and benthic foraminifers, calcareous nannofossils and magnetostratigraphy.

In contrast with these numerous biostratigraphic studies, comparatively few physical stratigraphic analyses have been undertaken. Mey *et al.* (1968), Garrido and Ríos (1972), Cuevas-Gonzalo *et al.* (1985) and Serra-Kiel *et al.* (1994) worked out the lithostratigraphic subdivisions of the succession, while sequence stratigraphic interpretations of the section were attempted by Mutti *et al.* (1985, 1988), Eichenseer (1987), Remacha and Zamorano (1989), Robador *et al.* (1991), Luterbacher *et al.* (1991) and Eichenseer and Luterbacher (1992). However, a comparison between these previous lithostratigraphic and sequence stratigraphic schemes shows important differences. Therefore, despite the paleontological importance of the Campo section, its physical stratigraphy is still unsettled.

The main purpose of this paper is to summarize and briefly discuss previous lithostratigraphic and sequence stratigraphic schemes, and to propose new ones for the upper Thanetian-middle Ilerdian interval. We have focused in this particular interval because it was then when the area evolved from shallow- to relatively deep-water conditions and, therefore, when benthic and planktic fossils occur together. In addition, sedimentological features relevant for the origin of the studied deposits are described.

Geological setting

During the early Paleogene the Pyrenean area was an E-W elongated gulf opening westwards into the paleobay of Biscay (Fig. 1a; see also Plaziat, 1981). The paleogeographic shape of this marine gulf remained relatively stable during Paleocene times, but evolved rapidly during early Eocene owing to active coeval tectonism (Muñoz *et al.*, 1986, 1998; Puigdefábregas *et al.*, 1986, 1992; Mutti *et al.*, 1988; Verges and Muñoz, 1990; Barnolas *et al.*, 1991; Teixell, 1998). During Paleocene and earliest Ilerdian times, the Campo area was situated in shallow-water (or even terrestrial) areas in the southeastern margin of the Pyrenean gulf (Fig. 1a). The intensity of compressional tectonics, generally low during Paleocene times, increased progressively during Ilerdian times, and eventually caused the subdivision of the area into different depocenters, such as the Tremp-Graus Basin and the Ainsa Basin (Fig. 1b) (Puigdefábregas *et al.*, 1986, 1992; Barnolas *et al.*, 1991; Muñoz, 1992). The former was a piggyback basin developed on top of the south-moving Cotiella-Montsec thrust sheet, and was mostly infilled by terrestrial and shallow-marine deposits. The latter was a foredeep turbiditic basin. The Campo area was located between these two basins, in a transitional zone situated next to an oblique lateral ramp of the South Pyrenean thrust-sheets (see also Mutti *et al.*, 1988, and Muñoz *et al.*, 1998).

The early Paleogene succession of the Campo section clearly reflects the general tectonic evolution of the zone. The Paleocene epoch is represented by a comparatively thin unit, which to the east passes rapidly into terrestrial deposits (Fig. 2a; see also Luterbacher *et al.*, 1991; Serra-Kiel *et al.*, 1994). However, the carbonates that made up the upper part of this unit at Campo (Navarri Formation in Fig. 2) pertained to an extensive carbonate shelf system, that stretched westwards with similar facies and thicknesses for more than 300 km (Baceta, 1996). In contrast, the Ilerdian succession is much thicker, reflecting an increased subsidence rate, and is typified by important lateral changes in facies and thicknesses (Fonnesu, 1984; Eichenseer, 1987; Eichenseer and Luterbacher, 1992; Serra-Kiel *et al.*, 1994). In the Campo section it reaches almost 900 m and is composed mostly of marine slope deposits with instability features. However, less than 10 km to the east, the Ilerdian succession is only 200 m thick and consists mostly of shallow-water deposits (including reefal facies), whereas to the west it passes rapidly into thick basinal turbiditic deposits of the Ainsa Basin (Fig. 2b).

The upper Thanetian-middle Ilerdian succession of the Campo section crops out in both sides of the N-S trending Esera river valley, about 2 km south of the Campo locality and 1 km to the northeast of Navarri, in the south central Pyrenees (see fig. 4.5 of Robador *et al.*, 1991). The succession is best seen in four closely-spaced exposures, respectively situated in the trenches of the roads Campo-Ainsa and Campo-Graus, in the local tract to Navarri, and in the banks of the Esera river. The litholog described below is in fact a composite column synthetisizing observations of the four exposures.

Lithostratigraphy

Previous studies

In a classical study of the area, Mey *et al.* (1968) subdivided the lower Paleogene succession into general lithological units of supposed chronostratigraphic significance. A specific lithostratigraphic scheme for the Campo section was developed by Garrido and Ríos (1972), who separated the lower Paleogene carbonate succession in two units (Laspun Dolostone and Navarri Limestone formations), and the overlying marly deposits in their Las Colladas and Morillo formations (Fig. 3). Nijman and Nio (1975) modified this scheme by introducing an informal

