Geologica Acta, Vol.1, Nº2, 2003, 177-200
Available online at www.geologica-acta.com

Marine and Transitional Middle/Upper Eocene Units of the Southeastern Pyrenean Foreland Basin (NE Spain)

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ABSTRACT

The stratigraphic basis of this work has allowed the use of larger foraminifers in the biostratigraphic characterisation of the new Shallow Benthic Zones (SBZ). This part of the volume presents a description of the sedimentary cycles formed by the transgressive-regressive systems of the Lutetian and Bartonian in the southeastern sector of the Ebro Foreland Basin. Concerning the Lutetian deposits studied in the Amer-Vic and Empordà areas, four sedimentary cycles have been characterised. The first and second are found within the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972), while the third and fourth cycles cover the Coll de Malla Marl Formation (Clavell et al., 1970), the Bracons Formation (Gich, 1969, 1972), the Banyoles Marl Formation (Almela and Ríos, 1943), and the Bellmunt Formation (Gich, 1969, 1972). In the Bartonian deposits studied in the Igualada area, two transgressive-regressive sedimentary cycles have been characterised in the Collbàs Formation (Ferrer, 1971), the Igualada Formation (Ferrer, 1971), and the Tossa Formation (Ferrer, 1971). The Shallow Benthic Zones (SBZs) recognised within the Lutetian are the following: SBZ 13, from the Early Lutetian, in the transgressive system of the first cycle; SBZ 14, from the Middle Lutetian, in the second cycle and the lower part of the transgressive system of the third cycle; SBZ 15, from the Middle Lutetian, in the remaining parts of the third system; SBZ 16, from the Late Lutetian, throughout the fourth cycle. The association of larger foraminifers in the first and second cycles of the Bartonian in the Igualada area has been used as the basis for the definition of SBZs 17 and 18 recognised in the Bartonian of the western Tethys.

KEYWORDS **Marine units. Middle/Upper Eocene. Lithostratigraphy. Biostratigraphy. Pyrenean basin.**

INTRODUCTION

The lithostratigraphy and biostratigraphy of the Middle Eocene (Lutetian and Bartonian) in the eastern sector of the South Pyrenean Foreland Basin (Ebro Basin), is well-known thanks to the works of previous authors. The regional lithostratigraphic-sedimentological studies by Reguant (1967), Kromm (1967), Ferrer (1971), Gich

(1972), Pallí (1972), Estévez (1973), Anadón (1978a), Colombo (1980), Busquets (1981) and Taberner (1983-84), together with the biostratigraphic studies by Ruíz de Gaona (1952), Ruíz de Gaona and Colom (1950), Hottinger (1960), Reguant (1967), Colom (1971), Ferrer (1971), Caus (1975), Schaub (1981), Serra-Kiel (1984) and Tosquella (1995), allowed the drawing of the first schematic stratigraphic correlations between the marine and transitional deposits of the Ebro Foreland Basin (Kromm, 1968; Gich, 1972; Rosell et al., 1973; Riba, 1975).

Despite these numerous stratigraphic, sedimentologic, and biostratigraphic studies carried out on the autochthonous Paleogene deposits of the South Pyrenean Foreland Basin, few authors have attempted an analysis based on the differentiation of stratigraphic units limited by allocyclic events, and thus identifiable throughout the basin. The difficulty in undertaking such an analysis lies in the important influence of tectonic activity on the sedimentation of the basin, especially in the northern part of the area, and in the effects of tectonic processes which subsequently truncated the physical continuity of the deposits. In this sense, the studies by Puigdefàbregas et al. (1986) and Barnolas (1992) use sequential analysis to divide the Paleogene marine deposits of the South Pyrenean Basin into depositional sequences, genetically related to the development of thrust sheets on the active margin (allochthonous Pyrenean units).

In this paper we use term ‘sedimentary cycles’ in the sense of Johnson et al. (1985) and Bates and Jackson (1987), used in the previous works on the Ebro Foreland Basin by López-Blanco (1996) and López-Blanco et al. (2000). These cycles comprise all deposits laid down in a basin between the beginning of two transgressive systems. These cycles are equivalent to the depositional sequences of Vail et al. (1984), Posamentier and Vail (1988) and Posamentier et al. (1988), without lowstand diposits that do not outcrop in the marginal parts of the basin, such as those studied here. In a marginal context, the transgressive and regressive systems are easy to identify from the sharp contacts found at lithologic changes, reflected both at stratigraphic and cartographic levels. Thus, we have differentiated six cycles. Four cycles are Lutetian in age and two are from the Bartonian.

The aim of this paper is to offer a synthesis of the stratigraphic (including lithostratigraphic and biostratigraphic) features of the marine deposits of the Middle Eocene (Lutetian and Bartonian) in the eastern sector of the South Pyrenean Foreland Basin (Ebro Basin). The chronostratigraphic results obtained from the correlation between the planktic foraminifer biozones, the larger foraminifers and the magnetostratigraphy were used to define the larger foraminifer biozones (Shallow Benthic Zones) published by Serra-Kiel et al. (1998a, b).

However, the synthesis presented here is not only based on a compilation of the existing bibliography. It also draws on the map and stratigraphic information included in the 1:25.000 scale maps produced by the Institut Cartogràfic de Catalunya -*Servei Geològic de Catalunya*- (Saula et al., 1994a; Mató et al., 1995a, b; Picart et al., 1995; Mató et al., 1996; Pi et al., 1997; Pi et al., 2000; Puig et al., 1997a, b; Saula et al. 1997; Losantos et al., 2000, 2001; Martínez et al., 2000), and on the lithostratigraphic and biostratigraphic work of the members of the Spanish Working Group of the IGCP 393.

The list of genera and species cited in this issue is included as Appendix II at the end of the volume.

GEOLOGICAL SETTING

Structure

The Pyrenees correspond to the western termination of an orogenic belt formed during the Tertiary due to clousure of the Tethyan Sea located between the converging African and European plates (Fig. 1). The relatively straight and E-W-trending Pyrenean orogen developed because of the northward subduction of the Iberian lithosphere beneath Europe as imaged by the deep seismic reflection ECORS-Pyrenees profile (Choukroune et al., 1989). The Pyrenees merge eastward with the highly arcuate Alps, which formed above a southward European lithospheric subduction underneath Adria, the northern promontory of the African plate (ECORS-CROP Deep Seismic Sounding Group, 1989). Both orogens are doubly sided with a major foreland basin on top of the lower plate (e.g. Muñoz, 1992; Pfiffner, 1992). The Ebro Foreland Basin and the Molasse Basin represent the latest evolutionary stage of the flexural foreland basin. Earlier foreland-basin strata are preserved in piggyback basins on top of thrust sheets. The Aquitaine Foreland Basin developed on the northwestern side of the Pyrenees and represents a retro-foreland basin.

The southeastern side of the Ebro Basin displays an irregular shape bounded by the Pyrenees to the north and the Catalan Coastal Ranges to the southeast (Fig. 2). This irregular geometry is due to both the oblique trend of the Catalan Coastal Ranges with respect to the Pyrenean chain and the succession of frontal and oblique segments of the Pyrenean front, inherited from the Mesozoic extensional-basin geometry (Puigdefàbregas et al., 1992; Vergés and Burbank, 1996). The Southeastern Pyrenean Foreland Basin developed almost entirely over pre-Mesozoic basement and stands in contrast to the more westerly Jaca Basin which developed on top of a detached Mesozoic section (e.g. Séguret, 1972; Teixell, 1996).

The Vallfogona Thrust represents the major thrust boundary between the southeastern Pyrenean thrust sheets

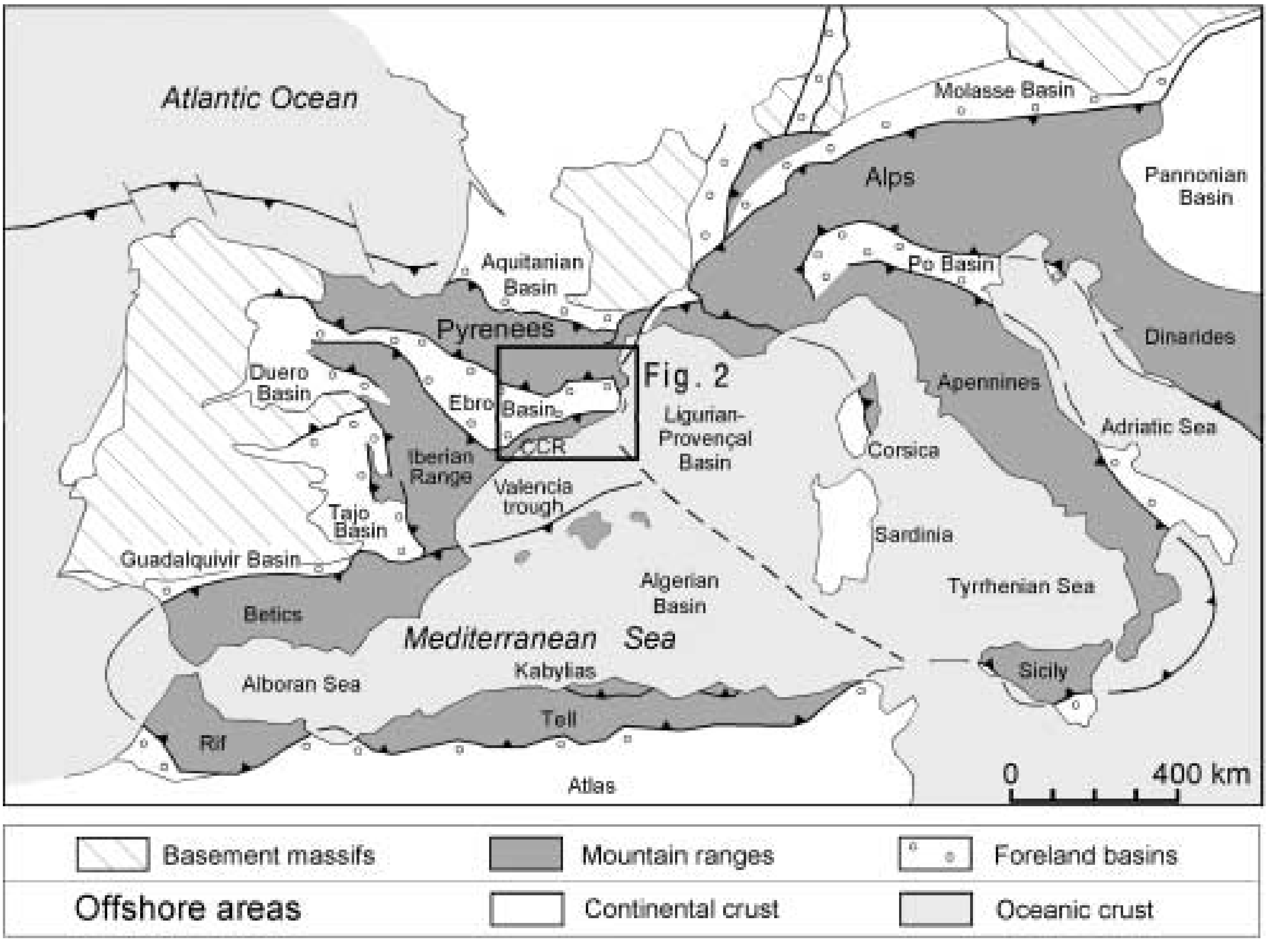

FIGURE 1 **Tectonic sketch of the Western Mediterranean showing the location of the Tertiary orogenic belts and adjacent foreland basins according to Vergés et al. (1998). Thick black lines with black triangles bound shaded orogenic regions. C. C. R.= Catalan Coastal Ranges.**

and the deformed foreland strata (Fig. 2). The thick Tertiary succession cropping out in the Cadí thrust sheet is folded at the Ripoll syncline (Muñoz et al., 1986; Puigdefàbregas et al., 1986). The lower segment of this stratigraphic succession represents the northern part of the former Southeastern Pyrenean Foreland Basin whereas the upper segment represents the individualization of this part of the trough as a piggyback basin, the Ripoll Basin, after the inception of the Vallfogona thrust. A large part of the Ebro Foreland Basin is deformed by a system of folds and thrusts, in part coeval to deposition (Puigdefàbregas et al., 1986; Burbank et al., 1992a), and detached above a suite of foreland evaporitic levels (Vergés et al., 1996; Sans et al., 1996).

