Geologica Acta, Vol.7, Nos 1-2, March-June 2009, 177-194
DOI: 10.1344/105.000000268
Available online at www.geologica-acta.com

geologica acta

Redefinition of the Ilerdian Stage (early Eocene)

V. PUJALTE [1] **J.I. BACETA** [1] **B. SCHMITZ** [2] **X. ORUE-ETXEBARRIA** [1] **A. PAYROS** [1] **G. BERNAOLA** [1] **E. APELLANIZ** [1] **F. CABALLERO** [1] **A. ROBADOR** [3] **J. SERRA-KIEL** [4] **J. TOSQUELLA** [5]

[1] **Department of Stratigraphy and Paleontology, Faculty of Science and Technology, University of the Basque Country**
P.O. Box 644, E-48080 Bilbao, Spain. Pujalte E-mail: victoriano.pujalte@ehu.es
Baceta E-mail: juanignacio.baceta@ehu.es Orue-Etxebarria E-mail: xabi-orueetxebarria@ehu.es
Bernaola E-mail: gilen.bernaola@ehu.es Payros E-mail: a.payros@ehu.es
Apellaniz E-mail: estibaliz.apellaniz@ehu.es Caballero E-mail: fernando.caballero@ehu.es

[2] **Department of Geology, University of Lund**
Sölvegatan 12, SE-22362 Lund, Sweden.E-mail: birger.schmitz@geol.lu.se

[3] **Área de Cartografía Geocientífica, IGME**
c/ La Calera 1, 28760, Tres Cantos, Madrid, Spain. E-mail: a.robador@igme.es

[4] **Departament d'Estratigrafia, Paleontologia i Geociències Marines, Facultat de Geologia, Universitat de Barcelona**
Zona Universitària de Pedralbes, 08028 Barcelona, Spain. E-mail: josepserra@ub.edu

[5] **Departamento de Geodinámica y Paleontología, Facultad de Ciencias Experimentales, Universidad de Huelva**
21071 Huelva, Spain. E-mail: josep@uhu.es

ABSTRACT

The Ilerdian Stage was created by Hottinger and Schaub in 1960 to accommodate a significant phase in the evolution of larger foraminifera not recorded in the northern European basins, and has since been adopted by most researchers working on shallow marine early Paleogene deposits of the Tethys domain. One of the defining criteria of the stage is a major turnover of larger foraminifera, marked by the FO's of *Alveolina vredenburgi* (formerly *A. cucumiformis*) and *Nummulites fraasi*. There is now conclusive evidence that this turnover was coeval with the onset of the Carbon Isotope Excursion (CIE) and, consequently, with the Paleocene-Eocene (P-E) boundary, a temporal correspondence that reinforces the usefulness of the Ilerdian as a chronostratigraphic subdivision of the early Eocene in a regional context. However, in addition to the paleontological criteria, the definition of the Ilerdian was also based on the designation of two reference sections in the southern Pyrenees: Tremp (stratotype) and Campo (parastratotype). In both sections, the base of the stage was placed at the lowest marine bed containing *A. vredenburgi* specimens. Using the CIE as a correlation tool we demonstrate that these two marine beds occur at different chronological levels, being older in Campo than in Tremp. Further, we show that both beds are in turn younger than the lowest strata with Ilerdian larger foraminifera at the deep-water Ermua section in the Basque Basin (western Pyrenees). Since the age of stage boundaries must be the same everywhere, the choice of these stratotype sections was misleading, since in practice it resulted in the Ilerdian being used as a facies term rather than as a chronostratigraphic unit. To eliminate that conflict, and yet be respectful with established tradition, we propose to redefine the Ilerdian Stage following a procedure similar to the one used by the International Commission on Stratigraphy to establish global chronostratigraphic standards, namely: by using a "silver spike" to be placed in the Tremp section at the base of the Claret Conglomerate, a widespread lithological unit that in the Tremp-Graus Basin coincides with the onset of the CIE. The redefined regional Ilerdian Stage becomes thus directly correlatable to the lower part of the global Ypresian Stage, as currently defined by the International Commission on Stratigraphy.

KEYWORDS **Ilerdian stage. Silver spike. Larger foraminifera. CIE. P-E boundary. Pyrenees. Tremp-Graus Basin.**

INTRODUCTION

The Ilerdian Stage was defined by Hottinger and Schaub (1960) to account for a significant step in the evolution of alveolinids, nummulitids, and other larger foraminifera not recorded in the northern European basins. Its utility was soon recognized, in spite of the fact that it was created around a century after the definition of the now standard early Paleogene Stages (Danian, 1849; Selandian, 1924; Thanetian, 1873; Ypresian, 1849). Thus, Luterbacher (1973) pointed out that "*the Ilerdian not only represents an important lapse of time in the evolution of the larger foraminifera, but also in the zonal successions based on pelagic microfossils*". Soon after, in a Meeting of the International Working Group on Paleogene Stratigraphy held in Paris in November 1974, it was concluded that "*the Ilerdian is not only a stage that allows a clarification of the Mesogean Paleogene stratigraphy, but that may also serve as the base to build up a stratigraphic succession better defined than the one of the north* [European] *basins*" (Pomerol, 1975, p. 213; in French in the original). Ever since, many researchers working on shallow marine deposits of the Tethys domain have adopted the Ilerdian Stage as a useful regional chronostratigraphic subdivision of the early Paleogene (e.g., Tambareau, 1972; Tambareau and Villatte, 1974; Tambareau et al., 1989; Freytet and Plaziat, 1970; Boukhary et al., 1995; Colakoglu and Ozcan, 2003; Murru et al., 2003; Rasser et al., 2005; Ozgen-Erdem et al., 2007).

Interest in the Ilerdian Stage was revived in connection with the activities of the IGCP projects 286 (Early Paleogene Benthos) and 308 (Paleocene-Eocene boundary events). Molina et al., (1992), for instance, reviewed the Ilerdian stratotype sections and concluded that "*the Ilerdian is today a very well defined stage, represented by thick and relatively continuous marine sediments, excellently exposed and very rich in marine microfossils, being now one of the best known European stages*". Likewise, Hottinger (1998) declared that "*the collaborators in IGCP 286 recommend to fix the Paleocene-Eocene boundary in shallow carbonate deposits at the level of the second turnover event* [of larger foraminifera], *which can be easily recognized by the microfacies documenting the rise of larger-sized and distinctly dimorph foraminifera in the field with a hand lens*". The second larger foraminiferal turnover mentioned by Hottinger (1998), which was later referred to by the acronym LFT (Orue-Etxebarria et al., 2001), is one of the criteria originally used by Hottinger and Schaub (1960) to pinpoint the base of the Ilerdian Stage.

The Global Standard Stratotype Section and Point (GSSP) for the base of the Eocene (or the P-E boundary) is now officially located in the Dababiya Quarry beds, near Luxor (Egypt), at the base of a 3.5 m thick interval characterized by a prominent global negative excursion in $\delta^{13}C$ (Carbon Isotopic Excursion, CIE) (Aubry el al., 2007). The onset of the CIE is in fact the main criterion for correlating the P-E boundary elsewhere (Luterbacher et al., 2000; 2004). It is now demonstrated that the LFT and the onset of the CIE are coeval (Orue-Etxebarria et al., 2001; Pujalte et al., 2003a; Pujalte et al., this issue; Scheibner and Speijer, this issue). This temporal concurrence further reinforces the usefulness of the regional Ilerdian Stage.

However, in addition to the aforementioned paleontological criterion (i.e., the LFT), the original definition of the Ilerdian was also based on the designation of two reference sections in the southern Pyrenees, the Tremp section (stratotype) and the Campo section (parastratotype) (Schaub, 1969). In both sections, the base of the stage was situated at the lowest marine bed containing specimens of *Alveolina vredenburgi* (formerly *A. cucumiformis*), two horizons that in most published papers were later assumed to be coeval. However, using the CIE as a correlation tool, it can now be demonstrated that the traditional base of the Ilerdian is older in Campo than in Tremp. Since the age of stage boundaries must necessarily be the same everywhere this diachrony is not acceptable.

A first purpose of this paper is to document the abovementioned diachronism, by providing information on stable isotopes and larger foraminifera from both the Tremp and Campo sections, and also from other important reference sections of the Pyrenees and the Basque Basin. We then show that, by following the guidelines of the International Commission on Stratigraphy, the inconsistency can be easily resolved. Finally, we briefly discuss the position of the redefined stage relative to the global Ypresian stage.

PALEOGEOGRAPHICAL SETTING

The early Paleogene paleogeography of the Pyrenean domain can be envisaged as an E-W elongated deep-water embayment, opening westward into the Bay of Biscay and flanked on its northern, southern and eastern sides by a shallow carbonate platform system (Fig. 1A). The platform system, which was laterally continuous and generally wider than 50 km, had a ramp geometry during the early Danian but was transformed into a rimmed shelf during the late Danian, following the growth of a well-defined coralgal barrier reef (Plaziat, 1981; Baceta, 1996; Baceta et al., 2004, 2005; Robador, 2005). That upper Danian barrier reef later persisted as a buried feature, which delineated the margin of the platform system during the Selandian, Thanetian and early Ilerdian time. Seaward of this margin a base-of-slope apron was created, which is represented by variable proportions of carbonate breccias, coarse- and fine-grained calciturbidites derived from the shelf, alternating with autochtonous (hemi)pelagic limestones and marls. One important section of these Paleocene-lower Ilerdian base-of-slope deposits occurs near the town of Ermua in the Basque Country (Baceta, 1996; Orue-Etxebarria et al., 1996; Schmitz et al., 2001), and is discussed below. In turn, the base-of-slope apron deposits grade to successions dominated by (hemi)pelagites toward the central part of the deep-water embayment, as exemplified by the well-known Zumaia section (Fig. 1A).

Landward from the carbonate platform system there existed subaerial coastal plains, which are represented by red-bed "Garumnian" deposits. These plains were particularly well developed in the so-called Tremp-Graus Basin, because of the abundant clastic supply coming from mountains created in the eastern Pyrenees during the late Cretaceous tectonic compressional phase (Puigdefàbregas and Souquet, 1986; Vergès and Martínez, 1988; Cuevas, 1992).