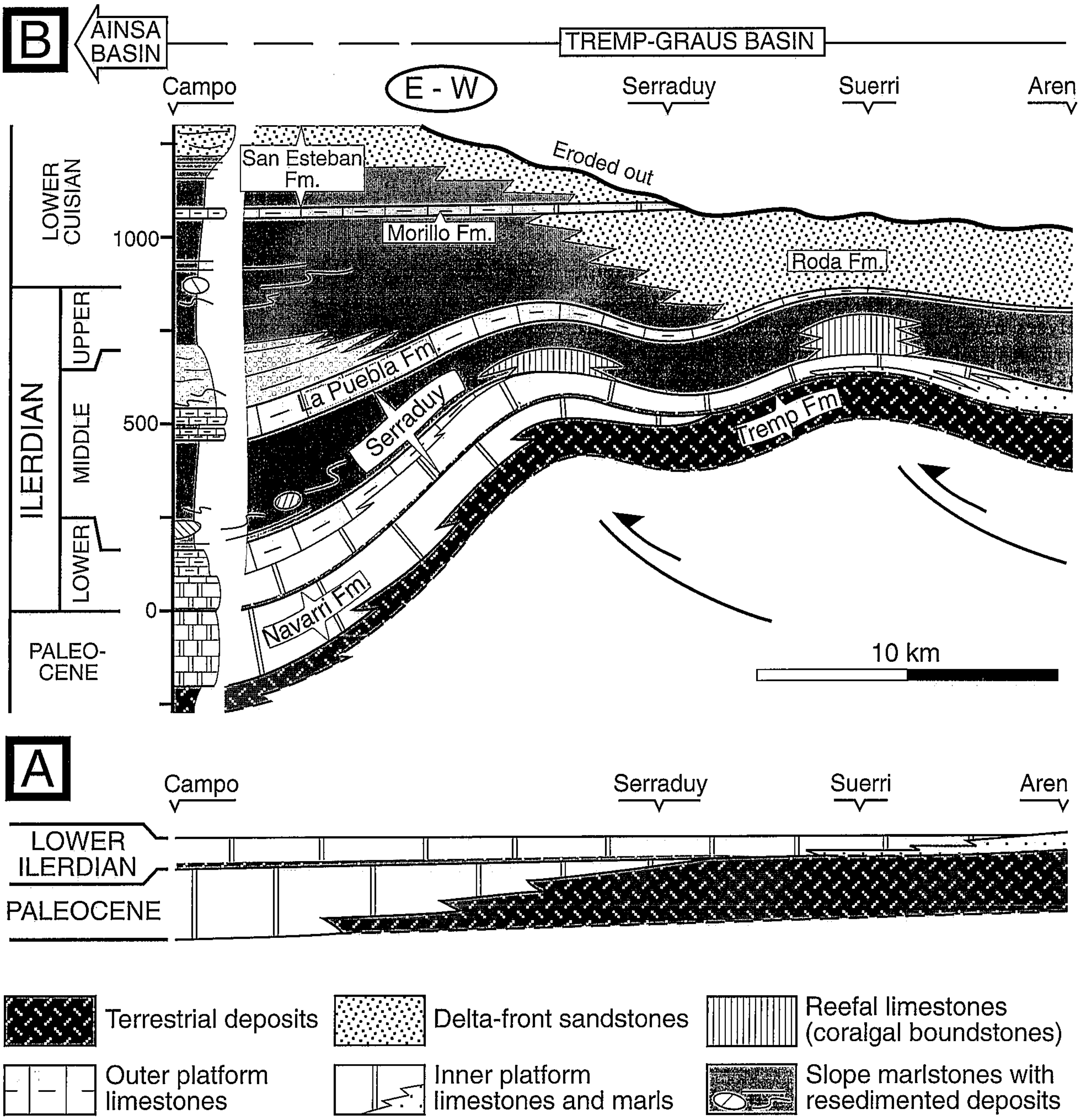

Figure 2.- Reconstructed E-W cross-sections showing stratigraphic relationships for the Paleocene-lowermost Ilerdian (A) and of the Ilerdian-Cuisian intervals (B) in the western part of the Tremp-Graus Basin. The Campo section is located in the transitional zone to the Ainsa Basin (see Fig. 1B). Based on Fonnessu, 1984; Eichenseer and Luterbacher, 1992; Serra-Kiel *et al.*, 1994; and our own data.

subdivision at the uppermost part of the Navarri Formation, named the «Alveolina Limestone» unit. In turn, Cuevas-Gonzalo *et al.* (1985) excluded the Alveolina Limestone from the Navarri Formation and included it within a new lithostratigraphic unit, named the Serraduy Formation. Accordingly, in the scheme of Cuevas-Gonzalo *et al.* (1985), the boundary between the Navarri and Serraduy formations would approximately coincide with the limit between the Thanetian and the Ilerdian Stages (Fig. 3).

Taking into account previous informations, Serra-Kiel *et al.* (1994) produced a more elaborated hierarchical scheme, in which the following units were distinguished (Figs. 2 and 3).

The Tremp Group is composed of the Tremp and Navarri formations. The Tremp Formation (ca. 60 m in Campo) is a fluviatile unit composed of cross-bedded sandstones and red mudstones, respectively interpreted as channel-fill and overbank deposits. The Navarri Formation is mostly made up of carbonate rocks, secondary dolostones in the lower part (~100 m) and of bioclastic limestones (locally rich in larger benthic foraminifers) in the upper one (~180 m). These carbonate rocks are attributable to inner platform settings, generally ranging from supratidal/intertidal conditions in the lower part to shallow subtidal conditions in the upper one.

The Ager Group is comprised of three main units, the Serraduy and La Puebla formations, plus an addi-