Stratigraphy

The Paleogene marine succession of the Southeastern Pyrenean Foreland Basin has been divided into four major tectostratigraphic units in the sense of Weller (1958) and modified after Vergés et al. (1998), (Fig. 3).

The first tectostratigraphic unit is Ilerdian to Early Cuisian in age and comprising SBZ 5 to SBZ 10, based on magnetostratigraphy (Serra-Kiel et al., 1994; Bentham and Burbank, 1996) correlated with biostratigraphic data (Serra-Kiel et al., 1994). This first unit started with a transgressive system composed of the shallow-marine *Alveolina* limestone of the Cadí Formation (Mey et al., 1968), Ilerdian and Early Cuisian in age. The age of the base of the Cadí Formation decreases toward the distal margin of the basin: Early Ilerdian in the Pyrenean thrust sheets, Middle Ilerdian in the centre of the Ebro Foreland Basin and Middle Ilerdian in the southern and distal margin (Orpí Formation, Ferrer, 1971). The shallow-marine carbonate shelf represented by both the Cadí and Orpí formations merged northward into the deeper-water calcareous mudstone of

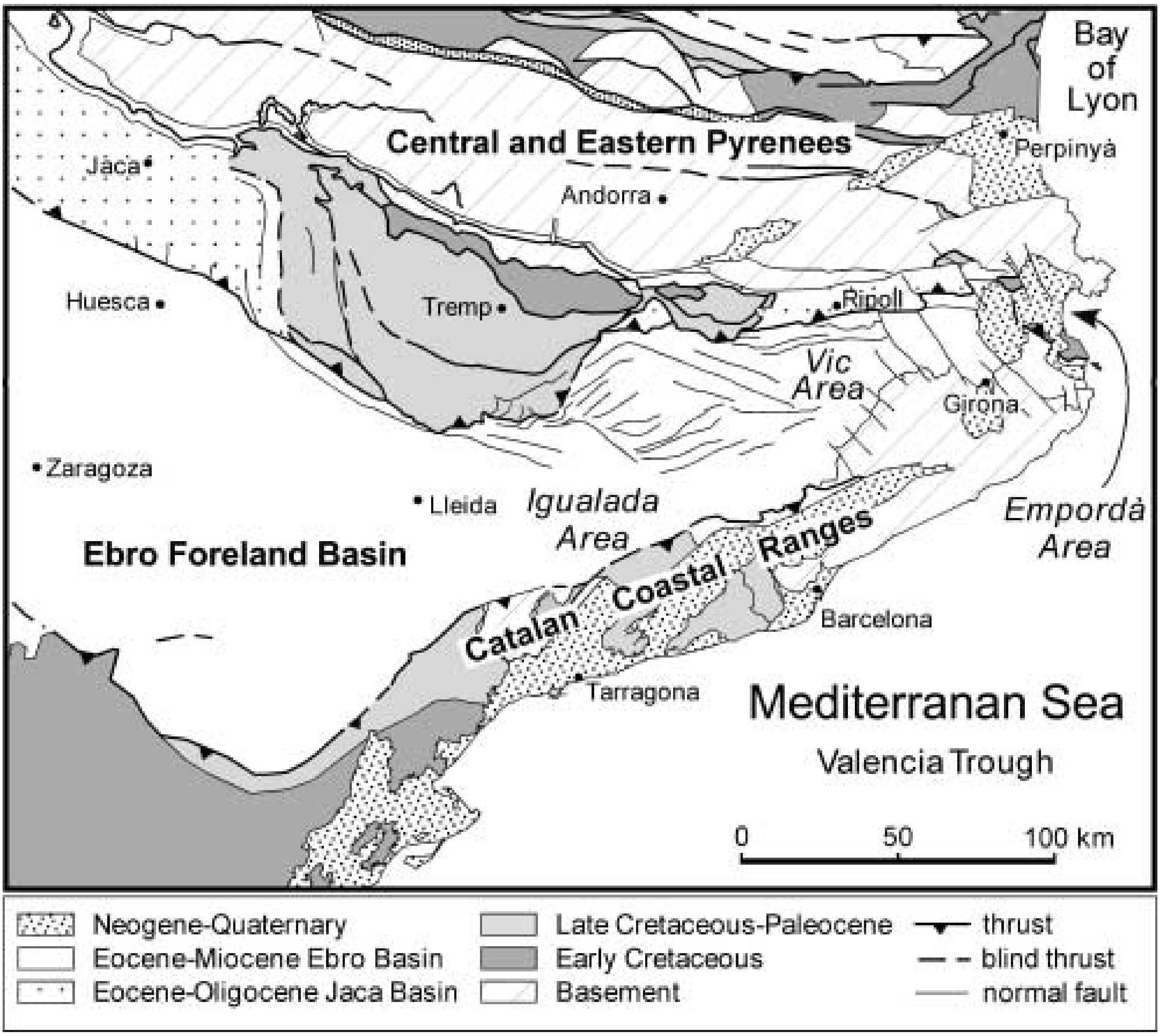

FIGURE 2 **Geological sketch of the eastern part of the Ebro Foreland Basin between the Pyrenean thrust belt to the north and the Catalan Coastal Ranges to the south according to Vergés et al. (1998).**

the Sagnari Formation (Gich, 1969, 1972; Giménez-Montsant, 1993). The regressive hemicycle is represented by a prograding deltaic clastic wedge capped by fluvio-lacustrine facies of the Corones Formation (Gich, 1969, 1972; Giménez-Montsant, 1993), Early Cuisian in age (Tosquella, 1995; Tosquella and Samsó, 1998).

The second tectostratigraphic unit is Middle Cuisian to Early Lutetian in age and contains SBZ 11 to SBZ 13, based on magnetostratigraphy (Bentham and Burbank, 1996) correlated with biostratigraphy (Samsó et al., 1994; Tosquella, 1995; Tosquella and Samsó, 1998). This unit starts with shallow-marine limestone of the uppermost part of the Corones Formation, which rapidly grade upward into a 1000 m thick, slope facies association mostly represented by carbonate and calcareous mudstone, often slumped from the Armàncies Formation (Gich, 1969, 1972). This slope association includes up to seven decametric megabreccia units interpreted as the result of slope resedimentation of coeval carbonate shelf. They constitute the Penya Formation (Estévez, 1970, 1973; Giménez-Montsant, 1993). In the northern margin of the basin, over 700 m of siliciclastic turbidites of the Campdevànol Formation (Gich, 1969, 1972), linked to southward-prograding fluvio-deltaic and fan-delta systems of the lowermost part of the Bellmunt Formation (Gich, 1969, 1972), overlie the slope marl facies.

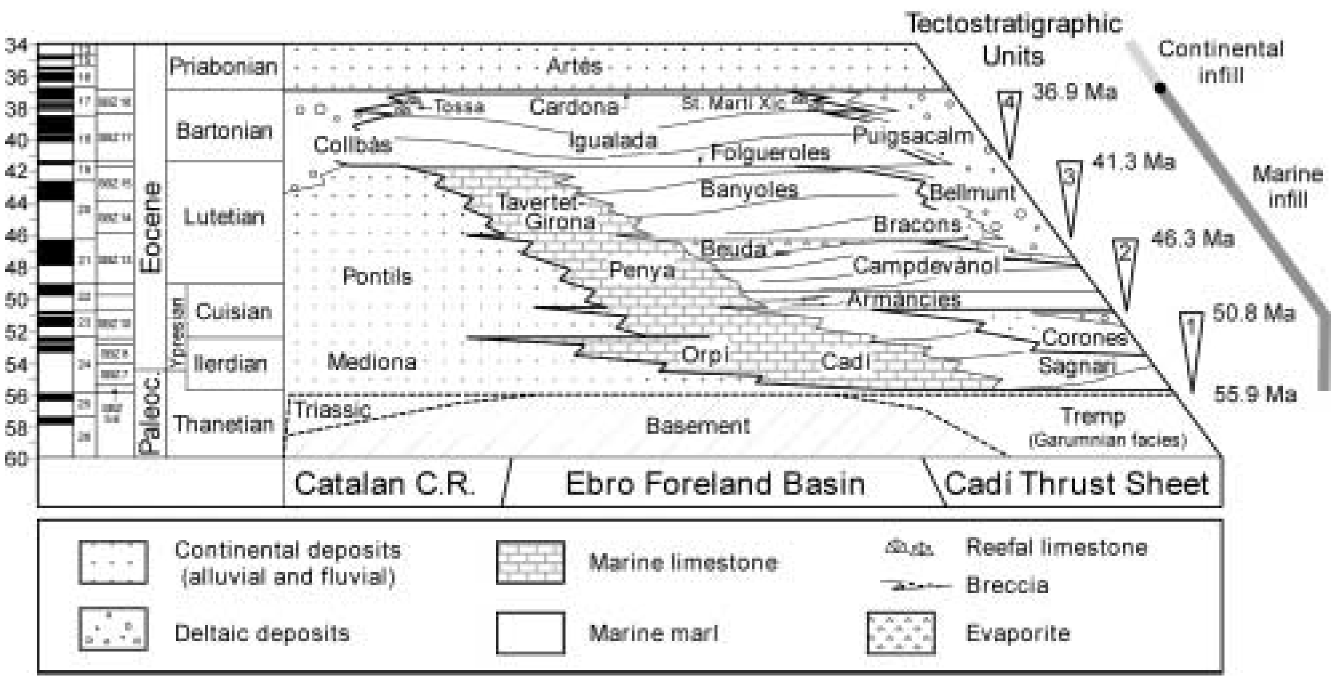

FIGURE 3 **Stratigraphic panel across the Ebro Foreland Basin between the Cadí Thrust Sheet and the Catalan Coastal Ranges, modified from Vergés et al. (1998). Biostratigraphic data combined with magnetostratigraphic information (Burbank et al. 1992a,b; Vergés and Burbank 1996; Serra-Kiel et al. 1994) define the chronostratigraphic framework of this area. Global polarity time scale from Cande and Kent (1995). Larger foraminifers biozones (SBZ) from Serra-Kiel et al. (1998a, b).**

The third tectostratigraphic unit is Early Lutetian to Late Lutetian in age. This third unit starts with a new transgression recorded at the northern basin margin by a glauconite-rich level. The mostly sandy, nearshore to deltaic facies association of the Cal Bernat Formation (Busquets, 1981) and Bracons Formation (Gich, 1969, 1972) unconformably overlie previous continental deposits of the lowermost part of the Bellmunt Formation (Gich, 1969, 1972). These sandy facies are overlain by a thick, southward-prograding, terrigenous fluvio-deltaic wedge (Mató et al., 1994), comprising the conglomerate, sandstone and marl of the Bellmunt Formation (Gich, 1969, 1972) and Banyoles Marl Formation (Almela and Ríos, 1943). The thickness of this fluvio-deltaic wedge is around 700 m to the east of the studied area. To the south, these fluvio-deltaic clastics overlap a southern carbonate platform of the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972).

The fourth tectostratigraphic unit is Bartonian in age. This fourth unit is characterised by the absence of well-developed shelf carbonate deposits. This is because of the simultaneous progradation from both basin margins: the Catalan Coastal Ranges to the south (e.g. Montserrat fan-delta deposits, López-Blanco et al., 2001) and the Pyrenean margin to the north with the Milany Formation (Gich, 1969, 1972) and the Puigsacalm Sandstone Formation (Gich, 1969, 1972). These formations are composed of deltaic facies (sandstone and marl), north-grading basinward to prodeltaic-offshore marl of the Igualada Formation (Ferrer, 1971). The overall offlapping trends of these terrigenous wedges are locally punctuated by minor onlapping transgressive pulses leading to the local and temporary development of carbonate facies and reef build-ups, such as the Tossa Formation (Ferrer, 1971) and Sant Martí Xic Limestone Formation (Reguant, 1967). These mixed terrigenous-carbonate wedges overlie the transgressive, glauconite-rich, shelf sandstone complex of the Folgueroles Sandstone Formation (Reguant, 1967) and are capped in turn by an evaporitic plug of the Cardona Formation (Riba, 1967, 1975) that fills up the marine basin.

THE MARINE LUTETIAN CYCLES

The marine sedimentation in the southern part of the South Pyrenean Foreland Basin was not significant until the Lutetian, although there is a prior intercalation of Ilerdian marine deposits in the Orpí Formation (Ferrer, 1971). The Lutetian marine units have an important lateral extension in the western part of the southern margin of the basin, stretching from Vic in the west to the Mediterranean sea in the east (Fig. 4). These materials (Fig. 5) change transitionally towards the south to the continental deposits of the Romagats Formation (Colombo,1980).