Both the Ilerdian stratotype and parastratotype, and also another important reference section discussed below (Esplugafreda), are situated in the Tremp-Graus Basin (Fig. 1B). The Campo section was near the Paleocene shoreline, and thus was alternatively in subaerial conditions or submerged by a shallow sea. Instead, the Tremp and Esplugafreda sections were placed well within the coastal alluvial plain and stayed subaerial throughout the Paleocene. The sea, however, also flooded both localities during the important transgression that started around the beginning of the Ilerdian Stage (Baceta et al., 2004).

THE ILERDIAN LOWER BOUNDARY: DEFINITION AND PRIOR USAGE

Since the deposits of their newly defined stage are extensively outcropping in the Lleida province (*Ilerda* in Latin language), Hottinger and Schaub (1960) named it "Ilerdian" and they chose the section exposed along the road C-1311, between Tremp and the Montllobat Pass, as its stratotype. That stratotype, however, was criticized because the Ilerdian deposits are encased there between non-marine sequences, making it impossible to define its limits with older and younger marine sequences. To overcome that problem, Schaub (1969) designed the Campo section as the Ilerdian parastratotype. Ever since, both sections have been extensively studied, and excellent descriptions of them can be found in Luterbacher (1973), Schaub (1973), Tambareau and Villatte, 1974, Molina et al., (1992, 1995), Eichenseer and Luterbacher (1992), Serra-Kiel et al., (1994) or De Renzi (1996), to name but

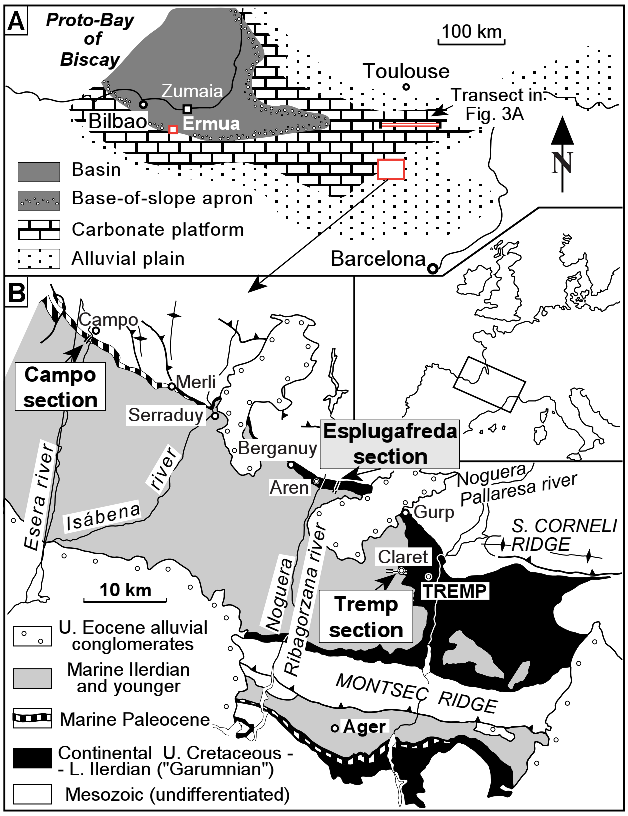

FIGURE 1 | **A) Simplified Paleocene paleogeography of the Pyrenean domain, showing the location of the Ermua and Zumaia sections. B) Geological map of the Tremp-Graus Basin, showing the location of the Esplugafreda, Tremp and Campo sections, and of other localities mentioned in the text.**

a few. Consequentely, the fossil content and the internal zonation of the Ilerdian Stage are now very well known. Yet, as discussed below, the position of the lower boundary is still problematic.

In effect, all the authors listed above did place the base of the Ilerdian Stage at Tremp and/or at Campo in accordance with the original definition of Hottinger and Schaub (1960) and Schaub (1969). Molina et al., (1992), for instance, explicitly stated that the lower limit of the Ilerdian in the Tremp section *"is perfectly exposed on the north side of the road, one kilometre from Claret,* ***overlying the continental deposits (Garumnian facies) of the Tremp Formation***" (p. 145, emphasis added). Further, the same authors indicated that, at Campo, "*the bottom of the Ilerdian parastratotype is perfectly exposed on the north side of the road to Ainsa beginning at 0.34 km*, adding that "*there is a brief non-marine intercalation that contains charophytes between the Late Thanetian* ***and the first Ilerdian sediments***" (emphasis added). Eichenseer and Luterbacher (1992) placed the lower boundary of the Ilerdian at exactly the same horizons as Molina et al., (1992), as shown in Fig. 2.

Indeed, the base of the Ilerdian is drawn in most geological maps of the Tremp-Graus Basin at the lower limit of the "*Alveolina* limestones" or correlative marine strata (eg., Institut Cartogràfic de Catalunya, 2002). The same procedure can be found in most papers dealing with Ilerdian deposits of the southern and northern Pyrenees, as exemplified by the reconstructed cross-sections reproduced in Figs. 3A and 3B. It should be noted that in these two cross-sections the Ilerdian strata always overlie non-marine deposits, a circumstance that entails that the non-marine/marine transition cannot be accurately dated. In spite of that, this transition was assumed to be isochronous, probably because the lowest Ilerdian marine beds in both transects contain larger foraminifera of the earliest Ilerdian.

To our knowledge, Payros et al., (2000) were among the first authors to challenge that general assumption, when they suggested that at Campo the base of the Ilerdian Stage may be located at the exposure surface with *Microcodium* remains capping the Thanetian marine carbonates and, therefore, that the overlying non-marine interval pertains to the Ilerdian. That suggestion, however,

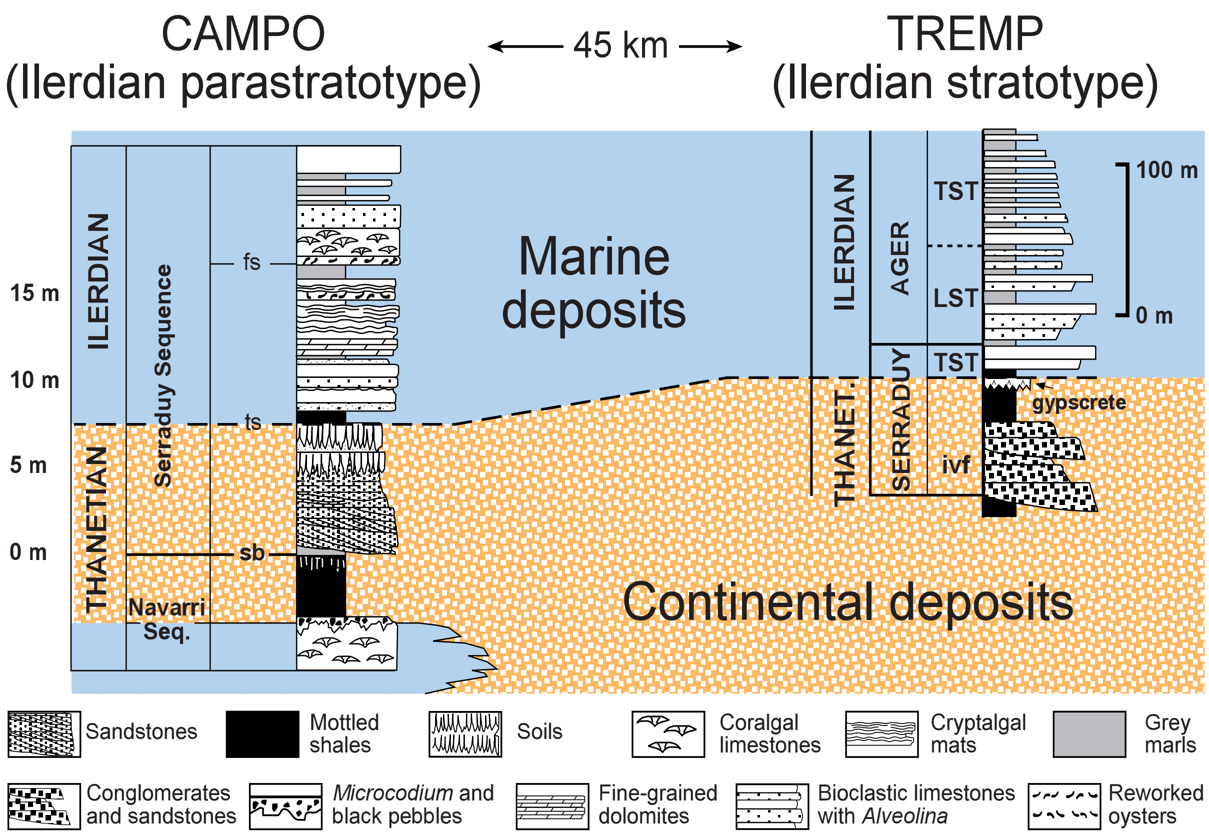

FIGURE 2 **An example of the classic correlation between the stratotypic and parastratotypic sections of the Ilerdian stage. (Columnar sections after Figs. 9 and 13 of Eichenseer and Luterbacher, 1992, slightly modified; note that the two sections are drawn at different scales). Ivf, incised valley fill; TST, transgressive system tract; LST, Lowstand system tract; sb, sequence boundary; ts, transgressive surface; fs, flooding surface.**

was not supported by paleontological data and, at any rate, it will be shown here to be inexact.

DESCRIPTION OF REFERENCE SECTIONS

The CIE as a correlation tool, and its application in the Pyrenees

Global temperatures increased abruptly by 5 to 10ºC about 55.5 million years ago, during an interval of ca. 200 thousand years (ky) known as the Paleocene-Eocene thermal maximum. This climatic event is commonly attributed to a massive release of ^{13}C-depleted carbon to the ocean-atmosphere system, because of its temporal concurrence with a global decrease of marine carbon isotope values (CIE) and an important shoaling of the calcite compensation level in the oceans (Dickens et al., 1997; Zachos et al., 2003). The initial shift in carbon isotopic values was geologically instantaneous, probably less than 10 ky (Bains et al., 1999; Röhl et al., 2000, 2007). Because of that, the onset of the CIE can be considered isochronous for all practical purposes.