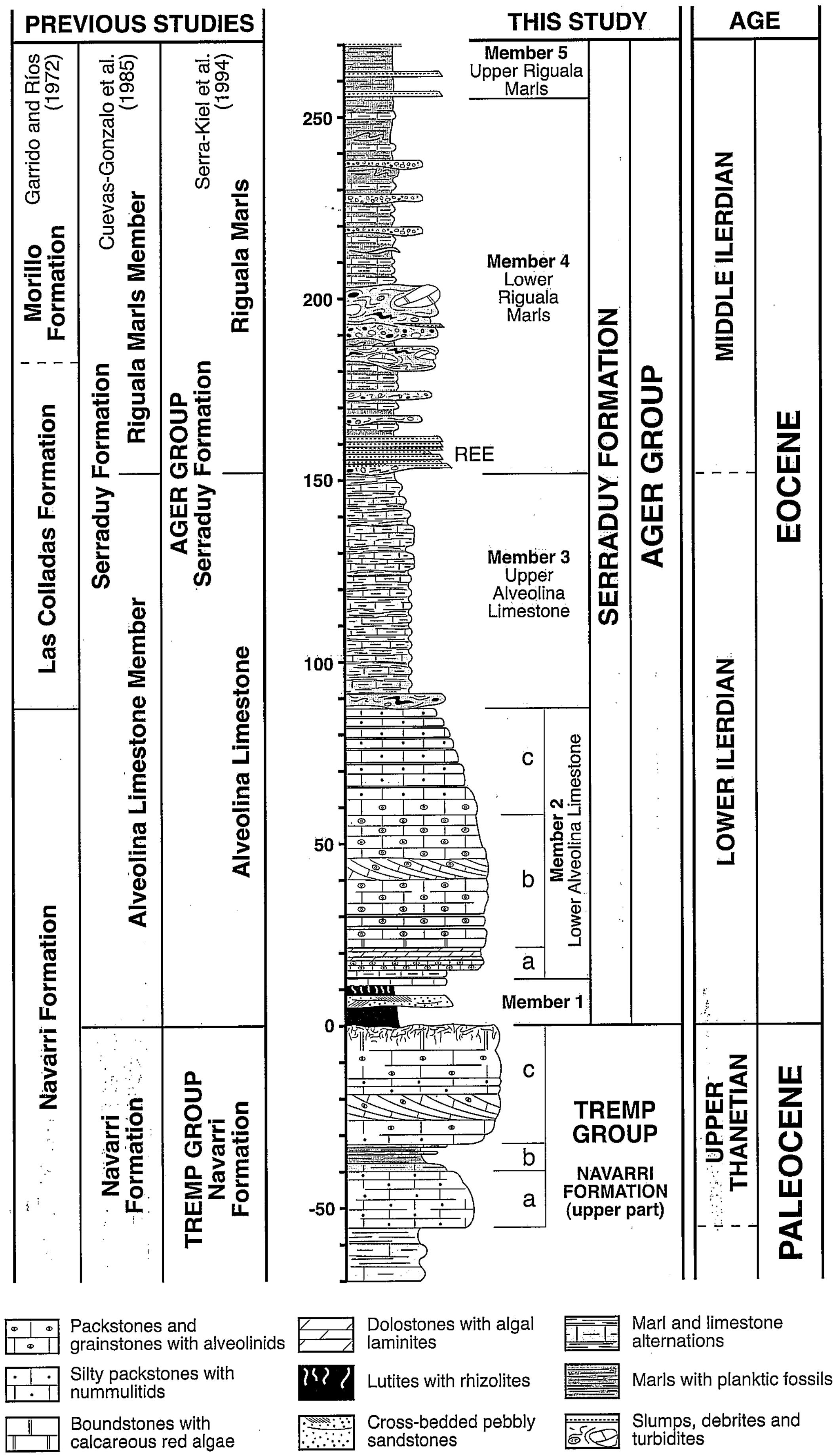

Figure 3.- Lithostratigraphic organization of the upper Thanetian-middle Ilerdian succession in Campo. Previous subdivisions are shown in the left-hand side, our own one in the right-hand side. Age data after Serra-Kiel *et al.* (1994).