The Lutetian marine units and their continental equivalents lie on alluvial-fluvial mudstone, sandstone and conglomerate that were attributed by Pallí (1972) to the Pontils Formation (Ferrer, 1971), and were later redefined by Colombo (1980) as the Vilanova de Sau and Romagats Formations in the Vic area. Both authors have assigned these deposits a Cuisian-Early Lutetian age.

Four sedimentary cycles have been differentiated within the Lutetian marine units. The description of the stratigraphic features of these cycles, and their outcrops, focuses on two separate geographical areas, each with different sedimentological characteristics. These are the Amer-Vic area in the western part of the basin, and the Empordà area in the east.

The Amer-Vic Area

The Amer-Vic area (Fig. 4) comprises autochthonous deposits located to the south of the South Pyrenean Frontal Thrust (Muñoz et al., 1986), extending from the Vic area in the west to the Camós-Celrà extensional fault (Saula et al., 1994b) in the east.

The four sedimentary cycles differentiated within the Lutetian marine units have been ordered by decreasing age. In this zone the four cycles can be clearly distinguished and, furthermore, their north-south evolution can be analysed.

First Sedimentary Cycle

Lithostratigraphically, the materials that make up the 1st Cycle are equivalent to the lower part of the Tavertet Limestone Formation (Reguant, 1967) and the lower part of the Girona Limestone Formation (Pallí, 1972).

Two parts can clearly be distinguished in this cycle: a lower siliciclastic interval, and an upper, much thicker, mostly carbonate interval. The cycle records a general deepening in the depositional environment.

The maximum thickness of the cycle is about 55 m (Fig. 6). The lower interval, up to 5 m thick, is made up of millimetric and centimetric sequences of siltstone, claystone, and variegated sandstone. There are also millimetric intercalations of alabastrine and columnar gypsum with some pelitic breccia intraclasts. In the upper interval, decimetric beds of sandstone are interbedded. These sandstone beds have medium-scale, planar cross laminations reflecting a north western paleocurrent. A ferruginous crust caps on top the last bed in this interval. This succession marks the base of the 1st Cycle, representing the transition between the underlying continental deposits of the Vilanova de Sau/Pontils Formations and the marine deposits of the base of the Lutetian.

The upper part of the cycle is 50 m thick and has a predominance of centimetric and decimetric sandstone beds. These sandstone beds display medium scale, low angle, planar cross lamination. Decimetric beds of bioturbated sandy limestone, oolitic and porcellaneous foraminifer grainstone and packstone occur interbedded with these sandstone beds. In the last 15 m, the sandstone beds become scarcer, being replaced predominantly by packstone and, to a lesser extent, by wackestone of miliolids and *Alveolina*, and monospecific packstone-wackestone beds of *Nummulites*. This vertical transition indicates a deepening trend in the depositional environment, changing from littoral facies at the base to inner shelf facies at the top of the cycle (Fig. 6). The upper boundary is marked by a sharp contact with the bottom facies of the overlying 2nd Cycle.

To the south (Fig. 5), the deposits of this 1st Cycle change transitionally into the continental deposits of the Romagats Formation (Colombo, 1980). To the north, the siliciclastic intercalations become less numerous, indicating that the detritic supply came exclusively from the south. The thickness of the deposits of the 1st Cycle is constant along this section.

Second Sedimentary Cycle

Lithostratigraphically, the materials of the 2nd Cycle are equivalent to the upper part of the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972).

This 2nd Cycle is mainly composed of carbonate beds (Fig. 6), and indicates a general trend towards a deepening of the depositional environment. The thickness of the deposits of the 2nd Cycle varies between 100 and 125 m.

This cycle is made up of metric thick sandstone beds, sandy limestone, and packstone and grainstone beds. The siliciclastic components are more important in the lower part of the unit and in the southern areas. There is widespread evidence of intense bioturbation. The sandstone beds include coarsening-upwards sequences, some of them with high-angle planar cross stratification, indicating a northward progradation. These facies alternate with fining-upwards sequences in which the upper part is richer in bioclasts. The packstone and grainstone, generally nodular in appearance, are characterised by an abundance of *Nummulites* and, to a lesser extent, of oysters, equinids and, in their upper part, of *Assilina*. The limestone beds have been interpreted as a protected shelf-inner shelf transitional facies, with abundant monospecific banks of *Nummulites*. Some of these banks, when reworked, became littoral bars made up of grainstone of *Nummulites* with imbrication and interpenetration marks between the specimens due to compaction.

The upper boundary of the 2nd Cycle is marked by an intensely bioturbated ferruginous biocalcarenitic level which locally shows a high glauconite content. This bed

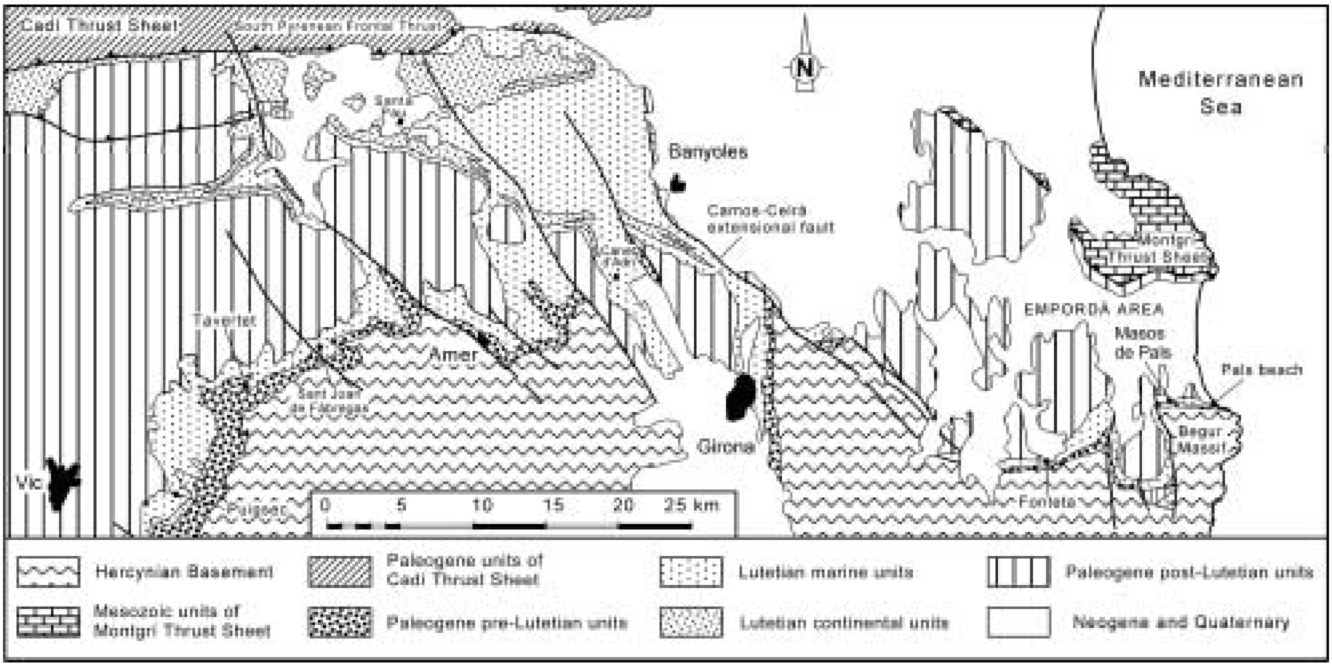

FIGURE 4 **Geological sketch of the Amer-Vic and Empordà area.**

has been interpreted as a condensation level and represents the transgressive surface which developped between the 2nd and 3rd Cycles.

To the south, the facies of the 2nd Cycle change transitionally to the alluvial continental facies of the Romagats Formation (Colombo 1980), with a progressive reduction of the marine deposits (Fig. 5). The transition is represented by a transgreded alluvial fan, a transgressive beach sequence and bars of reworked *Nummulites.* The transgressive system of this 2nd Cycle onlaps the previous cycle (Fig. 5).

Third Sedimentary Cycle

Lithostratigraphically, the materials of the 3rd Cycle are equivalent to the lower and middle parts of the Coll de Malla Marl Formation (Clavell et al., 1970), the Banyoles Marl Formation (Almela and Ríos, 1943), and to the lower part of the Bracons Formation (Gich, 1969, 1972) and Bellmunt Formation (Gich, 1969, 1972).

In the southern and central parts of the studied area of the basin, the upper boundary of this 3rd Cycle is marked by a sharp contact with the transitional facies of the 4th Cycle, while in the northern part this boundary can not be distinguished since the contact is located within the continental facies.

Two parts with different lithological and sedimentological features are distinguished in this cycle. The lower part up, to 50 m thick and mainly marly, is interpreted as a transgressive system. The upper part, with a thickness increasing to the north to up to 1000 m, composed of azoic marl grading upwards to interbedded azoic marl, sandstone and conglomerate. This upper part is interpreted as the regressive system assemblage (Figs. 5 and 6).

The bottom of the transgressive system consists of the above described transgressive surface, on top of the previous cycle. This surface contains, besides the glauconite, bioturbation and abundant fossils. The transgressive system continues with metric-scale levels of blue marl with fine centimetric and decimetric intercalations of blue marly limestone layers. These latter display bioturbation by *Thallassinoides* and *Ophiomorpha*, especially intense at the base, and are very rich in fossil content. The facies of this transgressive system are interpreted as siliciclastic shelf deposits. The upper boundary of this transgressive system is marked by a sharp contact with the azoic marl of the overlying regressive system.

To the south (Fig. 5), the transgressive system deposits clearly onlap the top of the 2nd Cycle materials, and pass laterally into the continental deposits of the Romagats Formation (Colombo, 1980). The extent of the landward ingression of these marine deposits is more important than that of the corresponding phase of the previous cycle. The transgressive system continues northward with very similar characteristics and thickness (Fig. 5).

The regressive system is composed of three intervals, each with different lithological features: the lowest inter-

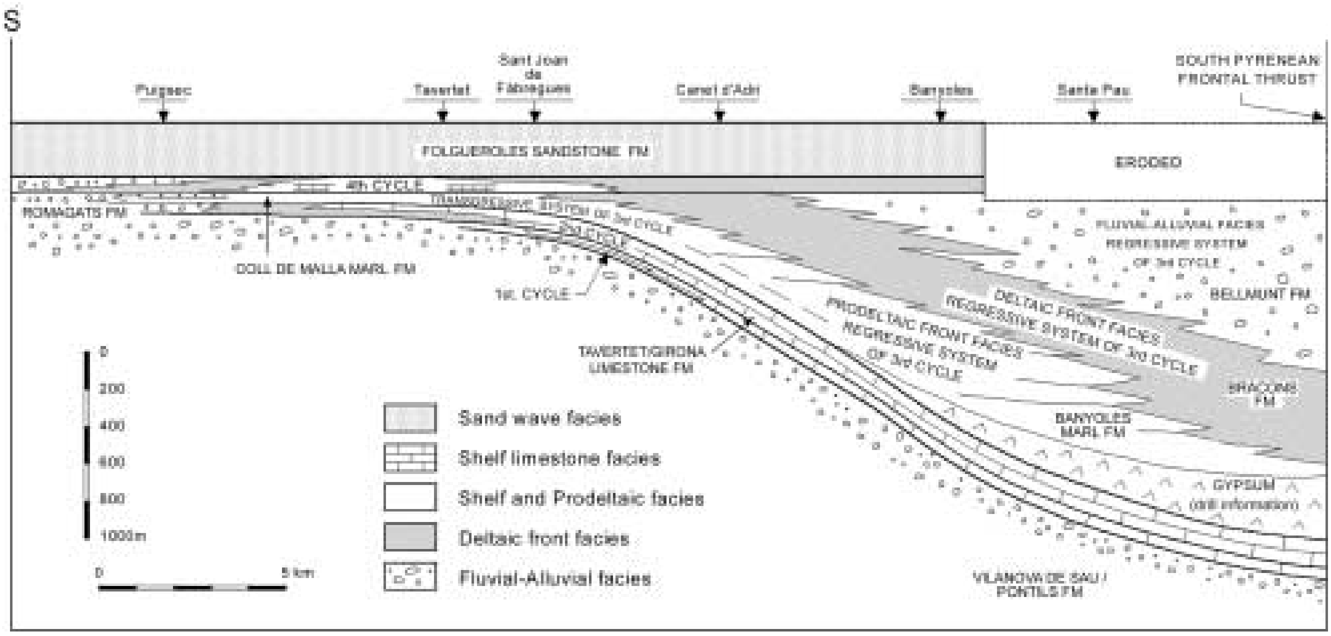

FIGURE 5 **Stratigraphical sketch across a north-south section showing the evolution of the facies during the Lutetian cycles.**

val comprises prodeltaic facies, the intermediate one consists of deltaic-front facies, and the upper one is composed of alluvial facies. The whole regressive system progrades southward (Figs. 5 and 6).