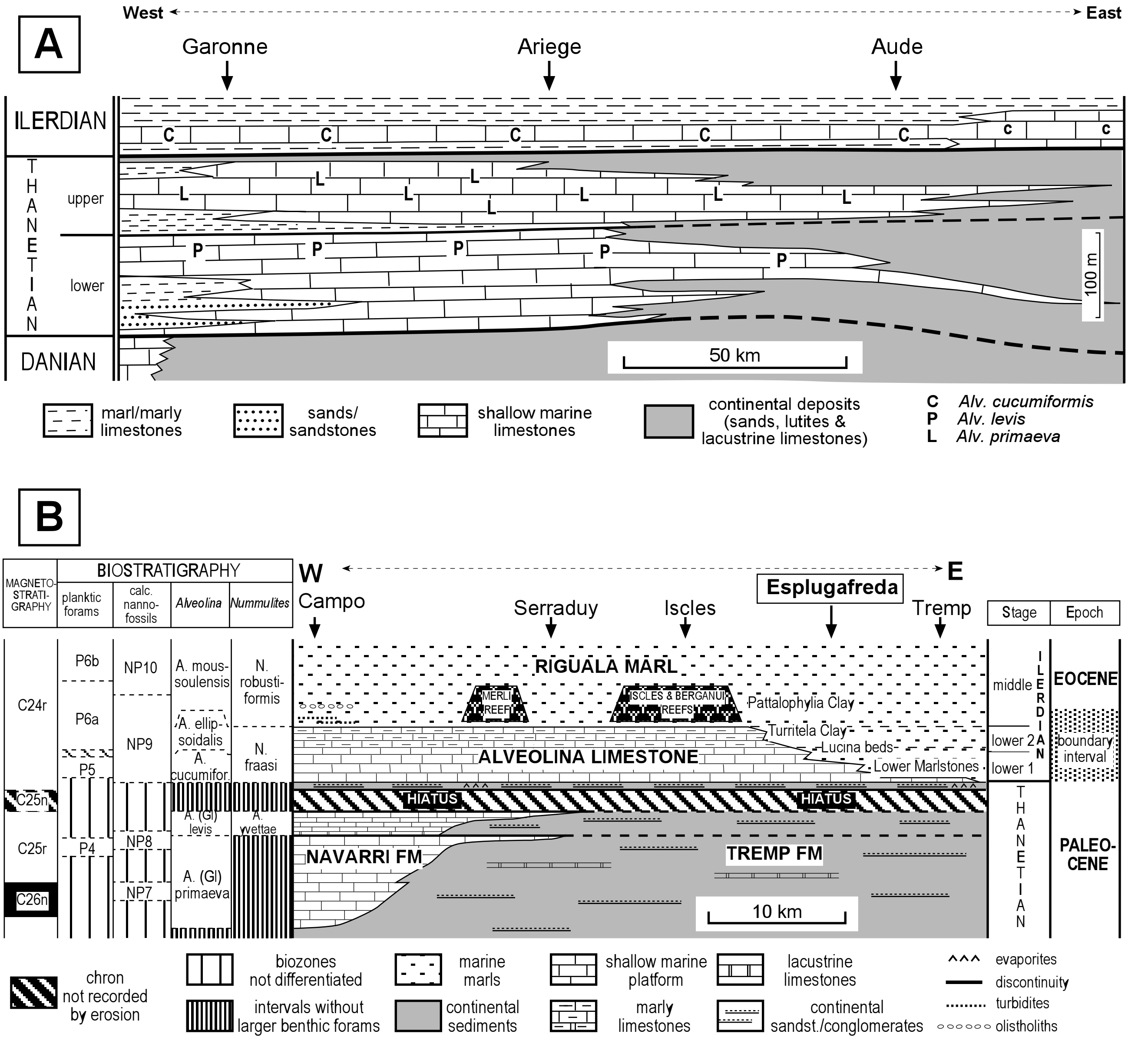

FIGURE 3 **Stratigraphic cross-sections of the Thanetian-early Ilerdian interval: A) Across the northern Pyrenees, after Plaziat, 1975 (for location, see Fig. 1A); B) Across the Tremp-Graus Basin, after Serra-Kiel et al. (1994).**

Reconstructions of past isotopic variations in seawater should ideally be based on analyses of well-preserved, monospecific calcite tests or shells, a material that is not possible to recover in Pyrenean P-E sections. However, isotopic analysis of whole-rock Paleocene and lowest Eocene fine-grained limestones of the Basque Basin have produced results that mimic records measured on well-preserved marine planktic and deep-benthic foraminiferal tests, probably because these limestones were indurated during early diagenesis and became closed systems with respect to carbon exchange (Schmitz et al., 1997, 2001). One of these sections (Ermua) was situated in the early Paleogene base-of-slope apron (Fig. 1A), and includes numerous calciturbidites that contain larger foraminifera (Orue-Etxebarria et al., 1996), thus permitting a direct comparison between the LFT and the CIE (see below).

The $\delta^{13}C$ and $\delta^{18}O$ values of carbonate soil nodules have long been recognized as valuable paleoclimatic proxies (Cerling, 1984), and have since been widely used to pinpoint the CIE in continental sections (Koch et al., 1995; Bowen et al., 2001). High-resolution studies in the expanded Bighorn Basin of Wyoming (USA) have further demonstrated that the CIE recorded in carbonate soil nodules accurately correlates with the one found in deep-sea oceanic boreholes (Bains et al., 2003; Koch et al., 2003). The CIE is therefore a powerful tool to correlate continental and marine sequences and, mainly because of that, the onset of the event has been selected as the main criterion to pinpoint the P-E boundary (Luterbacher et al., 2000, 2004; Aubry et al., 2007).

The "Garumnian" deposits of the Tremp-Graus Basin are rich in carbonate soil nodules, and their study has recently allowed location of the CIE in several reference P-E boundary sections, including the Ilerdian stratotype and parastratotype (Figs. 4 and 5; for actual isotopic data see: Schmitz and Pujalte, 2003, 2007, and Pujalte et al., this issue). The pre-CIE, syn-CIE and post-CIE intervals of the Tremp and Campo sections, and of two additional key sections of the Pyrenees, are described below.

Esplugafreda section

This section is located along a small tributary of the Esplugafreda ravine, in the northern margin of the Tremp-Graus Basin (Fig. 1B; UTM coordinates: base, X= 314772; Y= 4680040; top, X= 314713; Y= 4679583). Médus and Colombo (1991) gave a general description of the section and reported Thanetian pollen assemblages from its upper part. Schmitz and Pujalte (2003, 2007) and Pujalte and Schmitz (2005, 2006) later provided more detailed information, including abundant data on stable isotopes.

The pre-CIE interval of this section is ca. 225 m thick, of which only the upper part is shown in Figs. 4A and 5A. It is mostly made up of red silty mudstones with intercalated calcarenites and calcareous conglomerates. The mudstones contain numerous paleosols with well developed Bca horizons about 1 m thick, characterized by distinct carbonate nodules up to few cm in diameter. They may also contain carbonate rhizocretions and, occasionally, horizons with gypsum veins and crystals. The calcarenites and conglomerates generally occur as multistorey channelized bodies ranging between 2-9 m in thickness and 10-200 m in lateral extent. These deposits usually exhibit cross-bedding and clast-imbrication, indicative of traction currents, but examples of chaotic fabrics attributable to debris-flows have also been observed. *Microcodium* remains are abundant, mainly within the Bca soil horizons but also in the channelized calcarenites. The $\delta^{13}C$ composition of soil nodules from this interval shows a relatively stable trend, with values in the range –9‰ to –6‰, which are typical for soil carbonate nodules from the Paleocene (Bowen et al., 2001, 2002).

The syn-CIE interval at Esplugafreda is comprised of two different units, the Claret Conglomerate below and the "Yellowish Soils" above (Figs. 4A and 5A). The former is a sheet-like, complex alluvial unit composed mostly of coarse-grained calcareous conglomerates and pebbly calcarenites, with minor intercalations of grey and red mudstones (Schmitz and Pujalte, 2007). At Esplugafreda the thickness of the unit ranges between 2 and 4 m, but elsewhere may reach up to 8 m. However, despite its comparatively modest thickness, the Claret Conglomerate can be traced almost continuously in strike and dip sections (except where eliminated by recent erosion or buried under younger deposits) from Gurp to near Merli (ca. 30 km), and from Gurp to Claret (ca. 8 km; see Fig. 1B). Another distinctive feature of this unit is the abundance of reddish carbonate clasts, which in some exposures may constitute almost half of the total clast population.

The Yellowish Soils unit is about 20 m in the Esplugafreda section and is mainly made up of silty mudstones of a distinctive yellowish-orange colour in weathered exposures (Fig. 4A), with some intercalations of channelized sandstones. These Yellowish Soils contain abundant, small-sized (< 1cm) carbonate nodules that, in contrast to those of the pre-CIE red beds, are not concentrated in discrete Bca horizons but occur almost uniformly dispersed throughout the unit. Such a distribution is thought to indicate a cumulate soil (*sensu* Wright and Marriot, 1996), a type of soil that develops where small but frequent increments of sediments cause a near continuous aggradation of the sediment surface, with vertical displacement of the zone of active pedogenesis and abandonment of lower levels of the soil. The $\delta^{13}C$ values of the carbonate nodules throughout the Yellowish Soils vary between –14‰ to –12‰. This implies a negative shift of ca. 6‰, which can only reflect the unique carbon-cycle perturbation of the CIE. Instead, the $\delta^{18}O$ values throughout the section maintain a stable trend, which indicates that the large shift in $\delta^{13}C$ does not reflect a diagenetic artifact. At Esplugafreda, as in most other places, the Claret Conglomerate does not contain carbonate nodules. However, in a section situated 4 km to the west (Berganuy, fig. 1), the Claret Conglomerate encloses a 1.5 m thick intercalation of red mudstones with carbonate nodules (Fig. 5B), which gave $\delta^{13}C$ values between –13‰ to –9‰, with