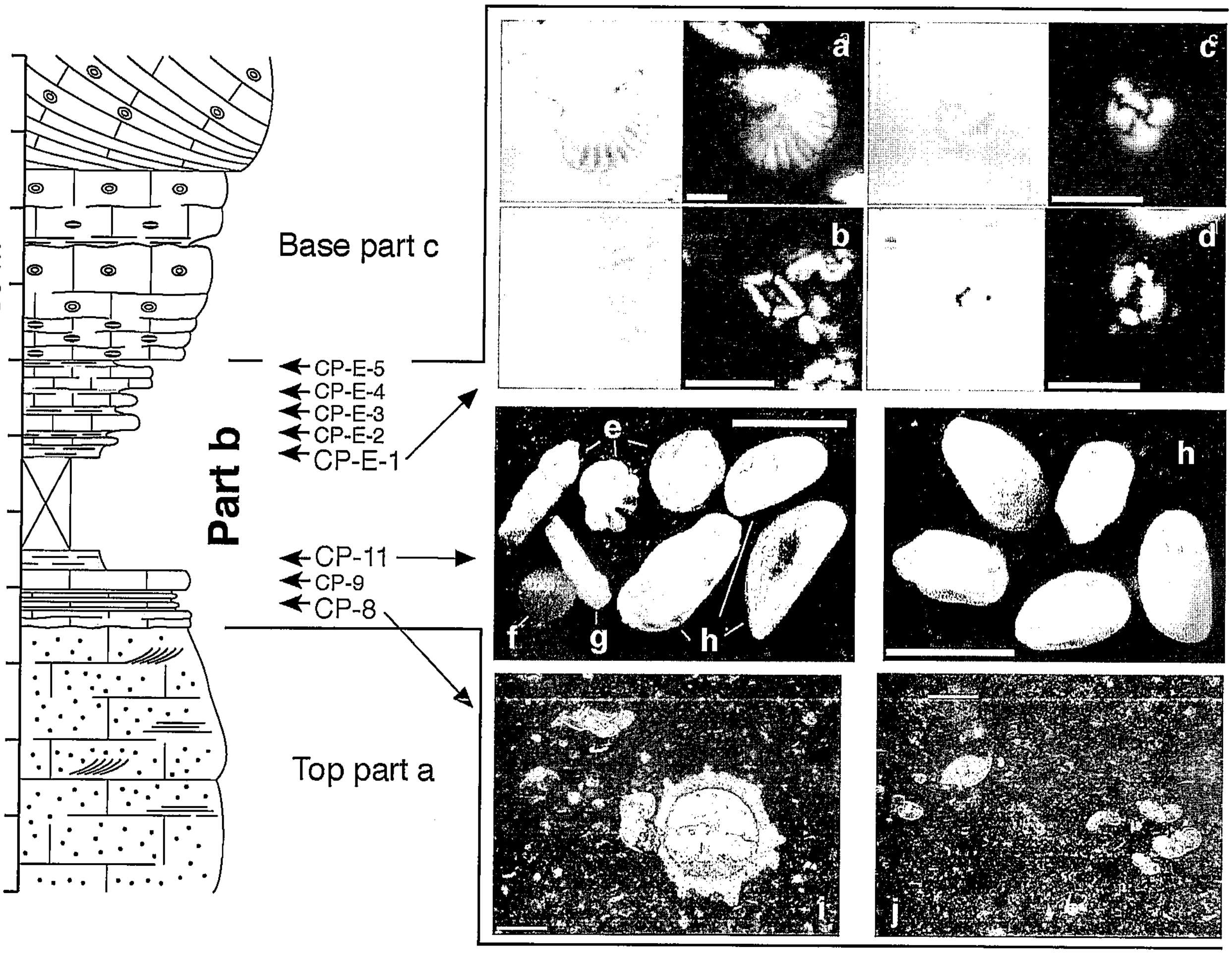

Figure 4.- Detailed section across part b of the upper package of the Navarri Formation. Open marine and terrestrial fossils occur together in this interval: a-d: calcareous nannofossils, respectively *Discoaster multiradiatus, Scapholithus apertus, Toweius pertusus* and *Coccolithus pelagicus*, indicative of the NP-9 Zone; e: benthic foraminifers; f: abraded charophyte oogonia; g: echinoid spine; h: ostracods; i: small foraminifers and possible fragmented charophyte stem in micritic matrix; j: foraminifer wackestone. Scale-bar 5 µm in a-d, 0.6 mm in e-h, and 0.3 mm in i-j.

tional unnamed unit laterally equivalent of the Roda Sandstone Formation, while the overlying Castigaleu Group consists of the Morillo and San Esteban formations (Figs. 2 and 3). The Serraduy Formation is a complex unit (see below), which records the transition from terrestrial to comparatively deep-water conditions. The La Puebla Formation is composed of alternating marly limestones and marls, occasionally slumped, probably laid down in a distally-steepened carbonate ramp. The overlying units, made up mostly of calcareous silty marls and sandstones, have been interpreted as fluvio-deltaic deposits fed from the rising Pyrenean chain to the northeast and reflect the progressive infilling and shallowing-up of the deep-water depression created during the accumulation of the Serraduy Formation.

The stratigraphic scheme developed by Serra-Kiel *et al.* (1994) has found general acceptance, being in fact used by the researchers currently working on the new 1:50000 MAGNA maps (Robador, *pers. com.*). Thus, we essentially follow the lithostratigraphic scheme of Serra-Kiel *et al.* (1994). We have focused our study in the interval comprising the upper part of their Navarri Formation and the middle part of their Serraduy Formation. However, we propose some modifications and subdivisions described below.

Revision of the Navarri Formation

The middle part of the Navarri Formation is represented by an alternation of marls and marly limestones with brackish-water ostracods, charophytes and occasional marine intercalations, which Tambareau and Villate (1974) interpreted as lagoonal deposits. This alternation is abruptly overlain by a 55 m thick package of marine deposits, in which three different parts are recognized, informally named here a, b and c (Fig. 3). Part a (14.5 m) is composed of partially dolomitized sandy packstones and grainstones, usually crudely bedded but occasionally with cross- and parallel-laminations that suggest current reworking. They contain miliolids, echinoderm plates, articulate coralline algal fragments and small rotaliids. Tambareau and Villate (1974) further reported occasional occurrences of *Assilina* sp., most probably equivalent to *Assilina yvettae*. The sedimentary structures and the faunistic assemblage of part a are indicative of relatively high-energy, shallow-marine conditions, likely representing a shoreface environment.

Part a and the underlying lagoonal marl/limestone alternation were jointly assigned to the *Alveolina (Glomalveolina) primaeva* Biozone (equivalent to Zone SBZ-3 of Serra-Kiel *et al.*, 1998) by Eichenseer (1987), Robador *et al.* (1991) and Serra-Kiel *et al.* (1994). However, if the equivalence of *Assilina* sp. and *Assilina yvettae* is confirmed, part a of the upper package of the Navarri Formation should be reassigned to the SBZ-4 Zone.

Part b is a comparatively thin intercalation (7 m) of micritic limestones (mudstones and wackestones), marls and marly limestones, which in the field appears as a distinct recessive unit between the more resistant parts a and c (Figs. 3 and 4). Thin sections of the micritic limestones (samples CP-8 and 9 in Fig. 4) have revealed the existence of small benthic foraminifers and broken charophyte stems. Washed residues of sample CP-11 have yielded a rich assemblage composed of miliolids, small rotaliids, algal fragments, ostracods, and deformed and/or abraded charophyte oogonia. The ostracod assemblage is composed of the genera *Thracella* (dominant), *Xestoleberis, Semicytherura, Paleomonsmirabilia, Oerthiella, Schizocythere, Hermainites, Pterygocythereis, Cytherella, Leguminocythereis, Occultocythereis, Bairdia* and *Paracypris*, which is characteristic of a sublittoral paleoenvironment (Ducasse *et al.*, 1985). Finally, smear slides from samples CP-11 and CP-E-1 to 5 consistently contain autochthonous assemblages of calcareous nannofossils.

Based on the occurrence of charophyte remains, some of the previous authors ascribed part b to lacustrine conditions, thus inferring a shallowing or regressive episode (e.g., Tambareau and Villate, 1974; Robador *et al.*, 1991; Serra-Kiel *et al.*, 1994). However, the bulk of the paleontological evidence outlined above rather suggests that part b represents the most open marine interval of the whole upper package of the Navarri Formation, indicative of a low-energy, relatively deep, outer ramp setting. We think, therefore, that the charophyte remains were transported into the depositional area by storm-related currents, a mechanism that would also explain their abraded and fragmented state.

Part c (32 m) mostly consists of thickly bedded bioclastic sandy packstones and grainstones, which in some cases exhibit large-scale cross stratifications (Figs. 3 and 4). They commonly contain abundant larger foraminifers, such as nummulitids and alveolinids, clearly attributable to the *A. (G.) levis* Biozone (i.e., SBZ-4 of Serra-Kiel *et al.*, 1998; see Serra-Kiel *et al.*, 1994). These organisms appear vertically arranged in several meter-thick sequences with nummulitids in their lower parts and alveolinids and miliolids in the upper ones (Fig. 4); these sequences are separated from each other by thin marly intervals with calcareous nannofossils. In the uppermost 6.5 m of part c the dominant fossils are crustose coralline algae (Fig. 3), which commonly develop rhodolithic structures, but coral and bryozoan fragments have also been observed. As a whole, the vertical fossiliferous changes, from larger foraminiferal grainstones to algae-dominated limestones suggest a shallowing-up character for part c. Furthermore, the top of the Navarri Formation is abrupt, represented by a sharp surface with abundant *Microcodium* remains, a clear indication of a long-lasting subaerial exposure (Tambareau and Villate, 1974; Robador *et al.*, 1991; Eichenseer and Luterbacher, 1992; Molina *et al.*, 1992).