The lower interval is made up of claystone and dark blue marl, finely stratified and practically azoic. These deposits have been interpreted as distal shelf and prodelta deposits. The transition from these facies to those of the intermediate section occurs upwards and laterally to the north. The thickness of this group of facies increases to the north, reaching more than 350 m in the area of Banyoles (Figs. 4 and 5), while in the south, the thickness is only 20 m. Lithostratigraphically, this interval corresponds to the upper part of the Coll de Malla Marl (Clavell et al., 1970) and the Banyoles Marl Formations (Almela and Ríos, 1943).

The middle interval shows a very varied lithology, with centimetric- to metric-thick layers of marl, marly limestone, sandstone, and some conglomerate beds (Fig. 6). These facies correspond to very varied depositional environments, from deltaic to mixed shelf environments. In general, the mixed shelf facies occur more frequently in the northernmost areas and in the lower part of the successions. Within these facies, there is an abundance of levels of bioclastic sandstone which are of metric thickness and show medium- to high-scale cross-bedding. Some of these levels are composed entirely of *Nummulites*, with imbricated specimens, and have been interpreted as bars of reworked *Nummulites*. The deltaic facies make up the thickest part of the series. Within these facies there are many metric thickening- and coarsening-upwards sequences, which grade from lutite, at the base, to coarse sandstone or conglomerate at the top, and with an abundance of small- and medium-scale tractive laminations and internal erosion scars. These deposits have been interpreted as the product of stream-mouth-bar progradation and the repeated infilling of interdistributary areas. In the northernmost areas, the thickness of the sequence reaches some 450 m, a thickness which reduces towards the southern part of the basin, in part due to the transition to the marl of the lower successions (Fig. 5). Lithostratigraphically, the intermediate successions correspond to the lower part of the Bracons Formation (Gich, 1969, 1972).

The upper interval is made up of alluvial sandstone, red mudstone and conglomerate (Fig. 6). The coarse-grain content is greater to the north and in the upper successions. Overall, facies relationships indicate north-south progradation. To the south, the continental facies change gradually into the facies of the middle interval, fining down and disappearing completely some kilometres north of the southern margin of the basin (Fig. 5). The maximum thickness of this upper interval is difficult to determine, given that in the northern areas the boundary with the succeeding cycle is located within the continental facies. The lithostratigraphic unit which corresponds to this upper interval is that of the Bellmunt Formation (Gich 1969, 1972), although in the northernmost areas this unit has also been attributed to the continental equivalents of the subsequent sedimentary cycle.

On the southern margin (Fig. 5), the regressive system is represented by bars of fine- to medium-grain sand which alternate with azoic marl. Thickness is about 20 m.

During the deposition of the regressive system assemblage of the 3rd Cycle the basin was asymmetric. This is reflected in the decrease in thickness of the deposits from a minimum of 1200 m in the northernmost outcrops to only tens of metres in the southern margin (Fig. 5). This asymmetry is interpreted as the result of the northward flexing of the lithosphere due to the emplacement of overthrust units on the hinterland basin. At the same time, the presence of these overlapping units brought about an important increase in the quantity of deposits which originated on the northern margin of the basin, producing the north-south progradation of the regressive system and the consequent predominance of continental deposits in the northern part of the basin (Fig. 5).

Fourth Sedimentary Cycle

Lithostratigraphically, the materials of this 4rd Cycle are equivalent to the upper parts of the Coll de Malla Marl Formation (Clavell et al., 1970), the Bracons Formation (Gich, 1969, 1972) and Bellmunt Formation (Gich, 1969, 1972). The upper boundary of the cycle is marked by a sharp contact with the sandstone beds of the 1st Bartonian Cycle.

The lithology of the materials which make up the base of this cycle is predominantly siliciclastic in the northern part of the basin. To the south, the lower part is carbonate and the upper is siliciclastic, representing a transgressive and a regressive system, respectively. The trend marked by the whole cycle is one of a progressive shallowing of the depositional environment, and the unit as a whole has a thickness of some 80 m (Figs. 5 and 6).

In the southern part of the basin, the transgressive system begins with decimetric beds of marl and marly limestone with *Nummulites*. It is followed by a metric-thick level of packstone and porcellaneous foraminifer grainstone (miliolids, *Idalina*, *Orbitolites*, *Alveolina*). These facies, which represent the transgressive system assemblage, are interpreted as having been laid down in inner and protected shelf environments.

In the northern part of the basin, the materials at the base of the cycle show a much greater siliciclastic influence, seen in the transition from the marly deposits at the base to the sandy limestone and sandstone. The calcareous sandstone beds which show evidence of bioturbation and contain abundant porcellaneous foraminifers, have been interpreted as being facies of the near shore-inner shelf transition, laid down in a transgressive context.

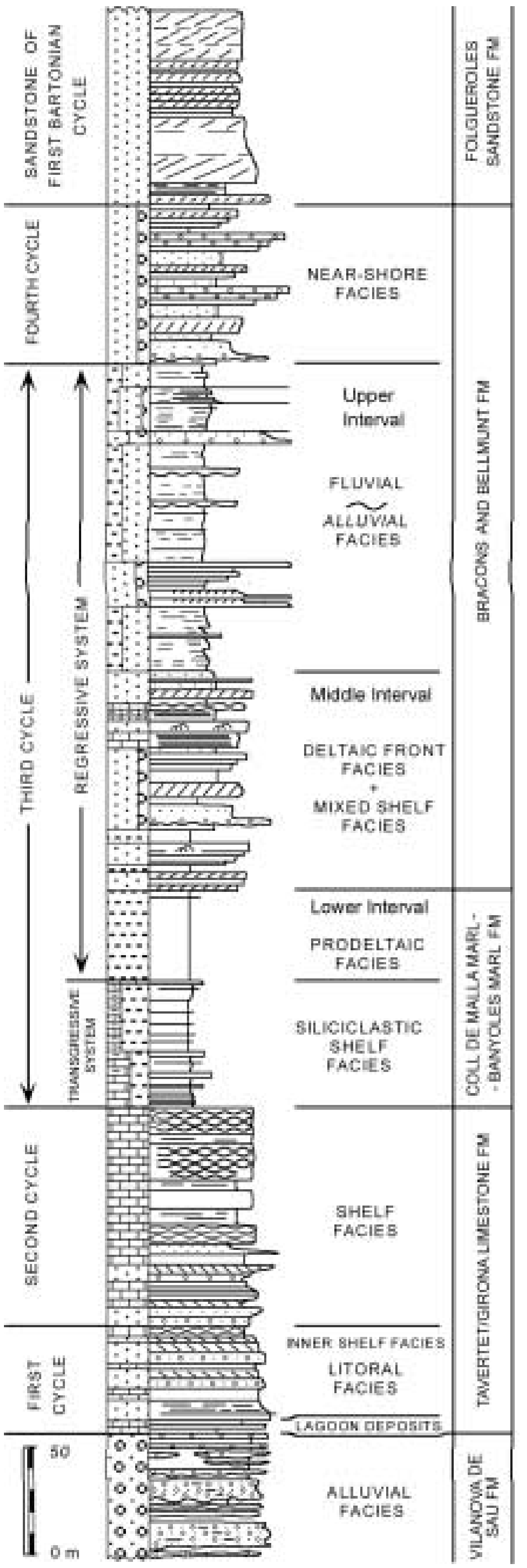

FIGURE 6 **Synthetic stratigraphical section of the Lutetian units from the Amer-Vic area.**

The facies of the regressive system, in the southern part of the basin, are characterised by decimetric beds of marl, marly limestone, limestone and sandstone which are particularly frequent in the upper part of the cycle. In the northern part of the basin, the sandstone beds dominate the regressive system, reaching thicknesses of metric order. Many ripples, together with medium-scale planar and trough cross-laminations, occur within these facies which have been interpreted as shoreface facies. In the northernmost areas, where the 4th Cycle deposits overlie the continental facies of the 3rd Cycle, the sedimentation is completely siliciclastic and is composed of decimetric beds of sandstone with occasional intercalations of conglomerate.

In the south of the basin, the marine deposits of this cycle pass laterally to the continental deposits of the Romagats Formation (Colombo, 1980), forming the deepest interfingered wedge in this formation (Fig. 5). This fact, together with the fact that to the north the marine deposits of this cycle overlie continental deposits to a distance of some 5 km, have led this cycle to be interpreted as the most transgressive of the Lutetian. In the northern part of the basin, the deposits of this cycle have been eroded over a large area, making it impossible to observe the marine-continental transition. However, it is probable that this transition was to the continental facies of the Bellmunt Formation (Gich, 1969, 1972) (Fig. 5).

Lithostratigraphically, the materials of the 4th Cycle correspond to the upper part of the Bracons Formation (Gich, 1969, 1972).

The Empordà Area

This area is further to the east than the Amer-Vic area and comprises the autochthonous deposits located to the south of the South Pyrenean Frontal Thrust between the Camós-Celrà extensional fault in the west and the Mediterranean sea in the east (Fig. 4). In the western part of this area, the outcrops are heavily affected by tectonic activity and the majority show only incomplete sections. For this reason, only the eastern outcrops are described here (Fig. 7).

The 1st and 2nd Cycles are described together since the boundary between them is not clear, probably because the contact is a paraconformity between similar marine deposits laid down in very near-shore-type environments (Fig. 7a, c).

First and Second Sedimentary Cycles

Taking these two cycles as a whole, two parts with different lithological features can be distinguished (Fig. 7a, c, d). The lower part is siliciclastic, while the upper one, much thicker, is carbonate. The general trend exhibited is one of facies laid down in a deepening depositional environment. The thickness of the unit as a whole increases from 25 m in the east to 75 m in the west.

The lower part (Fig. 7d), 5 m thick, is similar in character to the lower part of the 1st Cycle in the Amer-Vic area described above. It is composed of an alternating sequence of millimetric-centimetric beds of marl, siltstone, sandstone, and microconglomerate, laid down as lagoon and barrier-island facies, with an important siliciclastic supply from the continent. This section represents the base of the 1st Cycle and marks the transition from the continental facies of the Pontils Formation (Ferrer, 1971) to the Lutetian marine facies. In the easternmost part of the area (Fig. 7a), the thickness of the Pontils Formation is minimal or non-existent. The facies at the base of the 1st Cycle being is in discordant contact with the underlying Paleozoic. In these eastern parts of the Empordà area, the facies of the lower part are composed of a 2 to 3 m thick beds of coarse conglomerate and bioclast with abundant *Nummulites*.

The conglomerate beds are composed of Paleozoic pebbles supplied by the Begur Massif, located to the south. They are imbricated landwards as the result of their reworking by waves during the early stages of the marine transgression which marks the limit of the 1st Lutetian Cycle. In this area, the limestone layers of the 1st and 2nd Cycles change laterally to bioclastic ochre sandstone, representing beach facies, and then to continental conglomerate. The transition occur very rapidly in only about 20 m indicating a very low level of supply from the uplifted area on the southern margin of the basin.

The upper part of this unit is composed of carbonate deposits made up of packstone and grainstone, with an abundance of bioclasts and larger foraminifers. In the lower part of this section there are scattered intercalations of sandy limestone which indicate some terrigenous supply from the continent.

The upper boundary between the 1st and 2nd Cycles is marked by an intensely bioturbated biocalcarenitic ferruginous level with a high glauconite and fossil content. This level is similar to that described at the top of the 2nd Cycle in the Amer-Vic area and has been interpreted as a condensation level representing the transgressive surface between the 2nd and 3rd Cycles.

In the eastermost part of the Empordà area, the 3rd Lutetian cycle is missing (Fig. 7a) and the top of the 2nd Lutetian cycle shows a paleokarst that affects the limestone beds of the 1st and 2nd Lutetian cycles. This paleokarst is filled up with laminated bioclastic sandstone with *Nummulites* from the 4th Lutetian cycle.

Lithostratigraphically, the materials of the 1st and 2nd Cycles correspond to the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972).