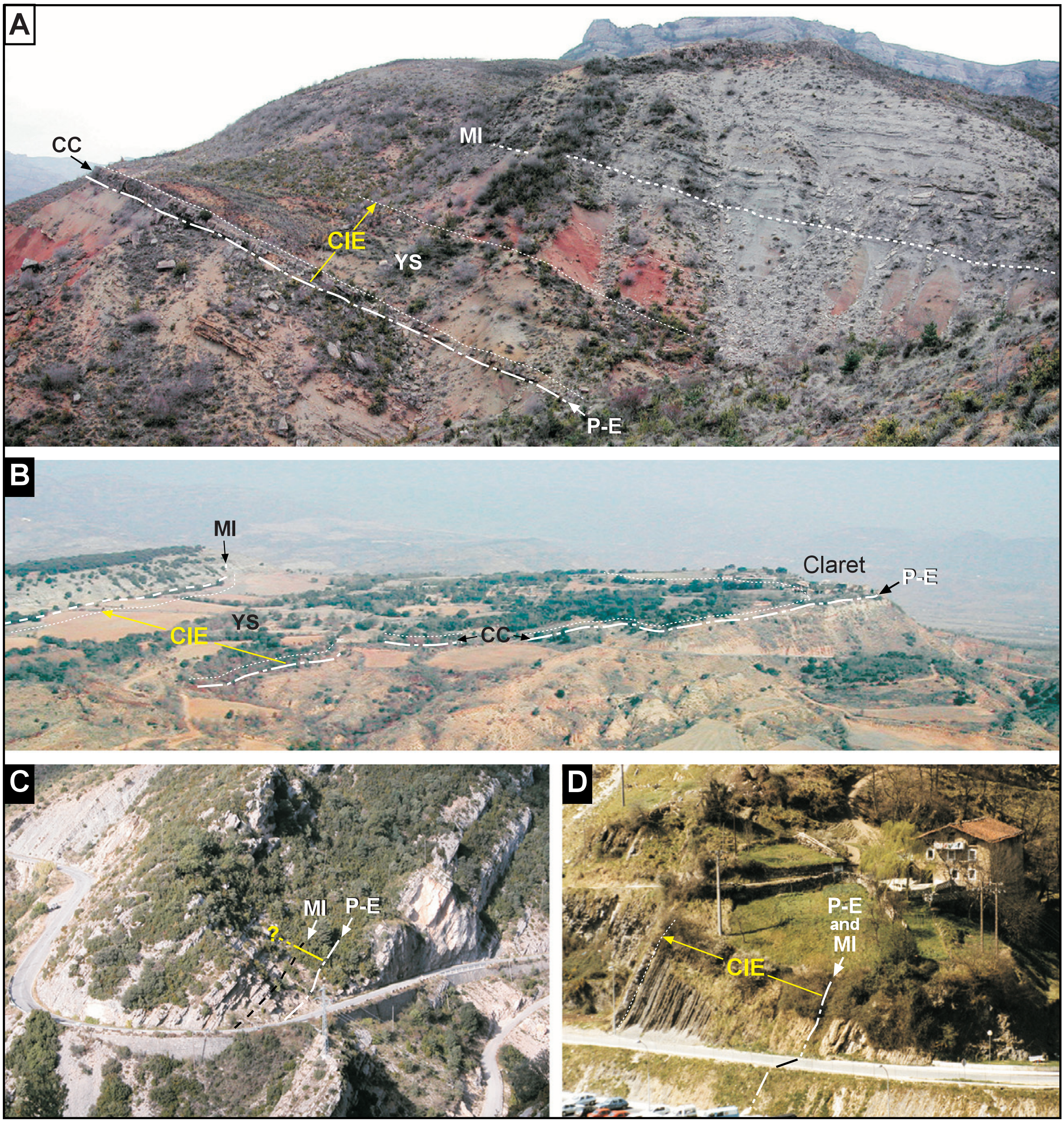

FIGURE 4 | **Field photos of the four sections described in the paper: A) Esplugafreda; B) Tremp (Ilerdian stratotype); C) Campo (Ilerdian parastratotype); D) Ermua. Key: P-E, Paleocene-Eocene boundary; CIE, Carbon Isotopic Excursion; CC, Claret Conglomerate; YS, Yellowish Soils; MI, Marine Ilerdian.**

most values in the upper, less negative part of this range (Schmitz and Pujalte, 2007; Fig. 5B). These intermediate values indicate that the Claret Conglomerate formed during the earliest part of the CIE, when a rapid buildup of ^{13}C-depleted carbon dioxide took place in the atmosphere.

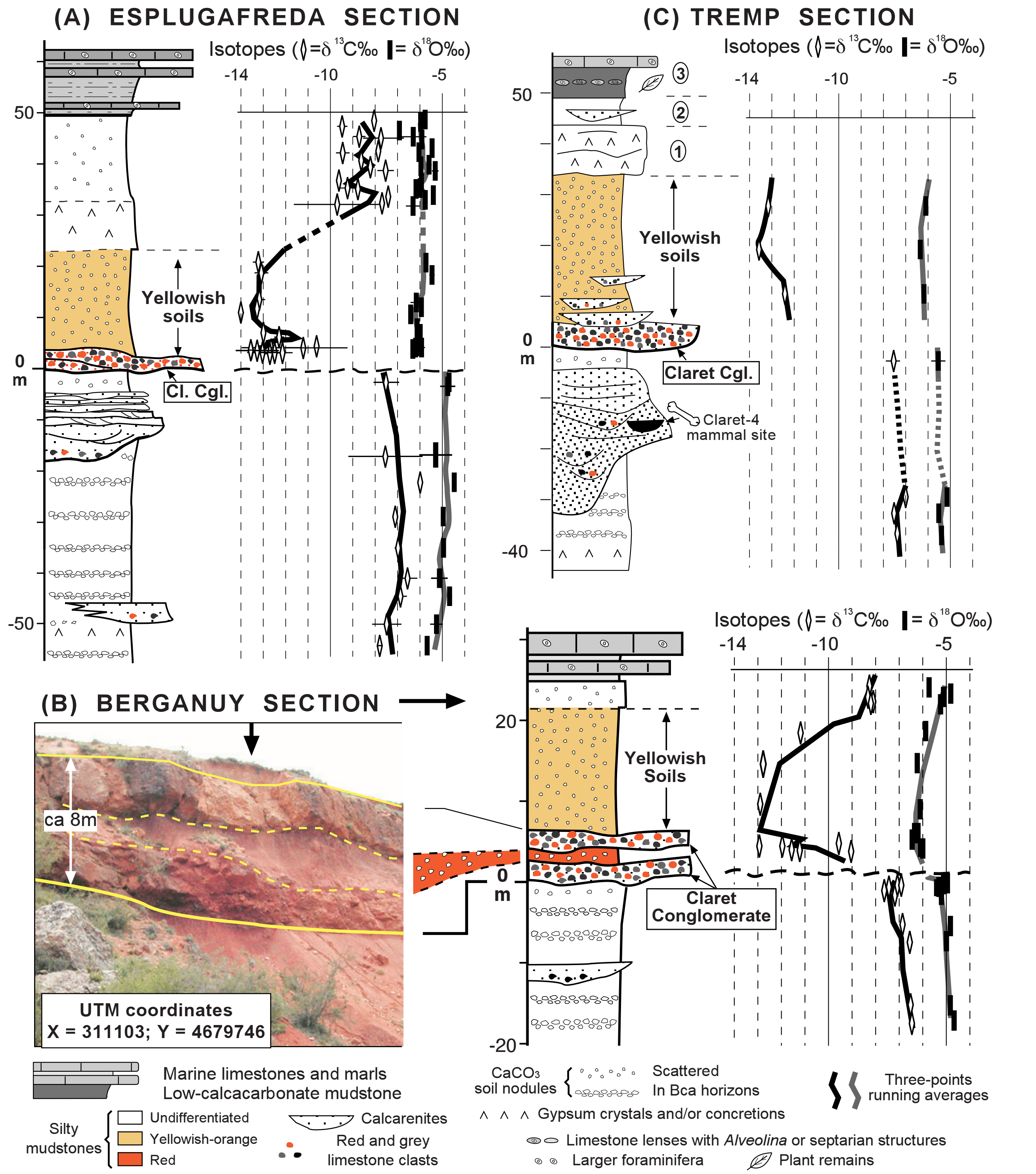

FIGURE 5 | **Columnar and isotopic profiles of key P-E sections of the Tremp-Graus basin, with a close-up photo of the Claret Conglomerate at Berganuy (after Schmitz and Pujalte, 2007, modified from their repository data file 2007048).**

The post-CIE interval at Esplugafreda includes three parts. The lower one is ca. 10 m thick and is mostly made up of silty mudstones of a deep red colour (Fig. 4A). These mudstones contain variable amounts of gypsum, either as veins, crystals or as root-like concretions, but no carbonate nodules have yet been found. The middle part consists of light red silty mudstones (Fig. 4A) with scattered small-sized carbonate nodules, suggestive of weakly developed soils. The $\delta^{13}C$ values of these nodules, and also of the ones collected just above the Yellowish Soils at Berganuy (Fig. 5B), vary between –9‰ and –7‰, recording the return to background conditions after the CIE. At Esplugafreda, the lowermost *Alveolina* limestones and marls, with marine molluscs and echinoids, overlie these light-red mudstones (Figs. 4A and 5A).

Tremp section

Located along the road C-1311, from about 1 km east of Claret to the Montllobat Pass, this long section encompasses the upper part of the Tremp Group "Garumnian" succession and the whole Ilerdian Stage (Luterbacher, 1973; Molina et al., 1992, 1995). The present paper, however, is only concerned with the lower part of the section (Figs. 4B and 5C; UTM coordinates: base, X= 324471; Y= 4669476; top, X= 322890; Y= 4669327).