Revision of the Serraduy Formation

This is a complex unit, both lithologically and sedimentologically (Figs. 2 and 3). To account for its variability, five different members have now been differentiated (numbered from 1 to 5). Member 1 is a comparatively thin, predominantly siliciclastic unit. Members 2 and 3, and members 4 and 5 respectively correspond to the lower and upper parts of the Alveolina Limestone and the Riguala Marl members of Serra-Kiel *et al.* (1994).

Member 1 is only well exposed in the trenches of the roads to Ainsa and to Navarri. It lacks indigenous marine fossils, but contains charophyte oogonia (Tambareau and Villate, 1974) and terrestrial palynomorphs (Núñez-Betelu *et al.*, 2000), an association clearly indicating non-marine conditions. Its lower boundary is the sharp, *Microcodium*-bearing surface that caps the marine carbonate deposits of the Navarri Formation. The character of this surface is a clear proof of an important relative fall of sea-level that led to the subaerial exposure and colonization by terrestrial plants of the former carbonate shelf deposits.

Member 1 is only 12 m thick, and consists of three parts. The lower part (5.5 m) is mostly made up of mottled greenish-grey calcareous lutites (interpreted as marsh soil deposits by Eichenseer and Luterbacher, 1992) with coalified plant remains and rare thin intercalations of limestones with abundant gastropods. The middle part (5 m) is composed of parallel- and cross-laminated calcareous sandstones (locally also granule- and pebble-sized conglomerates) that grade up into sandy siltstones with vertical rhizocretions, clearly a fluviatile, channel-fill sequence. The uppermost part is mostly represented by a thick bed of micritic limestone (1.25 m) overlain by greenish marls (0.25 m). The limestones contain abundant quartz grains and scattered fragments of charophyte stems, the latter indicating a fresh- or brackish water depositional environment. Thin sections of these limestones further reveal incipient pedogenic features, such as occasional circumgranular cracks and poorly-developed glaebulae. Member 1, therefore, records a temporary installation of terrestrial depositional conditions in the area.

Member 1 was attributed to the Thanetian in previous studies (Tambareau and Villate, 1974; Robador *et al.*, 1991; Serra-Kiel *et al.*, 1994). However, while the lower boundary of this terrestrial unit is abrupt, the upper one is conformable and gradational with the overlying Ilerdian shallow-marine limestones. Consequently, this siliciclastic unit has been transferred in this paper to the Ilerdian.

Member 2 gradually overlays Member 1, and records the return to marine conditions. Three parts (a, b and c in Fig. 3) are recognized. Part 2a (11 m) is dominated by dolostones, usually with algal laminations, but also includes packstones and wackestones with miliolids and alveolinids and oyster lumaquelles. Part 2b (35 m) is predominantly made up of thickly-bedded bioclastic limestones (grainstones and packstones), some of them with large-scale cross-bedding sets, and always rich in red algae and larger foraminifers, mainly alveolinids, but also miliolids and soritids. Part 2c (28 m) is composed of glauconitic marly limestones (packstones and wackestones), occasionally separated by thin marly intercalations. Nummulitids are the most abundant larger foraminifers of part 2c, although some alveolinids occur in the lower beds, while planktic foraminifers can already be found in the marly intercalations; bivalve shells are abundant constituents and crab shells have been occasionally observed.

Based on their respective lithological features and fossil content, Member 2a is depositionally thought to reflect inter- and supratidal flat conditions, Member 2b carbonate shoreface deposits and sublittoral sand bars (see also Eichenseer and Luterbacher, 1992), and Member 2c middle shelf, open marine conditions, as evidenced by the abundance of nummulitids and the presence of planktic foraminifers. Thus, the vertical depositional trend of Member 2, as a whole, demonstrates a transgressive character. This transgression took place during the early Ilerdian, as Member 2 accumulated during the time span of Zones SBZ-5 and 6 (Serra-Kiel *et al.*, 1994).

Member 3 (64 m) is equivalent to the upper part of the Alveolina Limestone Member of Serra-Kiel *et al.* (1994), but its features are very different to those of the underlying Member 2 (Fig. 3). The boundary between members 2 and 3 is abrupt, and it is marked by a slump 1 to 2 m thick (Fig. 3). Member 3 is entirely represented by a rhythmic alternation of grey marly limestones and marls. The marly limestones are thinly stratified (10-15 cm thick beds) and have a nodular aspect. The proportion of marls irregularly increases upwards, becoming dominant in the uppermost part of the unit. They include both small benthic and planktic foraminifers, but larger foraminifers are extremely scarce. On the basis of these lithological and paleontological features, Member 3 is assigned to an outer ramp or carbonate slope environment.

Member 4, the Lower Riguala Marl Member, is best exposed in the western bank of the Esera river and in the road to Aisa. It is 100 m thick and exhibits sedimentological features very different from those of the underlying Alveolina Limestone members 2 and 3 (Fig. 3). Its lower boundary is abrupt, being represented by a sharp surface that truncates the uppermost beds of Member 3. A mud supported debris-flow of variable thickness (0-1 m), with cm-sized clasts of Alveolina limestone, sits on this surface and is overlain by a 9-m-thick sub-unit composed of thin-bedded turbidites. In the field, this sub-unit forms a distinctive topographic feature, easily pinpointed because a big electrical transmission line pylon is planted on it. This pylon has a big sign reading REE (Red Eléctrica de España), an acronym used here to refer to this sub-unit. The REE turbidites generally have a mixed carbonate-siliciclastic composition, although some are almost pure bioclastic grainstones. Beds range in thickness between 5 and 35 cm, and in most cases exhibit base-missing Bouma sequences (either Tbc or Tcd). Their bases and tops are often bioturbated, attesting to periods of slowed or no sedimentation. However, marly interbeds are conspicuously absent and amalgamation between turbidite beds is frequent. The REE turbidites have a strike-parallel extent of just a few hundreds of meters and pinch out laterally, suggesting the confinement of the turbidite currents to a mainly by-passing depressed zone. They are therefore interpreted as the infilling of a shallow gully incised in the slope.