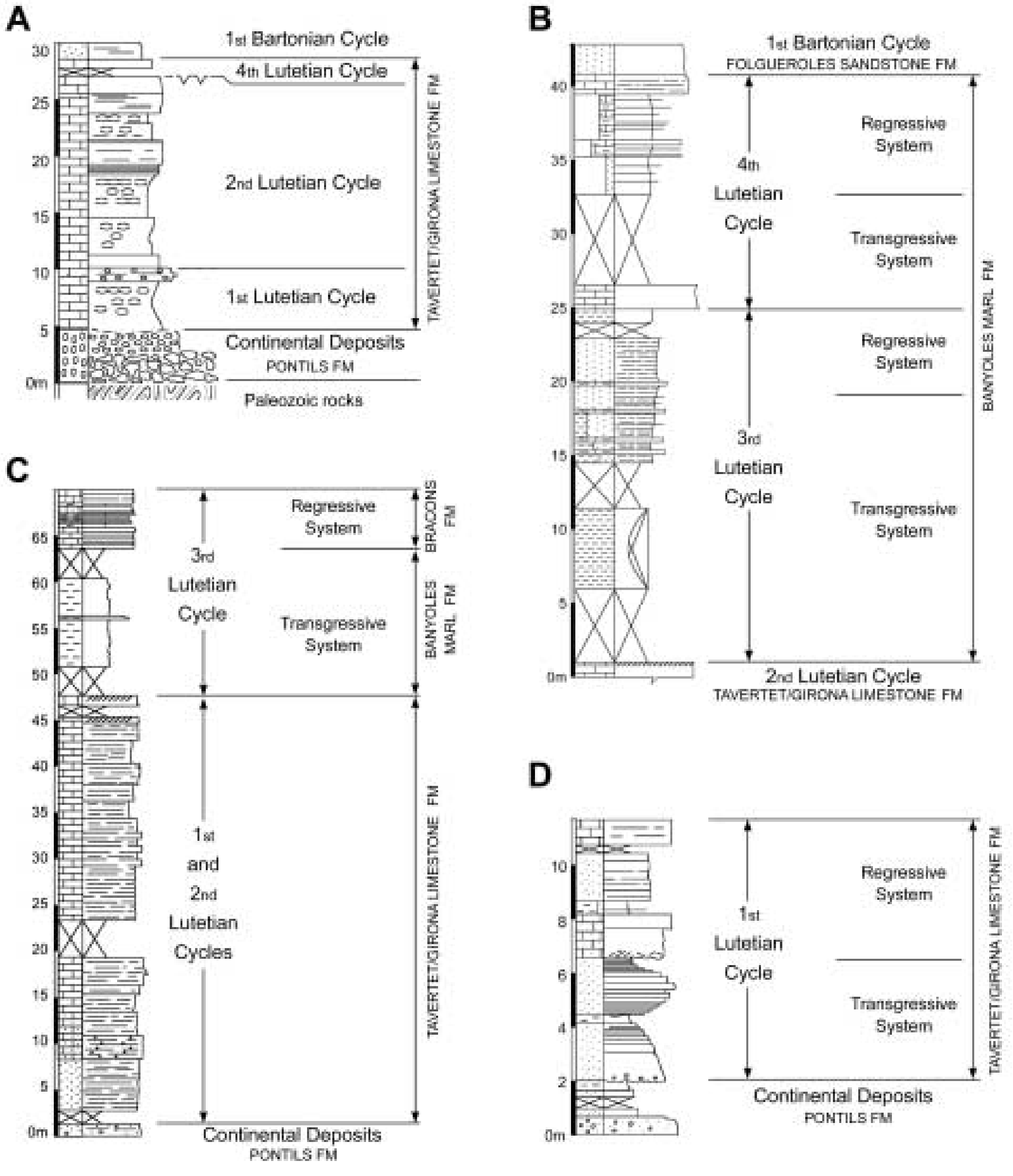

FIGURE 7 **Selected stratigraphic sections showing the Lutetian sedimentary cycles in the Empordà area (Fig. 6). A: Pals beach section, B: Massos de Pals section, C: Fonteta (Can Ametller) section, D: Fonteta (Can Barrabàs) section.**

Third Sedimentary Cycle

Two parts with different sedimentological features can be distinguished within this cycle (Fig. 7b, c).

The lower part presents a predominantly marly lithology, with a thickness that ranges from 10 m in the easternmost outcrops to 25 m in the west. This lower part, which represents the transgressive system, is very similar to the corresponding section in the Amer-Vic area. It is made up of metric beds of blue marl with fine, centimetric and decimetric intercalations of intensely bioturbated blue marly limestone with a high fossil content.

The upper part of the cycle is made up of marl with intercalations of limestone and fine sandstone. The thickness is about 10 m and this part is interpreted as corresponding to the regressive system assemblage of the cycle (Fig. 7c). In the central parts of the Empordà area, this upper part of the cycle is divided into a lower carbonate interval of packstone and bioclastic grainstone, and an upper section of limestone with the same texture and with intercalations of marl with oysters and bioclastic sandstone. The whole cycle has been interpreted as a group of protected shelf facies with a siliciclastic supply from the continent. In the easternmost outcrops of this cycle the regressive system comprises a sequence of siltstone with well-marked millimetric stratification, and with intercalations, in the lower part, of centimetric layers of marly limestone, fine sandstone and an abundance of miliolids, followed by azoic grey-ochre marl. This group of facies has been interpreted as protected-shelf grading to lagoon deposits.

Fourth Sedimentary Cycle

This cycle begins with a lower part (Fig. 7b), about 2 m thick, composed of grainstone-textured carbonate with a high porcellaneous foraminifer content (miliolids, *Idalina*, *Fabularia*, *Orbitolites*, *Alveolina*). These deposits, which form the transgressive system of the cycle, have been interpreted as protected-shelf-bar facies. The cycle continues with an upper part of siliciclastic materials, about 10 m thick, made up of marl with centimetric intercalations of bioclastic siltstone, marly limestone and occasional packstone with a high miliolid content and some gasteropods (*Velates*). This upper part corresponds to the regressive system.

The thickness of the deposits which make up this cycle increases in a westerly direction. The upper boundary is marked by a sharp contact with the arkosic sandstone of the 1st Bartonian Cycle (Fig. 7a, b). In the southern parts of the Empordà area, the base of the Bartonian is formed by a bioclastic calcarenitic layer rich in glauconite and iron.

The west-east reduction in the thickness of the Lutetian units, from the Amer-Vic area to the Empordà area, and the transition to shallower environments in the same direction, indicate a clear west-east reduction of basin subsidence and the proximity of the eastern margin of the South Pyrenean Foreland Basin. This fact, first presented by Pallí (1972), would indicate the presence of an uplifted massif in the area now occupied by the Mediterranean.

THE MARINE BARTONIAN CYCLES IN THE IGUALADA AREA

The Paleocene-Eocene deposits along the southeastern margin of the Ebro Foreland Basin in the Igualada area (Fig. 2) consist of 2100 m thick alternating marine and continental deposits. These deposits were divided from bottom to top, according to Rosell et al. (1966, 1973), Ferrer (1971, 1973), Anadón (1978a) and Anadón et al. (1983), into the following formal lithostratigraphic units:

-The Mediona level (Rosell et al., 1966), or Mediona Formation (Anadón et al., 1983), is composed of red mudstone layers with widespread paleosoils. It is up to 50 m thick and has been attributed to the Late Thanetian (Anadón, 1978a; Anadón et al., 1983).

-The Orpí Formation (Ferrer, 1971) is composed of limestone beds rich in miliolids, *Orbitolites* and *Alveolina.* It is up to 100 m thick and is attributed to the Early and Middle Ilerdian (Hottinger, 1960).

- The Pontils Formation (Ferrer, 1971), upgraded to the Pontils Group by Anadón (1978a), is composed of red mudstone, siltstone, sandstone and conglomerate, and limestone of lacustrine facies. It is up to 900 m thick and is attributed to the Late Ypresian-Early Bartonian age (Anadón and Feist, 1981).

-The Santa Maria Formation was defined and subdivided into the Collbàs, Igualada and Tossa Members by Ferrer (1971). This unit is composed of marine deposits, up to 1100 m thick, and is attributed to the Bartonian ("Biarritzian") and Early Priabonian (Ferrer, 1971). In this paper we consider the Santa Maria Formation as a group, and the Collbàs, Igualada and Tossa Members as formations, following Ferrer (1973), Rosell et al. (1973) and Anadón et al. (1983).

-The Collbàs Formation is composed of siltstone, sandstone and conglomerate in the lower part, marl in the middle part, and reefal limestone in the upper part. It is attributed to the "Biarritzian" (Ferrer, 1971, 1973) or Early Bartonian (SBZ 17) (Serra-Kiel et al., 1997).

-The Igualada Formation is composed of blue-grey marl. It is attributed to the Early and Late Bartonian (Serra-Kiel et al., 1997).

-The Tossa Formation is composed of reefal limestone in the type-section. In the other parts of the basin, this unit contains marl, sandstone and conglomerate of deltaic facies interbedded with reefal limestone. It is attributed to the Late Bartonian (Serra-Kiel et al., 1997; Romero, 1996, 2001; Romero et al., 1999).

-The Artés Formation (Ferrer, 1971) consists of red and grey claystone, sandstone and conglomerate and has been attributed to the Priabonian (Anadón et al., 1992).

At the top of the Tossa Formation and underlying the Artés Formation, Travé (1992) defined the Terminal Complex. This unit consists of dark marl, carbonate wackestone-packstone with miliolids, carbonate laminae constructed by bacterial mats alternating with gypsum laminae, and gypsum deposits (Òdena gypsum).

Stratigraphy of the Santa Maria Group

The geological sketch of the area (Fig. 8) and the lithostratigraphic correlation between 13 stratigraphic sections (Fig. 9), show the geometric relationships between the different units of the Santa Maria Group which, from bottom to top and, according to Travé et al. (1999), are:

Collbàs Formation

This unit is up to 600 m thick and outcrops all along the southern margin of the basin in the Igualada area (Fig. 9). It overlies the continental and lagoonal deposits of the Pontils Group (Anadón, 1978a; Anadón and Marzo, 1986). The type-section is that of the Ermita de Collbàs (3 in Figs. 8 and 9). Three parts have been differentiated in this formation which, from bottom to top, are: *Lower Collbàs, Middle Collbàs* and *Upper Collbàs.*

The *Lower Collbàs* comprises three intervals in the southwestern part of the basin (Santa Maria de Miralles and Ermita de Collbàs sections, 1 and 3 in Figs. 8 and 9). The first interval, up to 10 to 20 m thick, is composed of conglomerate and coarse sandstone. The second interval, from 63 to 85 m thick, consists of marl interbedded with carbonate siltstone and wackestone-packstone layers with widespread larger foraminifers (*Nummulites, Alveolina* and *Orbitolites*). Finally, the third interval, up to 10 m thick, consists of reefal limestone. In the eastern part of the basin (Els Mollons section, 13 in Figs. 8 and 9), the *Lower Collbàs* is composed of marl and packstone layers made up of larger foraminifers, mainly *Discocyclina*, and is up to 200 m thick. There are also small olistoliths of Triassic limestone up to 1 m thick.

The *Middle Collbàs*, in the western part of the basin (Santa Maria de Miralles and Ermita de Collbàs sections, 1 and 3 in Figs. 8 and 9), is composed of 120 m of marl with larger foraminifers (*Nummulites, Assilina, Operculina* and *Discocyclina*) and ahermatypical corals such as *Cyclolitopsis patera* (D'ACHIARDI, 1867). In the eastern part of the basin (Els Mollons section, 13 in Figs. 8 and 9), a stratiform olistolith of Triassic limestone about 1800 m long, 1500 m wide and up to 40 m thick is interbedded between the marl (Anadón, 1978b).

The *Upper Collbàs* in the western part of the basin (Santa Maria de Miralles and Ermita de Collbàs sections, 1 and 3 in Figs. 8 and 9), is composed of reefal limestone interbedded with marl with abundant larger foraminifers. In the eastern part of the basin (Els Mollons section, 13 in Figs. 8 and 9) the *Upper Collbàs* is made up of marl with nummulitic packstone layers up to 15 m thick. Towards the centre of the basin (La Pobla de Claramunt-Puig Aguilera section, 7 in Figs. 8 and 9), these deposits shade into marl. The thickness of the *Upper Collbàs* can reach up to 120 m.