The pre-CIE interval of the studied segment is about 100 m thick, and is mostly represented by brown and red mudstones, which are generally criss-crossed by gypsum veins a few millimetres thick. However, at several levels, the red mudstones are devoid of gypsum and contain well-developed pedogenic carbonate nodules. The interval also comprises a number of channel-like bodies, mostly composed of cross-bedded conglomerates and calcarenites. The thickest and more extensive of these bodies, which occurs near the top of the pre-CIE interval, is up to 30 m thick (Fig. 5C) and at least 1.6 km in lateral extent. In most previous description of the section this outsized clastic unit and the overlying Claret Conglomerate have been merged together (e.g., the ivf -incised valley fill- of Eichenseer and Luterbacher 1992, see Fig. 2). However, as shown in Fig. 5C, both units are separated by a 5 m-thick interval of red mudstones with carbonate nodules. These nodules gave a $\delta^{13}C$ value of ca.–7.5‰, very similar to the values obtained from carbonate nodules collected just below the outsized clastic body (Fig. 5C). There can be no doubt, therefore, that this part of the succession predates the onset of the CIE. Interestingly, in a patch of dark grey mudstones intercalated within the outsized clastic body, López-Martínez and Peláez-Campomanes (1999) found a mammal assemblage attributable to the late Cernaysian (= latest Thanetian) (Fig. 5B, their Claret-4 site). Their finding is a further proof of the Paleocene age of the outsized clastic unit, and also demonstrates that no important hiatus exists between the pre-CIE and syn-CIE intervals in this section.

The syn-CIE interval at the Tremp section is thicker, but otherwise very similar, to that of Esplugafreda, being also composed of the Claret Conglomerate and the Yellowish Soils (Figs. 4B and 5C). The Claret Conglomerate, on which the village of Claret is directly built (Fig. 4B), is up to 4 m thick and is mainly made up of clast-supported conglomerates, with clasts up to 30 cm in diameter. The lower boundary of the unit, best seen just to the east of the village, is a sharp, although essentially flat, erosional surface above the pre-CIE red mudstones. However, the transition to the overlying Yellowish Soils is gradual, being marked by several channel-like, fining-up sequences of conglomerates and cross-bedded calcarenites (Fig. 5B). The bulk of the Yellowish Soils, however, is composed of yellowish-orange silty mudstones with abundant, small-sized carbonate nodules evenly dispersed throughout the unit. The $\delta^{13}C$ values of these nodules vary between –14‰ and –12‰, typical of peak CIE conditions (Fig. 5B). This part of the succession is poorly exposed in the road section itself, where it is cultivated (Fig. 4B). Excellent exposures, however, can be found in the ravines situated to the south and north of the road section (e.g., in the Tendrui section of Schmitz and Pujalte, 2003; UTM coordinates: base, X= 322276; Y= 4670680; top, X= 322322; Y= 4670757)

The post-CIE interval at Tremp includes 3 successive units (Fig. 5B). Unit 1 coincides with the "gypscrete" horizon of Eichenseer and Luterbacher (1992; see Fig. 2). It is about 10 m thick and it is characterized by abundant gypsum deposits, which appear mainly as large-scale coalescing nodules encased within red mudstones. This character, and its position just above the Yellowish Soils, demonstrate that unit 1 correlates with the post-CIE gypsum-rich deep-red marlstones at Esplugafreda (Figs. 4A and 5A-C). Unit 2, the topmost non-marine unit of the "Garumnian" succession in this section, is 5 m thick and is composed of light brown silty mudstones with intercalated channel-like cross-bedded calcarenites. The mudstones are devoid of carbonate nodules, an indication of very immature soil development. The lower 6 m of unit 3 mainly consists of dark-grey low-calcareous mudstones with abundant plant remains, but also includes discontinuous levels of limestone lenses10-15 cm thick and 20-25 cm across (Fig. 5C). Some of these limestone lenses display a septarian structure, described in detail by García Veigas (1988), but others contain Ilerdian alveolinids. Probably because of that, Eichenseer and Luterbacher (1992) placed the lower limit of the Ilerdian Stage at the base of these mudstones (Fig. 2). The remainder of unit 3 is about 80 m thick and is made up of an alternation of grey marls, marly limestones and sandstones with abundant alveolinids of the *A. vredenburgi* Zone.

Campo section

The P-E segment of this classical section crops out in the western slope of the Esera river valley, along a now abandoned tract of the road from Campo to Ainsa (Fig. 4C; UTM coordinates: base, X= 285732; Y= 4696492; top, X= 285653; Y= 4696424). That segment is discussed at length by Pujalte et al. (this issue), and therefore only a brief summary is given here. Additional information can be found in other recent papers dealing with the section, such as Molina et al., (2000, 2003), Payros et al., (2000), Orue-Etxebarria et al., (2001), Pujalte et al., (2003b), Robador (2005) and Scheibner et al., (2007).

The pre-CIE interval of the P-E segment at Campo comprises the upper Thanetian shallow-marine carbonates and the lower 6 m of the overlying non-marine deposits (units 1 and 2 of Pujalte et al., this issue). The abrupt surface with *Microcodium* remains that separates these units has been considered by some authors to imply an important hiatus (e.g., Serra-Kiel et al., 1994; see Fig. 3B). That possibility, however, was refuted by Pujalte et al. (2003b) and, in any case, that surface is older than the CIE and thus unrelated with the event.

The onset of the CIE can now be confidently placed at the base of the channel-like body of cross-bedded pebbly sandstones that occurs in the middle part of the non-marine interval, which is the distal equivalent of the Claret Conglomerate (unit 3 of Pujalte et al., this issue). In addition to that channel-like body, the syn-CIE interval includes the remainder of the non-marine deposits and, at least, the lower limestone beds with Ilerdian alveolinids of the succession (i.e., units 3, 4, 5 and 6 of Pujalte et al., this issue). The top of the syn-CIE interval cannot be precisely located at Campo, but there is little doubt that it occurs within the early Ilerdian limestones, which make up the remainder of the post-CIE interval.

Ermua section

This section is located along the local road from Ermua to the Aixola reservoir, in the Basque province of Biscay, in the western Pyrenees (Fig. 4D; UTM coordinates: base, X= 540906; Y= 4780923; top, X= 540707; Y= 4780864). Orue-Etxebarria et al. (1996) produced an integrated report of the section that included a sequence stratigraphic interpretation and biostratigraphic zonations based on planktonic and small benthic foraminifera, palynomorphs, calcareous nannoplankton and, more important for the purposes of this paper, larger foraminifera. Schmitz et al. (2001) provided a high-resolution isotopic profile and a bed-by-bed litholog across a 65 m thick interval spanning the P-E boundary, as well as grain-size analysis of several samples. The description below is based on these two papers.

The Ermua section is mostly composed of hemipelagic marly limestones alternating with numerous gravity flow deposits, mainly calciturbidites but also slumps and calcidebrites. In addition, the section includes two key lithological units, both widespread in the Basque basin, that Schmitz et al., (2001) respectively named the "Greenish Limestone" and the "Siliciclastic Unit" (Fig. 6A). The former is only 50-70 cm thick in basinal sections of the Basque Basin (e.g., Zumaia, Figs. 1A and 6B), where it is formed almost exclusively of pure limestones of a characteristic greenish colouring due to a high content in glauconite. At Ermua, however, the Greenish Limestone contains about a dozen calciturbidites and therefore it reaches ca. 2m in thickness (Fig. 6A). Yet, when the cumulative thickness of the turbidites is subtracted, the limestone part at Ermua is also about 70 cm thick (Fig. 6B; Schmitz et al., 2001). The Siliciclastic Unit is 20.5 m thick at Ermua and mostly consists of calcite-depleted, fine-grained claystones, with intercalations of thin-bedded calciturbidites. The Greenish Limestone and the Siliciclastic Unit are separated at Ermua by about 1 m of alternating marly limestones and calciturbidites (Fig. 6A), but the cumulative thickness of the marly part only amounts to ca. 30 cm, very similar to the thickness of the coeval interval at Zumaia (Fig. 6B).

One hundred and fifty whole-rock samples were analysed for $\delta^{13}C$ (and $\delta^{18}O$) in the isotopic study of Schmitz et al., (2001). Sampling resolution varied from about 10 samples per metre in a segment just below the Siliciciclatic Unit to one sample every second metre in other parts of the section. Their data show that throughout the lower part of the section, including the Greenish Limestone, most $\delta^{13}C$ values vary between 0‰ and +2‰, which are typical pre-CIE values for marine carbonates (Fig. 6A; see also Fig. 3 of Schmitz et al., 2001). The onset of the CIE is recorded in the marly limestones intercalated between the Greenish Limestone and the Siliciclastic Unit by a seemingly gradual decline in $\delta^{13}C$ values, from 0‰ in the uppermost part of the Greenish Limestone to –4,5‰ in the uppermost part of the marly interval (Fig. 6A). Most samples from the overlying Siliciclastic Unit were too low in carbonate to allow reliable isotopic analyses, but in the few that produced sufficient CO_2 the $\delta^{13}C$ values remained very low (ca. -6‰ to -4‰; Schmitz et al., 2001), while in the overlying succession values gradually become positive (Schmitz et al., 2001, their Fig. 3). This indicates that the Siliciclastic Unit accumulated during the CIE (Fig. 4D), a conclusion reinforced by planktonic foraminifera and calcareous nannoplankton data from Ermua (Orue-Etxebarria et al., 1996), and by correlation with Zumaia (Schmitz et al., 2001).