The remainder of Member 4 consists of an irregular alternation of (hemi)pelagic marls (background deposits) and resedimented deposits. Background deposits are either pure marls or well bedded mar/limestone alternations. They contain small benthic and planktic foraminifers plus calcareous nannofossils, but not larger foraminifers. They are thought to represent the autochthonous sedimentation in relatively deep, open marine conditions. In the middle and upper part of the succession, however, these background deposits may show instability features (slumping and sliding), and in some cases they have evolved to intraformational, mud-supported breccias.

Resedimented deposits of Member 4 include a whole range of carbonate gravitational flow accumulations, grading from cm-thick fine-grained turbidites to carbonate breccias up to 7 m thick. These resedimented deposits include a variety of shallow water fossils, such as alveolinids, nummulitids and rhodoliths of coralline algae. Corals also occur, either as isolated broken specimens or within limestone clasts that may reach several meters across. Some

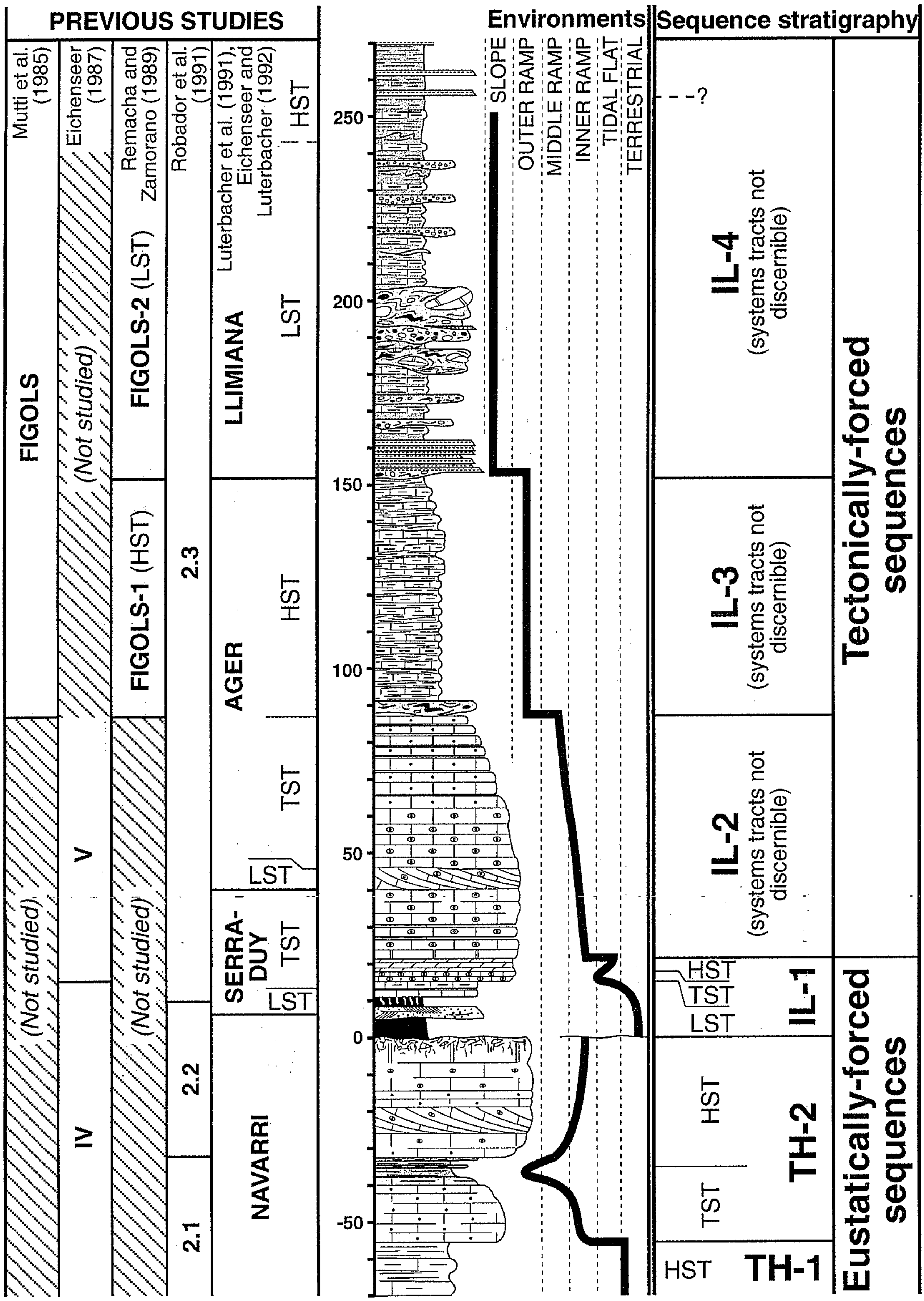

Figure 5.- Sequence stratigraphic schemes developed for the upper Thanetian-middle Ilerdian interval in the Campo section. Previous interpretations on the left-hand side; our own, in the right-hand side. Explanation within the text.

of the largest clasts are plastically deformed, a clear evidence that they were shed into the basin before fully indurated. The fossil content indicates a middle Ilerdian (SBZ-7) age for these deposits (Serra-Kiel *et al.*, 1994, 1998).