Castellolí Deltaic Complex

This unit is up to 100 m thick and is made up of conglomerate, sandstone, and azoic marl, organised in coarsening and thickening-upwards sequences with trough cross-bedding structures in the upper coarser facies (Riera de Castellolí, Castellolí-Can Tardà, Brucs E, and Els Mollons sections, 8, 10, 12 and 13 in Figs. 8 and 9). On the eastern margin, the deposits of the *Castellolí Deltaic Complex*, named Turó d´en Tort Conglomerate by Anadón (1978b), up to 300 m thick, are composed of conglomerate with interbedded sandstone developing a coarsening-upward sequence at the bottom and very coarse polygenic conglomerate at the top. Overlying the Turó d'en Tort Conglomerate, there are 30 m of massive sandstone, or marl and sandstone with red matrix, named Roca Cagadera Sandstone (Anadón and Marzo, 1986), which exhibit thickening- and coarsening-upward sequences. All these *Castellolí Deltaic Complex* deposits shade basinwards into azoic blue marl with tempestite beds (Fig. 9).

Castellolí Marl and Limestone

This unit developed on top of the *Castellolí Deltaic Complex* (La Pobla de Claramunt-Puig Aguilera, Riera de Castellolí, Castellolí-Can Tardà, Brucs W , Brucs E, and Els Mollons sections, 7, 8, 10, 11, 12 and 13 in Figs. 8 and 9). It is composed, from the bottom upwards, of limestone beds rich in oysters, packstone and grainstone made up of larger foraminifers, and locally, coral and coralline-algae build-ups interbedded with marl that contain larger foraminifers, bryozoa, and molluscs. The thickness of the *Castellolí Marl and Limestone* is up to 42 m in the Castellolí/Can Tardà section (10 in Figs. 8 and 9). Towards the centre of the basin, this unit shades into an interval, up to 220 m thick, of marl with marine fossils.

Upper Deltaic-Reefal Complex

This unit, up to 380 m thick, outcrops in the NE and in the SW sectors of the Igualada Area. It developed on top of the *Castellolí Marl and Limestone*, and below the *Terminal Complex* unit (Fig. 9). The features of this unit vary from area to area. On the western margin (La Tossa section, 2 in Figs. 8 and 9), this unit is characterised by 60 m of limestone with abundant corals and coralline-algae, interpreted as the aggradation of reefal facies with internal scars filled with grainstone, packstone and wackestone corresponding to small channelised areas within and between the reefal facies (Salas, 1979). On this west-

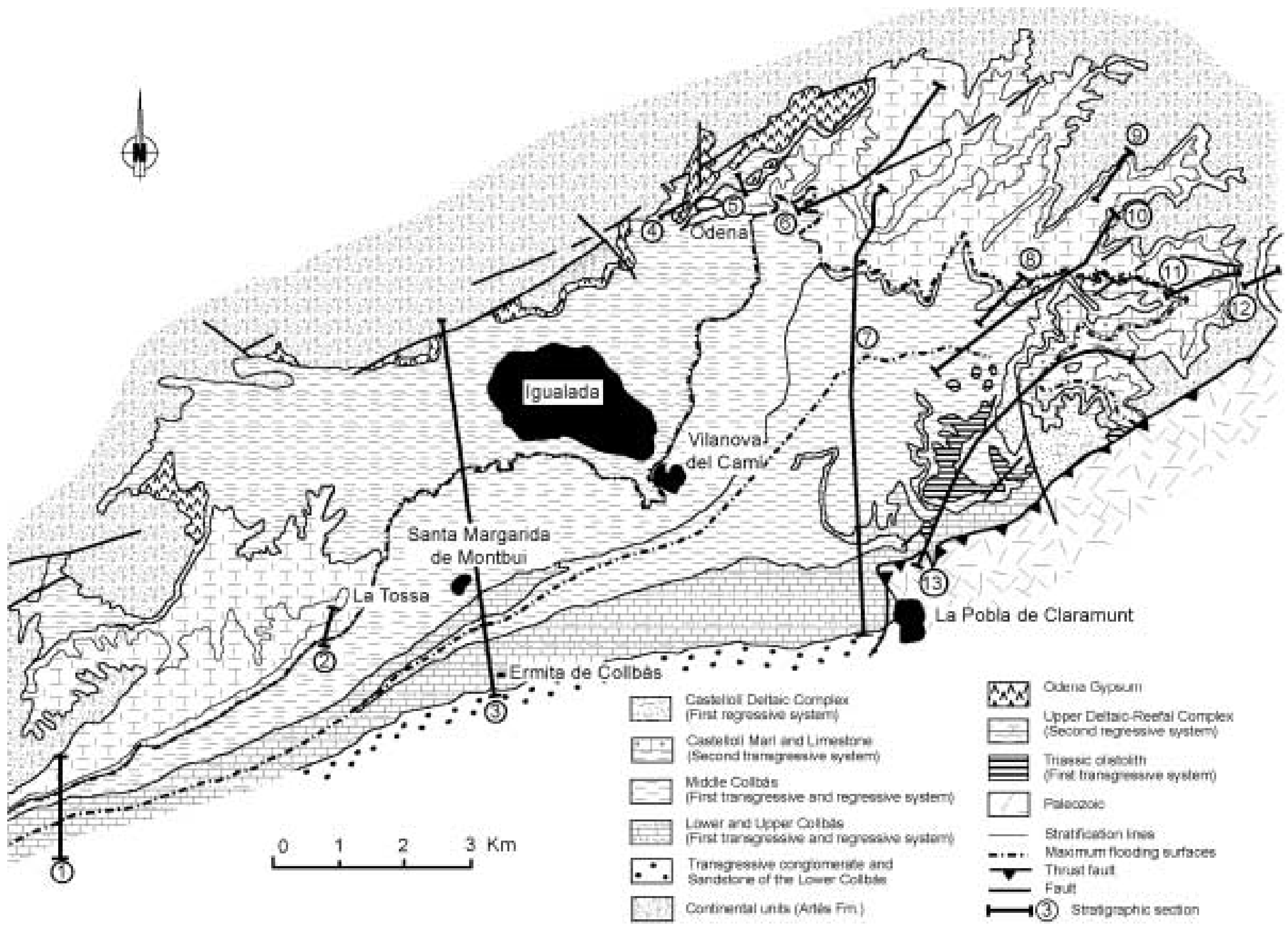

FIGURE 8 **Geological map of the Igualada area, with location of the stratigraphical sections. 1: Santa de Miralles section; 2: La Tossa section; 3: Ermita de Collbàs section; 4: Òdena section; 5: Can Valls section; 6: Can Aguilera de la Costa section; 7: La Pobla de Claramunt-Puig Aguilera section; 8: Riera de Castellolí section; 9: Torrent de Can Tardà section;10: Castellolí-Can Tardà section;11: Brucs W section; 12: Brucs E section; 13: Els Mollons section.**

ern margin the description of this unit corresponds to the Tossa Formation *sensu* Ferrer (1971, 1973) and Rosell et al. (1973). However, on the eastern margin of the basin, in the Can Aguilera de la Costa, La Pobla de Claramunt-Puig Aguilera, Torrent de Can Tardà, Castellolí-Can Tardà, Brucs W, Brucs E and Els Mollons sections (6, 7, 9, 10, 11, 12 and 13 in Figs. 8 and 9), this unit is up to 380 m thick and is formed of marl, sandstone and conglomerate, organised in coarsening-upward sequences interbedded with reefal limestone. Towards the east, corresponding to the areas closest to the basin margin, the *Upper Deltaic-Reefal Complex* passes to the red bed deposits of the Artés Formation (Ferrer, 1971), according to Anadón et al. (1985).

Characterisation of the sedimentary cycles

The analysis of the lithostratigraphic correlation between the 13 stratigraphic sections (Fig. 9) reveals two transgressive-regressive sedimentary cycles in the Bartonian marine deposits of the Igualada Basin, according to Travé et al. (1999).

First Sedimentary Cycle

This cycle extends from the beginning of the marine sedimentation up to the top of the Collbàs Formation (Fig. 9).

Transgressive System

The interval of coarse sandstone and conglomerate in the lower part of the Collbàs Formation has been interpreted as a transgressive barrier-island complex (Anadón and Marzo, 1986; Teixell and Serra-Kiel, 1988). On top of this transgressive base, carbonate siltstone, marl, wackestone-packstone horizons and patch reef facies of the *Lower Collbàs* are thought to have been deposited on the inner shelf (Teixell and Serra-Kiel, 1988). The most characteristic deposit of this inner shelf is a monospecific level of *Nummulites perforatus* (MONTFORT 1808) which is up to 3 m thick and several km long. This level has been interpreted as an *in situ* Nummulitic-bank parallel to the shoreline and located in the transitional facies from the shoreface to the inner shelf (Serra-Kiel and Reguant, 1984). More details on the paleocological features of this unit can be found in Romero and Caus (2000).

Most of the marl of the *Middle Collbàs* are characterised by the presence of abundant ahermatypical corals, i.e *Cyclolitopsis patera* (D'ACHIARDI, 1867), in the Santa Maria de Miralles section (1 in Figs. 8 and 9), and by larger foraminifers such as *Nummulites striatus* (BRUGUIÈRE 1792), *Assilina schwageri* (SILVESTRI 1928) and *Discocyclina* in the Ermita de Collbàs section (3 in Figs. 8 and 9). This association of fossils represents the middle-shelf facies. The uppermost part of the *Middle Collbàs* is composed of azoic marl.

In the La Pobla de Claramunt-Puig Aguilera section (7 in Figs. 8 and 9), the *Lower Collbàs* is thicker than in the other areas and is represented by grainstone and packstone with abundant echinoid fragments. In the Els Mollons section (13 in Figs. 8 and 9), the most frequent facies in the *Lower Collbàs* is a horizon of grainstone-packstone of *Discocyclina*, characteristic of outer-shelf facies (Caus and Serra-Kiel, 1984; Teixell and Serra-Kiel, 1988). Olistoliths of Triassic limestone (Anadón, 1978b) are interbedded in the *Lower* and *Middle Collbàs.* The location of these two sections (La Pobla de Claramunt-Puig Aguilera and Els Mollons sections, 7 and 13 in Figs. 8 and 9), close to the southeastern basin margin (Fig. 8), where a northward migrating thrust is present, suggests that a higher level of subsidence took place in this marginal area due to greater tectonic activity produced by the thrust, allowing the deposition of thick deep-facies successions and the emplacement of olistoliths within a transgressive context.

The maximum flood surface of this transgressive system is represented by the boundary between the marl with larger foraminifers and ahermatypical corals and the azoic marl of the uppermost part of the *Middle Collbàs*.

Regressive System

The first regressive system assemblage on the western margin (Santa Maria de Miralles and Ermita de Collbàs sections, 1 and 3 in Figs. 8 and 9) is formed by the upper part of the *Middle Collbàs* marl and the *Upper Collbàs*. The sequence consists of marl with some coral colonies with coralline-algae at the base, and followed upwards by reefal limestone with hermatypical corals and coralline-algae interbedded with marl. These reefal limestone layers are interpreted as patch reef facies formed in an inner-shelf environment, according to Teixell and Serra-Kiel (1988).

On the eastern margin of the basin (Riera de Castellolí, Castellolí-Can Tardà, Brucs E and Els Mollons sections, 8, 10, 12 and 13 in Figs. 8 and 9), the upper part of the regressive system is represented by the *Castellolí Deltaic Complex,* a succession of sandstone and conglomerate forming thickening- and coarsening-upwards sequences changing basinwards to azoic marl. These deltaic facies prograde to the northwest and are probably the distal equivalent of the Turó d´en Tort Conglomerate (Anadón, 1978b) and the Roca Cagadera Sandstone (Anadón and Marzo, 1986). They are interpreted as alluvial fan delta deposits (Anadón, 1978b).