Larger foraminifera at Ermua occur only within the gravity flow deposits, and are thus allochthonous. For

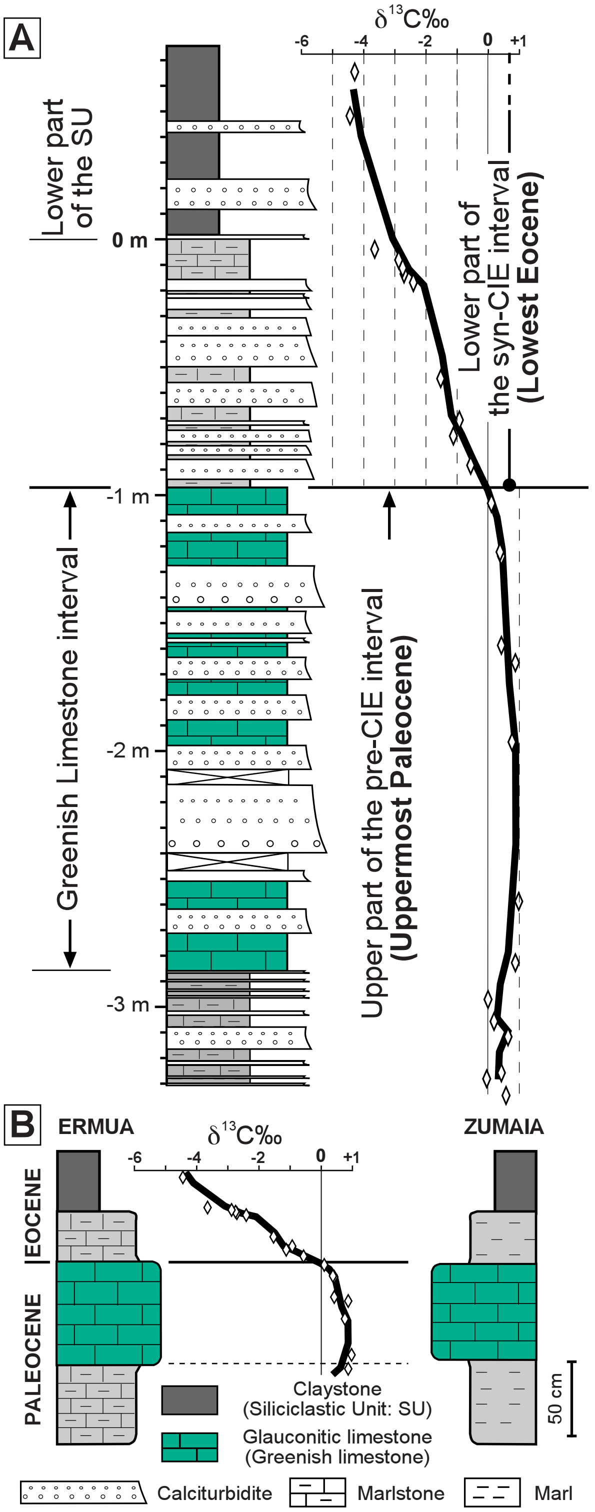

FIGURE 6 **A) High-resolution columnar and isotopic profile of the Ermua section across the P-E boundary; the high number of calciturbidites obscures the rapid onset of the CIE B) Same as above, after removal of the turbiditic beds, and comparison with the P-E boundary interval at Zumaia (modified after Schmitz et al., 2001, their Figs. 4B and 6).**

that reason, care must be taken when using them for biostratigraphic zonation. Larger foraminifera included in clasts of calcidebrites or of coarse-grained calciturbidites have obviously been derived from older, indurated shallow-marine limestones, and are thus biostratigraphically meaningless. In the calciturbidites, however, larger foraminifera abound as individual grains, which must have originated from unconsolidated shallow-water sediments. It is logical to assume that at least a sizable fraction of these larger foraminifera were brought into the base-of-slope apron soon after their death and, therefore, that the stratigraphically youngest specimens of the assemblages can be used for the zonation of the section.

Calciturbidite samples from the pre-CIE interval of the Ermua section contain *Assilina* cf. *azilenzis* and *A. yvettae*, which are typical fossils of the late Thanetian Shallow Benthic Zone SBZ-4 of Serra-Kiel et al., (1998) and *Miscellanea juliettae, Nummulites heberti* and *Ranikothalia sindensis*, which are species of the early Thanetian SBZ-3 of Serra-Kiel et al., (1998). However, not a single specimen of the SBZ-5 was found. Therefore, the pre-CIE interval of Ermua was assigned to the late Thanetian SBZ-4 by Orue-Etxebarria et al., (1996).

Instead, the youngest specimens in the calciturbidites from the lower part of the syn-CIE interval and from the Siliciclastic Unit are: *Nummulites* cf. *gamardensis*, *N.* cf. *solitarius*, *N.* cf. *fraasi*, *Assilina* cf. *ornata*, *A.* cf. *ammonea tectosaga* and *A.* cf. *dandotica*, all of them belonging to the earliest Ilerdian SBZ-5 of Serra-Kiel et al., (1998). This is an additional proof that, within the limit of biostratigraphic resolution, the early Ilerdian turnover of larger foraminifera was coeval with the onset of the CIE.

DISCUSSION

Redefinition of the regional Ilerdian Stage

Two different kinds of stratotypes have historically been used for the definition of chronostratigraphic units, unit stratotypes and boundary stratotypes. A unit stratotype is an actual stratigraphic section or a composite section of two or more nearby sections with a designated base and top (Hedberg, 1976; Salvador, 1994). A boundary stratotype, the so-called Global Standard Stratotype-section and Point (GSSP), defines only the lower limit of a stratigraphic unit, which becomes automatically the upper boundary of the underlying unit. The International Commission on Stratigraphy currently favours the use of boundary stratotypes for the (re)definition of global chronostratigraphic units (Remane et al., 1996).

The definition of the regional Ilerdian Stage by Hottinger and Schaub (1960), however, was based on the designation of two unit stratotypes, Tremp and Campo (see above). Hottinger and Schaub (1960) were in effect mostly concerned with the internal zonation of the stage, within which they recognized five successive larger foraminifera biozones, and not so much with the exact placement of the lower boundary of the stage. This is clearly evidenced, in approximate translation, in the following two excerpts from Schaub (1969; in French in the original): 1, *It is true that in the sections of the Tremp Basin no strata with marine faunas exist at the base or at the top of the Ilerdian. But this criticism does not touch its definition,* ***since the stage is defined by its constituent biozones*** (p. 260, emphasis added); and, 2: *The Campo section demonstrates the intercalation of the Ilerdian between the Paleocene with* A. primaeva *and* A. levis *below and the Cuisian with* N. planulatus *and* A. oblonga *above* (p. 262).

It is not surprising, therefore, that later authors placed the base of the stage at somewhat different levels. Eichenseer and Luterbacher (1992), for instance, chose the base of the post-CIE unit 3 at Tremp (Figs. 2 and Fig. 5B). Payros et al., (2000) and Orue-Etxebarria et al., (2001), on the other hand, selected the abrupt surface with *Microcodium* remains capping the Thanetian marine carbonates at Campo. Interestingly, in the geological map of the Tremp section, Rosell (1994) drew the base of the Ilerdian at a laterally continuous "*level of sandstone and conglomerates up to 8 m thick*" (his unit 53), which is none other than the Claret Conglomerate of Schmitz and Pujalte (2007).

A great majority of authors, however, customarily situate the base of the stage at the lowest bed of the succession with Ilerdian larger foraminifera (e.g., Luterbacher, 1973; Plaziat, 1975; Tambareau and Villatte, 1974; De Renzi, 1996; Molina et al., 2000, 2003; Serra-Kiel et al., 1994). Further, most authors implicitly or explicitly assume that these marine beds are basin-wide coeval or near coeval (Fig. 3). Until recently, such assumption could neither be proved nor disproved, since the biozones of larger foraminifera lack the required temporal resolution. Yet, the discovery of the CIE has provided the necessary tool to reassess this issue.

The data from the Ermua section demonstrate that the appearance of the Ilerdian larger foraminifera is coeval with the onset of the CIE (Fig 7A; see also Pujalte et al., this issue, and Scheibner and Speijer, this issue). However, the lowest bed with Ilerdian alveolinids is placed at Campo in the middle or upper part of the CIE, whereas at Tremp and Esplugafreda it is situated above the top of the isotopic event (Figs. 4 and 7A). In order to constrain the extent of the inherent diachronism, a tentative age model has been constructed in Figs. 7B and 7C, for which two main assumptions have been made: (1) a duration of 200 ky for the CIE, since recent estimates of the length of the event range between 170 ky (Röhl et al., 2007) and 231 ky (Giusberti et al., 2007); (2) that sedimentation rates in the alluvial coastal plains of the Tremp-Graus basin did not change appreciably during and after the CIE. According to that model, the age of the lowest marine Ilerdian bed at Campo, Tremp and Esplugafreda are respectively about 125 ky, 250 ky and 400 ky younger than at Ermua (Figs. 7B and C). Such diachronism is clearly related to the eastwards advance of the Ilerdian transgression on the former alluvial plains of the Tremp basin and, in all likelihood, it will steadily increase in more eastern sections of the Pyrenees. This is not acceptable for a chronostratigraphic unit, the age of which must be isochronous.

It is clear that, although unintended, the term "Ilerdian" has often been employed as a synonym of an early Paleogene marine facies, not unlike the term "Garumnian" designates Upper Cretaceous-Lower Paleogene non-marine deposits (Fig. 7B). Yet, this was not the purpose of Hottinger and Schaub (1960), who explicitly proposed the Ilerdian as a chonostratigraphic unit, nor that of the researches that later have used the term. As a way out of this dilemma we here propose to redefine the Ilerdian Stage following a similar procedure to the one employed by the International Commission on Stratigraphy to establish global chronostratigraphic standards: namely, by inserting a "spike" that we will call "silver spike" to differentiate it from the golden ones used to mark GSSPs. We further propose to place such silver spike at the base of the Claret Conglomerate in the Tremp section (Fig. 7A), so that the historical Ilerdian stratotype maintains its relevance.

Comparing global and regional Early Eocene Stages

In the afternoon of February 8, 2004, the GSSP for the P-E boundary was officially designated at the base of the syn-CIE interval at the Dababiya Quarry section (Egypt), thus concluding several years of debate about the selection of this important boundary (Aubry et al., 2007). According to the International Commission on Stratigraphy, such designation entailed placing the base of the global Ypresian Stage at the revised P-E boundary (Fig. 8; Gradstein et al., 2004); this decision, however, was considered inappropriate by some members of the P-E Working Group (e.g., Aubry et al., 2003; Thiry et al., 2006), arguing that such procedure involved lowering the boundary of the regional Ypresian Stage as defined by its stratotype in Belgium. It is far beyond the scope of this paper to discuss the pros and cons of these alternatives. Instead, we will just mention the merits for lowering the base of the regional Ilerdian Stage.