Member 5 is discontinuously exposed along the road to Ainsa and in the western side of the Esera valley. By far, the most abundant deposits of this unit are dark-grey or bluish marls and silty calcareous shales with open marine fossils, including planktic foraminifers and calcareous nannofossils (Fig. 3). The only other deposits of some volumetric importance are plane-parallel siliciclastic turbidites, that in the lower part of the unit occur as isolated beds up to 90 cm thick. Upwards in the succession (not included in this study) turbidite beds become progressively thinner and less frequent.

Significance of the lithostratigraphic succession

The vertical facies and fossiliferous changes described above can be used to reconstruct the evolution of the Campo area from late Thanetian to middle Ilerdian times (Fig. 5). The onset of part a of the upper Navarri package records a change from a lagoonal to shallow-water, high energy, fully marine conditions, probably indicating a rapid transgression. This transgression peaked during the deposition of part b, when open marine conditions were established. Part c records the inversion of the previous trend, being typified by a progressive shallowing-up that culminated in the subaerial exposure of the upper Navarri carbonate shelf, a halt in deposition and the encroachment of plant communities. The extent of the inherent hiatus, however, cannot be ascertained.

Sedimentation resumed in terrestrial conditions, during earliest Ilerdian times, with the deposition of Member 1 of the Serraduy Formation. Soon after that, the area was again flooded by the sea, and the Alveolina Limestone was deposited. As a whole, Member 2 of the Serraduy Formation (lower part of the Alveolina Limestone) records a progressive change from nearshore environments (part a) to middle ramp (part c), a clear indication of a relative sea-level rise. The area, nevertheless, still remained in comparatively shallow waters, as evidenced by the continuous occurrence of larger foraminifers.

Member 3 of the Serraduy Formation records a dramatic paleoenvironmental change, which resulted in the establishment of aphotic deep-water conditions in an outer ramp setting (Fig. 5). The rapid transition between members 2 and 3 of the Serraduy Formation, and the presence of the slump bed at the base of Member 3, point to tectonism as the most likely mechanism driving this change.

Deposition of Member 4 of the Serraduy Formation implies another abrupt change. The REE turbidites and the overlying carbonate breccias and turbidites record deposition in a deep-water area. Further, the mixed siliciclastic and carbonate nature of the REE turbidites record sourcing from adjacent emergent terrains and from coeval shallow-water reefs into a slope setting. As a whole, the features of Member 4 suggest a rapid deepening of the Campo area and the uplift of adjacent zones, induced most likely by strong tectonic activity. Member 5, in turn, records the progressive cessation of resedimentation processes and the stabilization of the depositional area.

Sequence stratigraphy

In the interval of the succession here revised, 3rd-order depositional sequences were described by Mutti *et al.* (1985), Eichenseer (1987), Remacha and Zamorano (1989), Robador *et al.* (1991), Luterbacher *et al.* (1991) and Eichenseer and Luterbacher (1992). A comparison between them readily shows that conflicting interpretations still remain (Fig. 5). The discrepancies concern to both the number of depositional sequences and to the location of their bounding surfaces. However, most previous authors did suggest a good correlation between their sequences and the global sea-level changes predicted by Haq *et al.* (1987), therefore implying an eustatic imprint.

Our own results should be considered still preliminary, as they are based exclusively on the study of the Campo section. We have recognized six depositional sequences, two of them of Thanetian age, the other four of Ilerdian age. They have been coded with the initial letters of their age followed by a number indicating their correlative order (i.e., TH-1, TH-2, IL-1, IL-2, IL-3 and IL-4). Sequences TH-1, TH-2 and IL-1 differ in some significant points with sequences IL-2, IL-3 and IL-4, and therefore the two groups of sequences are described separately.

Late Thanetian and earliest Ilerdian sequences

Only a part of sequence TH-1 is included in the studied interval, namely the middle Thanetian lagoonal deposits, that are attributed to a late highstand systems tract (LHST). Depositional sequence TH-2 comprises the remainder of the upper Thanetian succession (Fig. 5). Its lower boundary corresponds to the base of part a of the upper package of the Navarri Formation, and it is considered a transgressive surface. Part a is assigned to the transgressive systems tract (TST) because of its marine, deepening-up character, while part b is considered to contain the maximum flooding surface. Part c, in turn, is assigned to the HST because of its overall shallowing-up character. Further, the shallowing-upwards calcarenitic units separated by marly intercalations are considered to represent stacked parasequences bounded by flooding surfaces. The occurrence of larger foraminifers in these calcarenites suggests oligo-mesotrophic conditions. The coralline algal limestones of the uppermost 6.5 m of part c suggest an increasing influx of nutrient rich waters of continental derivation, linked perhaps to the late HST progradation.

The lower boundary of sequence IL-1 coincides with the *Microcodium*-bearing surface capping the Navarri Formation. This sequence comprises both Member 1 and part a of Member 2 of the Serraduy Formation. Member 1 is provisionally interpreted as the LST, because of its terrestrial character. However, it could also represent an incised valley fill, and hence be part of the TST. Member 2a opens with two alveolinid-rich limestone beds, interpreted as the TST (or, perhaps, the maximum flooding surface) and then it grades up into tidal-flat algal limestones and dolostones, ascribed to the HST (Fig. 5). Sequence IL-1 approximately corresponds with the Serraduy sequence of Luterbacher *et al.* (1991) and Eichenseer and Luterbacher (1992), although the position of the respective sequence boundaries is different in each case (Fig. 5).

Despite having very different facies and vertical development, the three sequences described above share some distinct features: they reflect successive deepening and shallowing-up trends, they are bounded by transgressive or subaerial exposure surfaces, and they allow the recognition of systems tracts. In other words, they can be interpreted using the «Exxon model» of sequence stratigraphy.

Ilerdian sequences

One inspection of figure 5 readily shows that our sequence stratigraphic interpretation of the upper Thanetian-lowermost Ilerdian interval differs from that of previous authors in the horizons selected as sequence boundaries. In contrast, for the lower-middle Ilerdian interval we generally coincide with previous authors in the position of the sequence boundaries, although we still differ in the interpretation of the origin of the sequences themselves.