Second Sedimentary Cycle

This cycle includes the *Castellolí Marl and Limestone* and the *Upper Deltaic-Reef Complex* (Figs. 7 and 8). During the deposition of this second cycle, sedimentation was much more homogeneous over the whole area, indicating

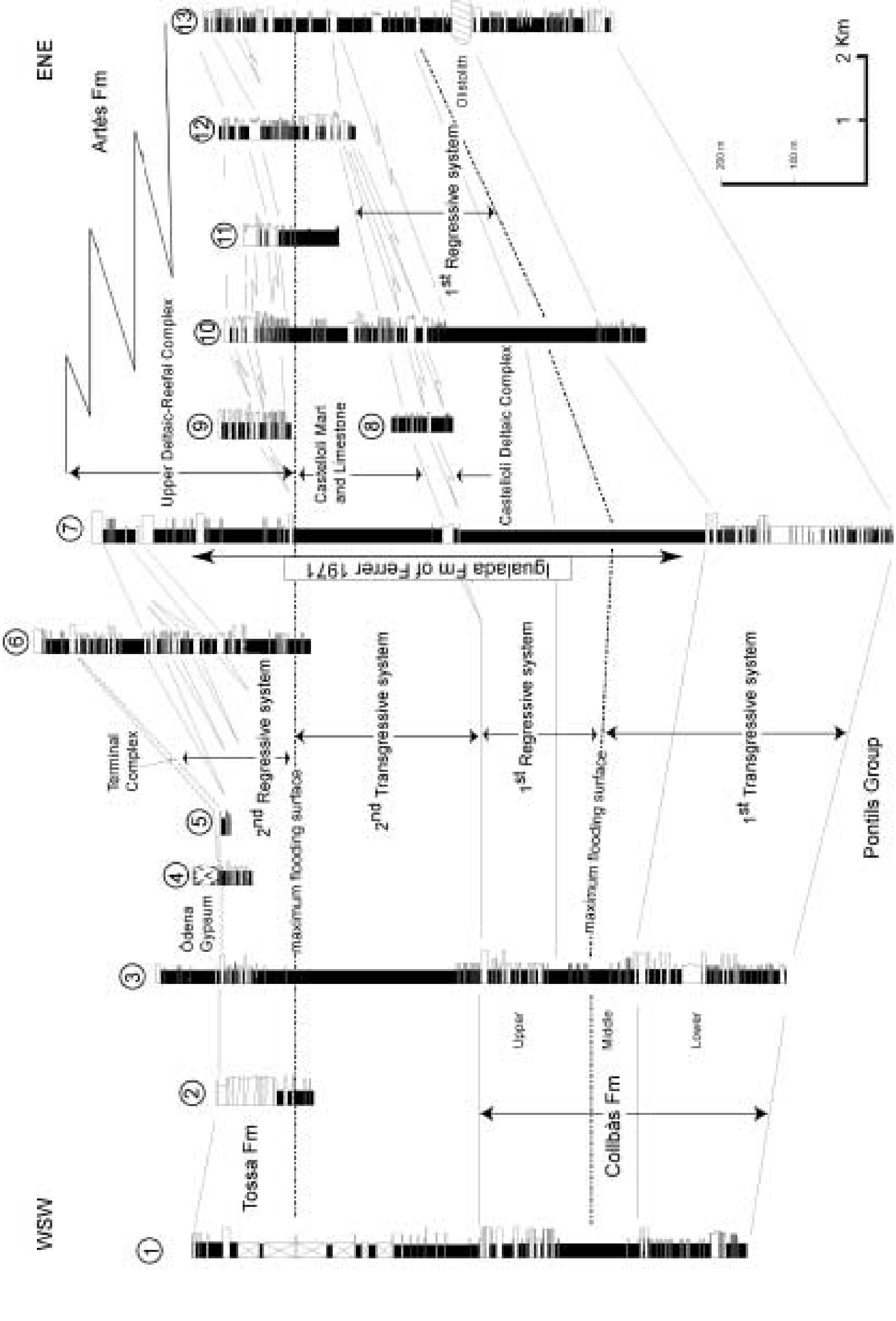

FIGURE 9 | **Correlation of the stratigraphical sections taking as datum the maximum flooding surface of the second Bartonian transgressive system. 1: Santa Maria de Miralles section; 2: La Tossa section; 3: Ermita de Collbàs section; 4: Òdena section; 5: Can Valls section; 6: Can Aguilera de la Costa section; 7: La Pobla de Claramunt-Puig Aguilera section; 8: Riera de Castellolí section; 9: Torrent de Can Tardà section;10: Castellolí-Can Tardà section; 11: Brucs W section; 12: Brucs E section; 13: Els Mollons section.The location of the sections is in Figure 8.**

the end of the large differences in subsidence rates and therefore a decrease in the tectonic activity in the basin margin. Details on the paleocological features of this unit can be found in Romero and Caus (2000) and Romero et al. (2002).

Transgressive System

The second transgressive system is represented by the *Castellolí Marl and Limestone* (La Pobla de Claramunt-Puig Aguilera, Riera de Castellolí, Castellolí-Can Tardà, Brucs W, Brucs E and Els Mollons sections, 7, 8, 10, 11, 12 and 13 in Figs. 8 and 9). Towards the eastern margin of the basin, this succession of deposits forms a marine wedge onlapping continental deposits. The base of the transgressive system in this sector of the basin consists of conglomerate and coralline-algae limestone layers. These limestone layers change vertically and laterally basinwards to marl with *Assilina*, *Operculina* and *Discocyclina*. In the more distal deeper part of the basin (Ermita de Collbàs and La Pobla de Claramunt-Puig Aguilera sections, 3 and 7 in Figs. 8 and 9), the succession is represented by marl containing abundant branching bryozoa. The absence of organisms requiring light for their development, such as larger foraminifers, indicates aphotic conditions during the sedimentation of this marl.

The maximum flood surface of this transgressive system is represented by a marker level that covers the basin, which is made up of abundant *Discocyclina*, *Asterocyclina*, *Assilina* and *Operculina*. This horizon, interpreted as a condensation level, has been observed in the Santa Maria de Miralles, La Tossa, Ermita de Collbàs, Can Aguilera de la Costa, La Pobla de Claramunt-Puig Aguilera, Castellolí-Can Tardà, Brucs W and Brucs E sections (1, 2, 3 6, 7, 10, 11 and 12 in Figs. 7 and 8) and has been chosen as the datum for the stratigraphic correlation (Fig. 9).

Regressive System

The second regressive system is represented by the *Upper Deltaic-Reefal Complex*. This complex is made up of detritic deltaic facies interbedded with reefal limestone facies (the Ermita de Collbàs, Òdena, Can Valls, Can Aguilera de la Costa, La Pobla de Claramunt-Puig Aguilera, Torrent de Can Tardà, Castellolí-Can Tardà, Brucs W, Brucs E, and Els Mollons sections, 3, 4, 5, 6, 7, 9, 10, 11, 12 and 13 in Figs 7 and 8) and reefal limestone beds in the La Tossa section (2 in Figs. 8 and 9). The *Upper Deltaic-Reef Complex* is aggradational and progrades towards the northwest, filling up the sedimentary trough. Towards the east, the *Upper Deltaic-Reefal Complex* changes laterally to the non-marine red bed deposits of the Artés Formation, according to Anadón et al. (1985).

The prograding coral reef and deltaic complexes of the 2nd Bartonian regressive system are overlain by the fluvial deposits of the Artés Formation (Ferrer, 1971), located along the margin of the Ebro Foreland Basin. In between, and bounded at the bottom and at the top by unconformities, a unit showing a great variability of facies is found. This unit, referred to as the “Calders Complex” in Vilaplana (1977) and “Post-reefal deposits filling up the Vic Basin” in Taberner (1983-84), was formally defined by Barnolas et al. (1983) and characterised by Travé (1992) as Terminal Complex.

The lithological features of this unit show a great variability of facies: siliciclastic continental and marine materials, carbonate shallow-shelf beds, stromatolitic facies and gypsum.

Despite the interest of these deposits because of their relationship with the evaporitic basin (Cardona Formation, Riba, 1967, 1975), and because they are the last marine episode in the Ebro Foreland Basin, the foraminifer content did not supply any data to the new biozones of larger foraminifers (Shallow Benthic Zones). Details on the lithostratigraphic and paleocological features of this unit can be found in Taberner (1983-84), Travé (1992), and Travé et al. (1994, 1996).

CONCLUDING REMARKS. CHRONOSTRATIGRAPHIC FRAMEWORK

The chronostratigraphic analysis of the Lutetian and Bartonian marine sedimentary cycles of the Ebro Foreland Basin was carried out using biostratigraphy based on larger foraminifers, planktic foraminifers and magnetostratigraphic data. The biozones of planktic foraminifers, the Shallow Benthic Zones of larger foraminifers (SBZ) and the magnetostratigraphic scale are those proposed by Berggren et al. (1995), Serra-Kiel et al. (1998a, b) and Cande and Kent (1995), respectively.

A magnetostratigraphic profile by Burbank et al. (1992a), from the Coll de Malla Marl Formation (Clavell et al., 1970) to the continental facies of the Artés Formation (Ferrer, 1971), shows five chrons, with eight magnetozones of normal polarity. This magnetostratigraphic section, together with old and new data on larger foraminifers and scattered data on planktic foraminifers, has allowed us to improve the precision of the correlation between the magnetostratigraphic scale and the biozones of larger foraminifers and planktic foraminifers for the interval comprised between the Middle Lutetian and the Late Bartonian. Only biostratigraphic data on larger foraminifers, mainly on *Nummulites* and *Alveolina*, were available for the Early Lutetian- Middle Lutetian interval.

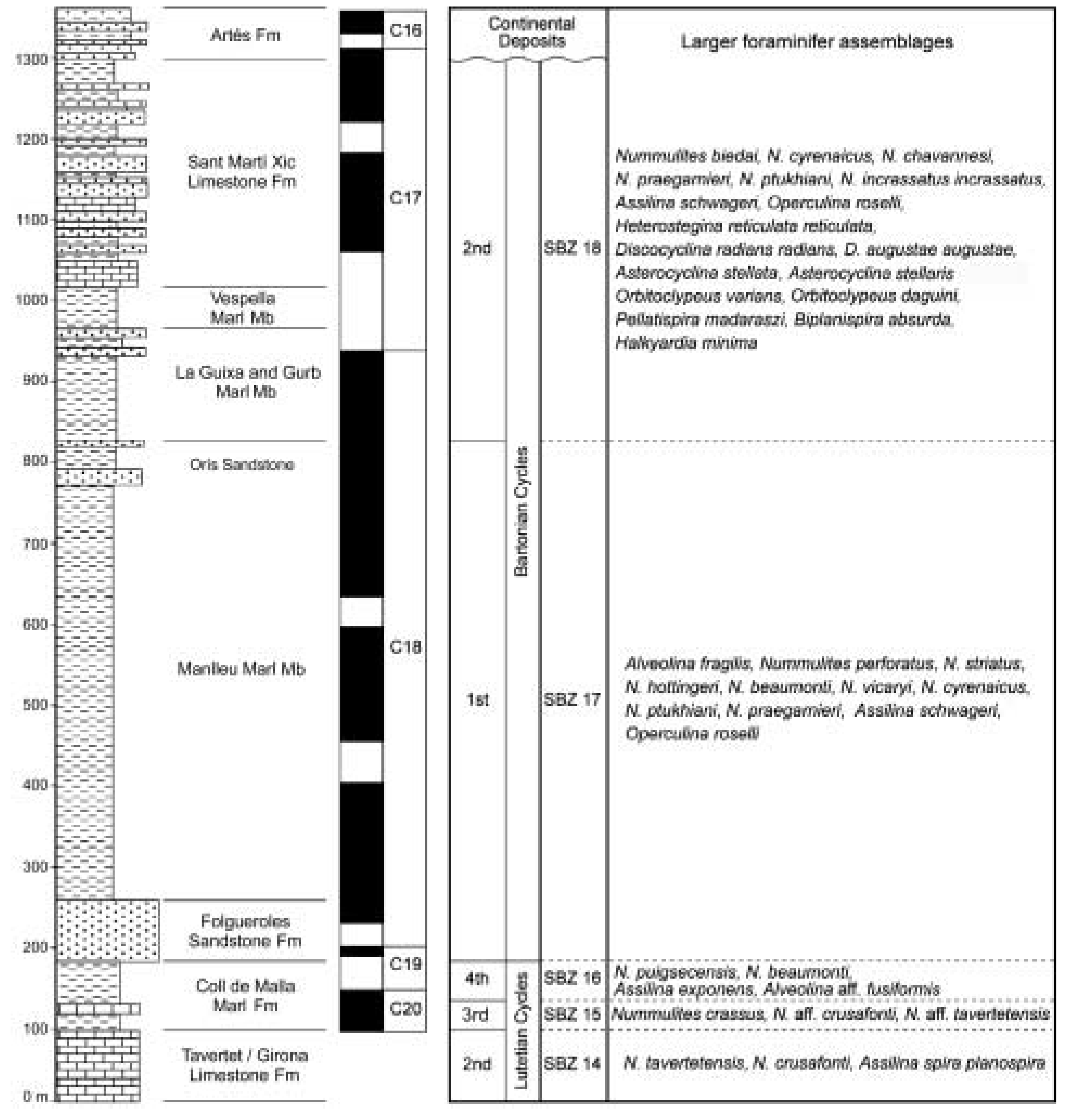

FIGURE 10 **Chronostratigraphic correlation between the magnetostratigraphic section of Burbank et al. (1992a), marine Lutetian and Bartonian sedimentary cycles, lithostratigraphic units, larger foraminifer biozones (SBZ) and larger foraminifer assemblages.**

Lutetian

First Sedimentary Cycle

The carbonate levels of the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972) of this 1st Lutetian cycle, according to Hottinger (1960), Serra-Kiel (1984) and Serra-Kiel et al. (1997), contain *Alveolina stipes* HOTTINGER 1960 and *Alveolina frumentiformis* SCHWAGER 1860, whereas the monospecific levels contain *Nummulites verneuili* D'ARCHIAC and HAIME 1853. Such an association allows the placing of this cycle in SBZ 13 (Early Lutetian).