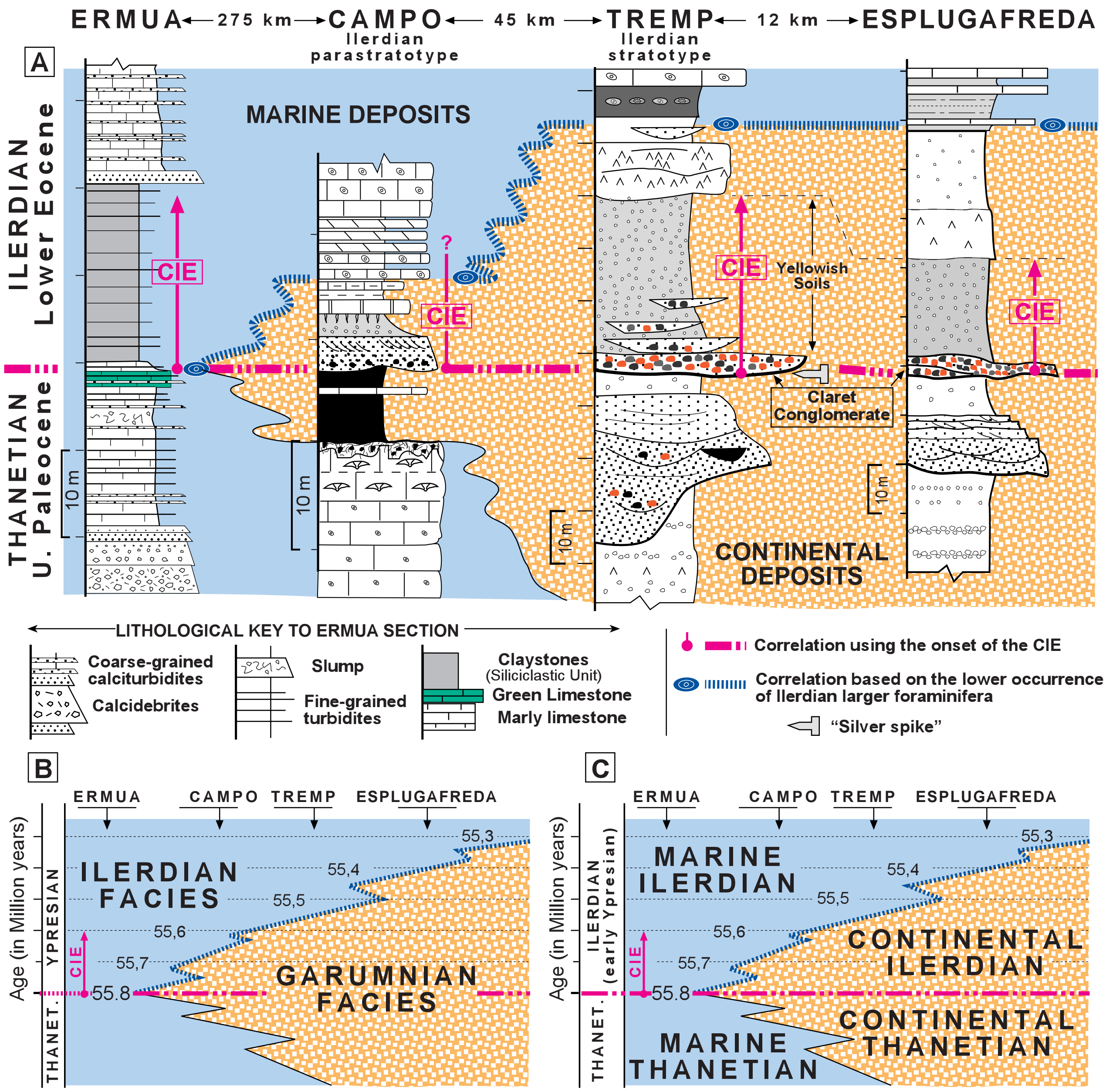

FIGURE 7 | **A) Alternative correlation of the four sections described in the paper based on the two different criteria discussed in the text. See Figs. 2 and 5 for key to symbols and lithologies B) and C): Tentative chronostratigraphic schemes across the P-E boundary, showing the two possible meanings of the Ilerdian: a facies name or a chronostratigraphic unit.**

The correlation between the global and regional early Eocene stages, as respectively favoured by the International Commission of Stratigraphy and proposed in this paper, is shown in Fig. 8. In the resulting scheme, the time span of the global Ypresian becomes very similar to the joint time span of the redefined regional Ilerdian Stage plus the Cuisian Stage, as defined in the Pyrenees by Schaub (1992). As discussed above, the redefinition proposed in this paper implies a lowering of the base of the Ilerdian Stage at its stratotype (Tremp Section). However, we support that lowering for two main reasons: (1) the biostratigraphic criteria originally used to define the Ilerdian Stage (i.e., the development of the well-known larger porcellaneous orbitolitids, alveolinids, lacazinids and lamellar perforate involute nummulitids) should take preference to the position of the base of the stage at its type section. Indeed, the election of the Tremp section as stratotype can be regarded, in perspective, as a mere accident: had it been placed in a more western or eastern location the base of the stage would have been situated,

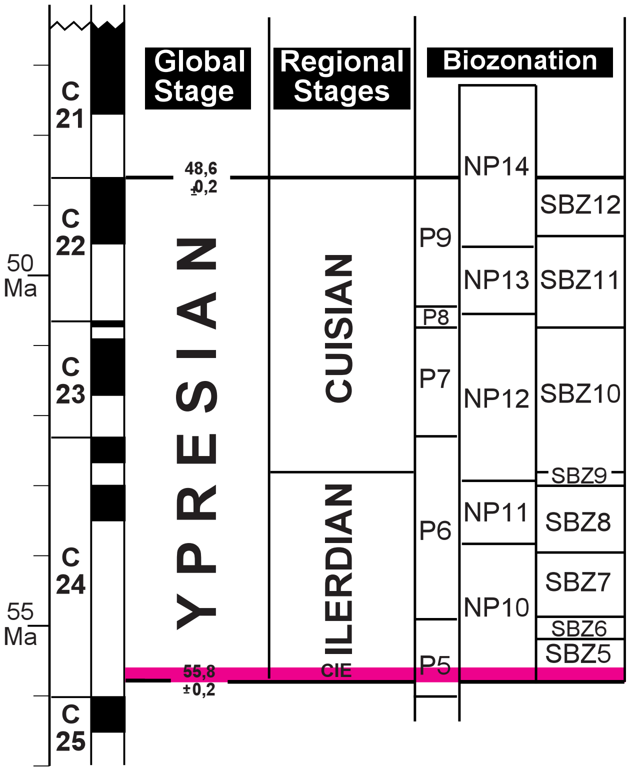

FIGURE 8 | **Correlation of Early Eocene Stages: the global Ypresian Stage (as currently accepted by the International Commission on Stratigraphy, (Gradstein et al., 2004) and the regional Ilerdian and Cuisian Stages, after lowering the base of the Ilerdian as proposed in this paper. P and NP, planktonic foramiminifera and calcareous nannoplankton zones, after Berggren et al., 1995; SBZ, shallow benthic zones, after Serra-Kiel et al., 1998, slightly modified.**

respectively, at a stratigraphycally older or younger position than at Tremp (see Figs. 7B and C). (2) Lowering of the base of the Ilerdian will enable a friendly coexistence and an easy correlation between the global Ypresian (*sensu* International Commission of Stratigraphy) and the regional stages of the Early Eocene (Fig. 8). It is worth mentioning that Hottinger and Schaub (1960) and Schaub (1973, 1992) recognized 5 successive larger foraminiferal biozones within the Ilerdian Stage, Tosquella et al., (1996, publ. 1998) revised and improved the precision of nummulitid biostratigraphic scales in the Pyrenean basin, and Serra-Kiel et al., (1998) codified these zones as SBZ-5 to SBZ-9, whereas only two planktonic foraminiferal and calcareous plankton biozones are recognized for the same time interval (Fig. 8). Therefore, in the particular lapse of geological time discussed here, larger foraminifera offer a better biostratigraphic resolution than calcareous plankton, at least theoretically. This fact, together with the scarcity or even absence of calcareous plankton in many shallow water sequences of the Tethys domain, is probably the main reason explaining the persistence in the usage of the Ilerdian Stage.

CONCLUSIONS

The Ilerdian represents an important time interval in the evolution of the larger foraminifera, and has been used by most researchers working in shallow marine facies of the Tethys from the Pyrenees to Pakistan. In the Pyrenees at least, most maps and publications dealing with lower Paleogene strata, including recent ones, do not use the global Ypresian Stage, but instead the regional Ilerdian and Cuisian Stages (e.g., Rasser et al., 2005; Scheibner et al., 2007; Robador and Zamorano, in press). It seems unlikely that this deep-rooted practice will be abandoned in the foreseeable future, if ever. The confirmation that the appearance of Ilerdian larger foraminifera was coeval with the onset of the CIE, and consequently with the newly defined P-E boundary, has further reinforced the usefulness of the stage.

We have demonstrated, however, that the criterion most frequently employed to place the base of the Ilerdian Stage in transitional continental-marine successions (i.e., the lowest occurrence of marine beds with "Ilerdian" larger foraminifera) is inadequate and, in practice, it has led to the usage of the Ilerdian as a shallow marine lower Eocene facies rather than as a chronostratigraphic unit, as was originally intended. To overcome that problem, we propose to redefine the Ilerdian with a silver spike, to be placed at the onset of the CIE in the Tremp section, the original stratotype of the stage. In fully marine successions, correlation with this silver spike will be possible either with the CIE or with the several biotic turnovers associated with the event, including the LFT. However, in continental or transitional continental-marine successions such as the ones that characterize the Tremp-Graus Basin (the type area of the stage), the primary means of correlation will have to be the onset of the CIE.

The base of the Eocene series is now officially defined with a GSSP situated at the onset of the CIE in the Dababiya Quarry section of Egypt (Aubry et al., 2007). Therefore, the bases of both the global Ypresian Stage (as accepted by the International Commission of Stratigraphy) and the regional Ilerdian Stage (as here redefined) can be considered coeval for all useful purposes. This fact greatly simplifies the chronostratigraphy of the Early Eocene and permits the coexistence of both stages.

ACKNOWLEDGEMENTS

Funds for this research for Spanish researches were provided by the projects CGL2005-01721/BTE, CGL2005-02770/BTE and CGL2005-00404/BTE and from the Basque Government Research Group Grant GIC07/122-IT-215-07. Funds for B. Schmitz were provided by the Swedish Research Council. The manuscript has benefited greatly from constructive reviews by H. Luterbacher, E. Molina and an anonymous referee. This paper is dedicated to the memory of Yvette Tambareau, a keen researcher of the early Paleogene of the Pyrenees, and a kind and pleasant friend of all of us.