The lower limit of depositional sequence IL-2 corresponds to a laterally continuous oyster-shell accumulation at the base of a coralline algal boundstone bed, situated in the lowermost part of member 2b of the Serraduy Formation (Fig. 5). Both the shell accumulation and the overlying algal biostrome were probably formed after an episode of reworking by high-energy currents and a relative rise of sea level. Therefore, the lower boundary of sequence IL-2 is coincident with a ravinement surface, suggestive of a rapid transgression. The lower boundary of sequence IL-3 is placed at the abrupt facies change between members 2 and 3 of the Serraduy Formation (Fig. 5), and coincides with the base of the Figols-1 sequence of Remacha and Zamorano (1989). The lower boundary of sequence IL-4 is situated at the base of the REE turbidites, a limit that also marks the base of the Figols-2 and Llimiana sequences of Remacha and Zamorano (1989) and Eichenseer and Luterbacher (1992), respectively.

Despite the ample consensus about the sequence boundaries, the Ilerdian succession is not easy to interpret following Exxon's sequence stratigraphic concepts. On the one hand, none of these sequence bounding surfaces show evidence of subaerial exposure. More important, the strata between successive sequence boundaries do not show the transgressive-regressive trend typical of eustatically-forced sequences, making the delineation of systems tracts difficult or even impossible. Thus, sequence IL-2 is entirely made up of larger foraminifer bearing limestones with a deepening-upward character, suggestive of a TST. This interpretation, however, would imply that the sequence was composed of a single systems tract. The slumped bed at the lowermost part of Member 3 records a rapid deepening of the Campo area and the coeval onset of gravitational instabilities leading to episodic resedimentation. Otherwise, the outer ramp marls and marly limestones of Member 3 of the Serraduy Formation make up the bulk of sequence IL-3 and, as stated by Remacha and Zamorano (1989), differentiation of systems tracts in this succession is hampered by its monotonous character. Finally, the resedimented deposits in the lower part of sequence IL-4 led previous authors to consider them as belonging to the LST (Fig. 5). However, it must be taken into account that during the formation of this part of the sequence the Campo area was submerged at a greater depth than in previous times, indicating a sea-level rise rather than a drop.

Comparison with the western Pyrenees

Further insights about the significance and the factors controlling the development of the late Thanetian-middle Ilerdian 3rd-order sequences in the Campo section can be gained through a comparison with coeval sequences occurring in the western Pyrenees.

Campo depositional sequences TH-1, TH-2 and IL-1 can be also recognized in the western Pyrenees and in the Basque Basin (see Baceta, 1996; Pujalte *et al.*, 2000). Their extent is a clear evidence of subdued tectonism and homogeneous subsidence rates along the whole southern margin of the Pyrenean gulf during the late Thanetian and earliest Ilerdian times. Furthermore, the sea-level changes that created these sequences are also reported from the North Sea area (Neal, 1992; Pujalte *et al.*, 1998) and probably from the SE and E margins of North America (Baum and Vail, 1988; Davidoff and Yancey, 1993; Mancini *et al.*, 1995). It can thus be concluded that the sea-level changes that created the sequences TH-1, TH-2 and IL-1 were at least of regional and perhaps of even global (eustatic) extent.

In contrast, the lower-middle Ilerdian sequences are highly variable both at the local scale and regional scales (see facies and thicknesses changes in Fig. 2). As a whole, the lower-middle Ilerdian successions in this part of the south central Pyrenees record a deepening-upward trend related to an overall transgression. Interestingly, contemporary deposits in the western Pyrenees display an aggradational to slightly progradational character, indicating a general regression (Payros, 1997; Pujalte *et al.*, 2000). Therefore, it is likely that the deepening-upward trend in the Campo section is a local feature, reflecting an accelerated tectonic subsidence in the study area. Besides, the Ilerdian depositional sequences in Campo (IL-2 to IL-4) cannot be correlated with coeval depositional sequences developed in other areas of the Pyrenean domain (see Payros, 1997; Pujalte *et al.*, 2000). All these facts indicate that the local tectonic imprint in Campo was strong enough to hide the eustatic signal.

Concluding remarks

The upper Thanetian to middle Ilerdian interval is one of the most important segments of the Campo section. During its deposition a change from shallow- to (comparatively) deep-water conditions took place in the area, making it possible the coexistence of larger foraminifers, planktic foraminifers and calcareous nannofossils. These are the fossils best suited for the biozonation of early Paleogene successions, and their co-occurrence allows the intercalibration of their respective biostratigraphic schemes across the Paleocene/Eocene boundary interval, a crucial time in the evolution of the biosphere (Schmitz *et al.*, 2000).

In this paper we have revised previous lithostratigraphic divisions of that important interval and we have proposed some refinements to the lithostratigraphic scheme of Serra-Kiel *et al.* (1994), the most accurate account of the lower Paleogene succession of the Campo area so far available. Our modifications concern mostly to the Serraduy Formation, specially to the chronostratigraphic placement of its lower boundary and to its subdivision in members. A similar revision of the Navarri Formation is beyond the scope of this paper, although we have provided some new information concerning this unit.

On the other hand, six unconformity-bounded depositional sequences have been recognized in the studied succession. Those of the Paleocene and earliest Ilerdian times (TH-1, TH-2 and IL-1) fit with the Exxon's sequence stratigraphic model, being bounded by transgressive or subaerial exposure surfaces and having transgressive-regressive vertical trends. They are though to record eustatically-driven sea-level changes, a possibility reinforced by comparison with coeval sequences of the western Pyrenees and elsewhere. In contrast, the Ilerdian sequences (IL-2, IL-3 and IL-4) are bounded by sharp surfaces recording abrupt deepening events of the depositional setting, linked to episodic increases in the rate of tectonic subsidence of the Campo area. These Ilerdian sequences, therefore, have a more reduced lateral extent than those of previous times, although they are still invaluable to reconstruct the local evolution of the Campo area. We must stress, however, that our sequence stratigraphic analysis should be considered preliminary, as it is solely based in the study of one single section. Ongoing research in the area will test its validity, and will determine the opportunity of further changes.

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