Second Sedimentary Cycle

In the Tavertet/Girona Limestone Formation (Reguant, 1967; Pallí, 1972) the beds of the 2nd Lutetian cycle, according to Reguant (1967), Serra-Kiel (1984), and Serra-Kiel et al. (1997), contain different monospecific levels of the *N. tavertetensis* REGUANT and CLAVELL 1967 and *N. crusafonti* REGUANT and CLAVELL 1967. The siliciclastic and carbonate levels of the middle shelf contain *Assilina spira planospira* BOUBÉE 1831. The association allows us to situate this cycle in SBZ 14 (Middle Lutetian).

Third Sedimentary Cycle

In the lower part of the transgressive system of the cycle (lower part of the Coll de Malla Marl Formation (Clavell et al, 1970), the species *Nummulites tavertetensis* REGUANT and CLAVELL 1967, *N. crusafonti* REGUANT and CLAVELL 1967 and *A. spira planospira* BOUBÉE 1831 were identified, whereas the middle and upper parts of this cycle contain, according to Serra-Kiel et al. (1997), *Nummulites crassus* BOUBÉE 1831 and transitional forms from *N. crusafonti* REGUANT and CLAVELL 1967 to *N. puigsecensis* REGUANT and CLAVELL 1967 (*N.* aff. *crusafonti*) and from *N. tavertetensis* REGUANT and CLAVELL 1967 to *N. deshayesi* D'ARCHIAC and HAIME 1853 (*N.* aff. *tavertetensis*). From these associations, the lower part of the cycle is attibuted to SBZ 14 (Middle Lutetian), and the middle and upper parts to SBZ 15 (Middle Lutetian).

According to Burbank et al. (1992a) the lower and middle part of the Coll de Malla Marl Formation (Clavell et al., 1970) show a normal magnetozone interpreted as Chron 20n.

Fourth Sedimentary Cycle

The marl with *Nummulites* at the base of this cycle, the upper part of the Coll de Malla Marl Formation (Clavell et al., 1970) and the Bracons Formation (Gich, 1969, 1972) in the Vic-Amer area contain *N. puigsecensis* REGUANT and CLAVELL 1967, *N. beaumonti* D'ARCHIAC and HAIME 1853 and *Assilina exponens* (SOWERBY 1840), whereas the carbonate levels of the middle and upper parts of this cycle contain *Alveolina* aff. *fusiformis* SOWERBY 1850, according to Barnolas et al. (1983) and Serra-Kiel et al. (1997). In the Empordà area, the carbonate levels of this cycle (Fig. 7a) contain *N. herbi* SCHAUB 1981, *N. praepuschi* SCHAUB 1981 and *N. discorbinus* (SCHLOTHEIM 1820), according to Serra-Kiel et al. (1997). From this association, this cycle is situated in SBZ 16 (Late Lutetian). According to Burbank et al. (1992a), the upper part of the Coll de Malla Marl Formation (Clavell et al., 1970), which belongs to the 4th Lutetian cycle, and lower part of the Folgueroles Sandstone Formation (Reguant, 1967), which belongs to the lower part of the 1st Bartonian cycle, show a normal magnetozone interpreted as Chron 19. These data allow us to include SBZ 16 within the magnetozone 19.

Bartonian

First Sedimentary Cycle.

Transgressive system

In the Igualada area (Fig. 9), the 1st Bartonian transgressive system is represented by the lower and middle parts of the Collbàs Formation (Ferrer, 1971; Travé et al., 1999). In this 1st transgressive system, the association of the larger foraminifers, according to Hottinger (1960), Ferrer (1971, 1973), Schaub (1981), Serra-Kiel (1984) and Serra-Kiel et al. (1997), contain *Alveolina fragilis* HOTTINGER 1960, *Nummulites perforatus* (MONTFORT 1808), *N. striatus* (BRUGUIÈRE 1792), *N. hottingeri* SCHAUB 1981, *N. beaumonti* D'ARCHIAC and HAIME 1853, *N. vicaryi* SCHAUB 1981, *N. cyrenaicus* SCHAUB 1981, *N. ptukhiani* KACHARAVA 1969, *N. praegarnieri* SCHAUB 1981, *N. variolarius* (LAMARCK 1804), *Assilina schwageri* (SILVESTRI 1928) and *Operculina roselli* (HOTTINGER 1977). This association indicates SBZ 17, Early Bartonian or Biarritzian *sensu* Hottinger and Schaub (1960).

In the magnetostratigraphic section elaborated by Burbank et al. (1992a), this 1st Bartonian transgressive system is represented by the Folgueroles Sandstone Formation (Reguant, 1967) and the lower part of the Manlleu Marl Member (Reguant, 1967; Barnolas et al., 1983), showing that this interval is located within the magnetozone 18.

Regressive system

The 1st Bartonian regressive system in the Igualada area (Fig. 9) is represented by the upper part of the Collbàs Formation (Ferrer, 1971), according to Travé et al. (1999), and the Castellolí Deltaic Complex (Travé et al., 1999). According to Ferrer (1971), the upper part of the Collbàs Formation contains *N. striatus* (BRUGUIÈRE 1792), *N. ptukhiani* KACHARAVA 1969, *N. praegarnieri* SCHAUB 1981, *Assilina schwageri* (SILVESTRI 1928) and *Operculina roselli* (HOTTINGER 1977). This association indicates SBZ 17, Early Bartonian.

In the magnetostratigraphic section (Burbank et al., 1992a), this regressive system is represented by the middle-upper part of the Manlleu Marl Member (Reguant, 1967; Barnolas et al., 1983) and the Orís Sandstone (Burbank et al., 1992a). Both lithostratigraphic units are equivalent to the Puigsacalm Sandstone Formation (Gich, 1969, 1972), and show that this interval is located within magnetozone 18.

Second Sedimentary Cycle

Transgresive system

The 2nd Bartonian trangressive system in the Igualada area (Fig. 9) corresponds to the middle-upper part of the Igualada Formation (Ferrer, 1971) and the Castellolí Marl and Limestone (Travé et al., 1999). The maximum flooding surface of this transgressive system at the base of the Tossa Formation (Ferrer, 1971), according to Ferràndez-Cañadell (1999), contains the association made up of *Discocyclina radians radians* (D'ARCHIAC 1850), *D. augustae augustae* WEIJDEN 1940 and *Asterocyclina stellaris* (BRÜNNER 1848 *in* RÜTIMEYER 1850), indicating SBZ 18, Late Bartonian.

According to Ferrer (1971), the middle and upper parts of the Igualada Formation contain the planktic foraminiferal associations that indicate the P 14 and P 15 biozones respectively.

In the magnetostratigraphic section elaborated by Burbank et al. (1992a), this 2nd Bartonian transgressive system is represented by the Gurb Marl Member (Reguant, 1967; Barnolas et al., 1983), showing that this interval is located between the lower part of the magnetozone 18 and the upper part of magnetozone 17. Moreover, according to Ferràndez-Cañadell in Serra-Kiel et al. (1997), the presence of *Discocyclina radians radians* (D'ARCHIAC 1850), *D. augustae augustae* WEIJDEN 1940, *Asterocyclina stellaris* (BRÜNNER 1848 *in* RÜTIMEYER 1850), *Orbitoclypeus daguini* (NEUMANN 1958) and *Assilina schwageri* (SILVESTRI 1928), indicates SBZ 18, Late Bartonian.

Regressive system

The 2nd Bartonian regressive system in the Igualada area (Fig. 9) comprises the middle part of the Igualada Formation (Ferrer, 1971), the Tossa Formation (Ferrer, 1971) and the Upper Deltaic Reefal Complex (Travé et al., 1999), according to Travé et al. (1999). These units, according to Papazzoni and Sirotti (1995), Serra-Kiel et al. (1997), Ferràndez-Cañadell (1999), Romero (1996, 2001), Romero et al. (1999) and Hottinger et al. (2001), contain *Nummulites chavannesi* DE LA HARPE 1878, *N. praegarnieri* KACHARAVA 1969, *N. ptukhiani* KACHARAVA 1969, *N. incrassatus incrassatus* DE LA HARPE 1883, *Assilina schwageri* (SILVESTRI 1928), *Operculina roselli* HOTTINGER 1977, *Asterocyclina stellaris* (BRÜNNER 1848 *in* RÜTIMEYER 1850), *Discocyclina pratti* (MICHELIN 1846), *D. radians radians* (D'ARCHIAC 1850), *D. augustae augustae* WEIJDEN 1940, *D. augustae olianae* ALMELA and RÍOS 1942, *Heterostegina reticulata reticulata* RÜTIMEYER 1850, *H.* aff. *reticulata italica* HERB 1978, *Assilina schwageri* (SILVESTRI 1928), *Operculina roselli* HOTTINGER 1977, *Pellatispira madaraszi* (HANTKEN 1875), *Biplanispira absurda* UMBGROVE 1938, and *Halkyardia minima* (LIEBUS 1919), indicating SBZ 18.

In the magnetostratigraphic section elaborated by Burbank et al. (1992a), this regressive system is represented by the Vespella Marl Member (Reguant, 1967; Barnolas et al. 1983) and the Sant Matí Xic Limestone Formation (Reguant, 1967), equivalent to the Tossa Formation (Ferrer, 1971). According to Reguant (1967), Papazzoni and Sirotti (1995) and Serra-Kiel et al. (1997), the Sant Martí Xic Limestone Formation contains the following species of larger foraminifers: *Nummulites biedai* SCHAUB 1962, *N. praegarnieri* SCHAUB 1981, *N. ptukhiani* KACHARAVA 1969, *N. cyrenaicus* SCHAUB 1981, *Nummulites chavannesi* DE LA HARPE 1878, *Assilina schwageri* (SILVESTRI 1928), *Operculina roselli* HOTTINGER 1977, *Heterostegina reticulata reticulata* RÜTIMEYER 1850, *Discocyclina augustae augustae* WEIJDEN 1940, *D. radians radians* (D'ARCHIAC 1850), *Asterocyclina stellaris* (BRÜNNER 1848 *in* Rütimeyer 1850), *Orbitoclypeus varians* (KAUFMANN 1867) and *Halkyardia minima* (LIEBUS 1919), which indicate SBZ 18, Late Bartonian. The magnetostratigraphic data show that this interval is located within magnetozone 17.

The chronostratigraphic correlation between the magnetostratigraphic section elaborated by Burbank et al. (1992a), the marine Lutetian and Bartonian sedimentary cycles and the larger foraminifer biozones (SBZ) is shown in Figure 10.

Finally, the presence in the carbonate beds of the Terminal Complex of *Malatyna vicensis* SIREL and AÇAR 1998, *Rhabdorites malatyaensis* (SIREL 1976) and *Orbitolites*, according to Sirel and Açar (1998), does not permit a biostratigraphic attribution. This unit is located above the well-dated Bartonian marine units and below the Priabonian continental deposits of the Artés Formation (Ferrer, 1971), which were well-dated using charophytes (Anadón et al., 1992). The unit is placed in an uncertain chronostratigraphic position, Late Bartonian-Early Priabonian.

ACKNOWLEDGMENTS

This research was supported by DGICYT grant PB 98-1293, and is a contribution to IGCP 393 of the Grup Consolidat de Recerca "Geodinàmica i Anàlisi de Conques" of the Barcelona University and the Research Commission of the Generalitat de Catalunya. We thank J. Telford and J. Soler for the revision of the English version.

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Manuscript received January 2002;
revision accepted December 2002.