REFERENCES

Aubry, M.P., Ouda, K., Dupuis, C., Berggren, W. A.,Van Couvering, J. A., and the Members of the Working Group on the Paleocene/Eocene boundary, 2007. The Global Standard Stratotype-section and Point (GSSP) for the base of the Eocene Series in the Dababiya section (Egypt). Episodes, 30(4), 271-286.

Aubry, M.-P., Berggren, W.A.,Vancouvering, J.A., Ali. J., Brinkuis, H., Cramer, B., Kent, D.K., Swisher III, C.C., Dupuis, C., Gingerich, P.D., Heilmann-Clausen, C., King, C., Ward, D.J., Knox, R. W. O´B., Ouda, K., Stott, L.D., Thiry, M., 2003. Chronostratigraphic terminology at the Paleocene/ Eocene boundary. In: Wing, S.L., Gingerich, P.D., Schmitz, B. and Thomas, E., (eds.). Causes and Consequences of Globally Warm Climates in the Early Paleogene. Geological Society of America Special Paper 369, 551-566.

Baceta, J.I., 1996. El Maastrichtiense superior, Paleoceno e Ilerdiense inferior de la Región Vasco-Cantábrica: secuencias deposicionales, facies y evolución paleogeográfica. Doctoral thesis. Universidad del País Vasco, 372 pp.

Baceta, J.I., Pujalte, V., Serra-Kiel, J., Robador, A., Orue-Etxebarría, X., 2004. El Maastrichtiense final, Paleoceno e Ilerdiense inferior de la Cordillera Pirenaica. In: Vera, J. A. (ed.) Geología de España, Sociedad Geológica de España - Instituto Geológico y Minero de España, 308-313.

Baceta J.I., Pujalte V., Bernaola G., 2005. Paleocene coralgal reefs of the Western Pyrenean basin, northern Spain: New evidence supporting an earliest Paleogene recovery of reefal ecosystems. Palaeogeography, Palaeoclimatology, Palaeoecology, 224, 117-143.

Bains, S., Corfield, R.M., Norris, R.D., 1999. Mechanisms of climate warming at the end of the Paleocene. Science, 285, 724–727.

Bains, S., Norris, R.D., Corfield, R.M., Bowen, G.J., Gingerich P.D., Koch, P.L., 2003. Marine-terrestrial linkages at the Paleocene–Eocene boundary. In: Wing, S.L., Gingerich, P.D., Schmitz, B., Thomas, E. (eds.). Causes and Consequences of Globally Warm Climates in the Early Paleogene. Geological Society of America Special Paper, 369, 49-64.

Berggren, W.A., Kent, D.V., Swisher III, C.C., Aubry, M.P., 1995. A revised Cenozoic geochronology and chronostratigraphy. In: Berggren, W.A., Kent, D.V., Aubry, M.P., Hardenbol, J. (Eds.). Geochronology, Time Scales and Global Stratigraphic Correlations. Society for Sedimentary Geology (SEPM) Spec. Publ. 54, 129-212.

Boukhary, M., Bassiouni, M. E. A., Hussein-Kamel, Y., 1995. Nummulites Luterbacheri, unexpected large nummulites from basal Ilerdian (Early Eocene) from El Quss Abu Said, Farafra oasis, Western Desert, Egypt. Revue de Micropaléontologie, 38, 285-298.

Bowen, G.J., Koch, P.L., Gingerich, P.D., Norris, R.D., Bains, S., Corfield, R.M., 2001. Refined isotope stratigraphy across the continental Paleocene–Eocene boundary on Polecat Bench in the northern Bighorn Basin. In: Gingerich, P.D. (ed.). Paleocene–Eocene Stratigraphy and Biotic Change in the Bighorn and Clarks Fork Basins, Wyoming. University of Michigan Papers on Paleontolology, 33, 73-88.

Bowen, G.J., Clyde, W.C., Koch, P.L., Ting, S., Alroy, J., Tsubamoto, T., Wang, Yu., Wang, Y., 2002. Mammalian Dispersal at the Paleocene/Eocene Boundary. Science, 295, 2062-2065.

Cerling, T.E., 1984. The stable isotopic composition of modern soil carbonate and its relationship to climate. Earth and Planetary Science Letters, 71, 229-240

Colakoglu, S., Ozcan, E., 2003. Orthophragminid foraminiferal assemblages from an Ilerdian-early Cuisian reference section (Sakarya section, Haymana-Polatli Basin, central Anatolia-Turkey). Revista Italiana di Paleontologia e Stratigrafia, 109 (1), 45-61.

Cuevas J. L., 1992. Estratigrafía del “Garumniense” de la Conca de Tremp. Prepirineo de Lérida. Acta Geologica Hispanica, 27(1-2), 95-108.

De Renzi, M., 1996. La influencia de los factores tafonómicos y paleoecológicos en la distribución de los moluscos en el área tipo del Ilerdiense (Conca de Tremp, Cataluña, España). Revista Española de Paleontología, nº extraordinario, 204-214.

Dickens, G.R., Castillo, M.M., Walker, J.C.G., 1997. A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate. Geology 25(3), 259-262.

Eichenseer, H., Luterbacher, H., 1992. The marine Paleogene of the Tremp Region (NE Spain) - depositional sequences, facies history, biostratigraphy and controlling factors. Facies, 27, 119–152.

Freytet P., Plaziat J.C., 1970. New elements of lithologic and paleontologic correlation between marine and continental formations of early Eocene in Languedoc Region – Problem of continental Ilerdian. Comptes Rendus des Seances de l’Academie des Sciences, Paris, 270(8), 1076-1079.

García Veigas, F. J., 1988. La formación evaporítica de las facies garumnienses de la Conca de Tremp (Lleida): petrología, geoquímica y sedimentología. Tesis de Licenciatura, Universitat Autónoma de Barcelona, 175 pp.

Giusberti, L., Rio, D., Agnini, C., Backman, J., Fornaciari, E., Tateo, F., Oddone, M., 2007. Mode and tempo of the Paleocene-Eocene thermal maximum in an expanded section from the Venetian pre-Alps. Geological Society of America Bulletin 119 (3/4), 391-412.

Gradstein, F.M., Ogg, J.G., Smith, A.G., Bleeker, W., Lourens, L.J., 2004. A new geologic time scale with special reference to the Precambrian and Neogene. Episodes, 27, 83-100.

Hedberg, H.D. (ed.), 1976. International stratigraphic guide. A guide to stratigraphic classification, terminology and procedure. (International Subcommittee on Stratigraphic Classification, IUGS Commission on Stratigraphy). New York, John Wiley and Sons, 200 pp.

Hottinger, L., 1998. Shallow benthic foraminifera at the Paleocene-Eocene boundary. Strata, 9, 61-64.

Hottinger, L., Schaub, H., 1960. Zur Stufeneinteilung das Paleocaens uns das Eocaens. Einführung der Stufen Ilerdien und Biarritzien. Eclogae Geologicae Helvetiae, 53(1), 453-479.

Institut Cartogràfic de Catalunya, 2002. Mapa geologic de Catalunya. Escala 1:250 000.

Koch, P.L., Zachos, J.C., Dettman, D. L., 1995. Stable isotope stratigraphy and paleoclimatology of the Paleogene Bighorn Basin (Wyoming, USA). Palaeogeography, Palaeoclimatology, Palaeoecology, 115, 61-89.

Koch, P.L., Clyde, W.C., Hepple, R.P., Fogel, M.L., Wing, S.L., Zachos, J.C., 2003. Carbon and oxygen isotope records from paleosols spanning the Paleocene–Eocene boundary, Bighorn Basin, Wyoming. In: Wing, S.L., Gingerich, P.D., Schmitz, B., Thomas, E. (eds.). Causes and Consequences of Globally Warm Climates in the Early Paleogene. Geological Society of America Special Paper, 369, 49-64.

López-Martínez, N., Peláez-Campomanes, P., 1999. New mammals from south-central Pyrenees (Tremp Formation, Spain) and their bearing on late Paleocene marine-continental correlations. Bulletin de la Société Géologique de France, 170(5), 681-696.

Luterbacher, H.P., 1973. La sección tipo del Piso Ilerdiense. In: Enadimsa (ed.). Libro-Guía del XIII Coloquio Europeo de Micropaleontología (España, 1973), 101-128.

Luterbacher, H.P., Hardenbol, J., Schmitz, B., 2000. Decision of the voting members of the International Subcommission on Paleogene Stratigraphy on the criterion for the recognition of the Paleocene/Eocene boundary. Newsletter of the International Subcommission on Paleogene Stratigraphy, 9, 13.

Luterbacher, H.P., Ali, J.R., Brinkhuis, H., Gradstein, F.M., Hooker, J.J., Monechi, S., Ogg, J.G., Powell, J., Röhl, U., Sanfilippo, A., Schmitz, B., 2004. The Paleogene Period. In: Gradstein, F.M., Ogg J., Smith, A. (eds.). A Geological Timescale 2004. Cambridge, Cambridge University Press, 384-408.

Médus, J., Colombo, F., 1991. Succession climatique et limite stratigraphique Crétacé-Tertiaire dans le N.E. de l'Espagne. Acta Geologica Hispanica, 26, 173-179.

Molina, E., Canudo, J.I., Guernet, C., McDougall, K., Ortiz, N., Pascual, J.O., Parés, J.O., Samsó, J.M., Serra-Kiel, J., Tosquella, J., 1992. The stratotypic Ilerdian revisited: Integrated stratigraphy across the Paleocene/Eocene boundary. Revue de Micropaléontologie, 35, 143-156.

Molina, E., De Renzi, M., Álvarez, G., 1995. El estratotipo del Ilerdiense y su registro fósil. In: G. López et al. (eds.). Actas de las XI Jornadas de Paleontología, 211-220.

Molina, E., Angori, E., Arenillas, I., Monechi, S., Schmitz, B., 2000. Integrated stratigraphy across the Paleocene/Eocene boundary at Campo, Spain. GFF- Geologiska Föreningens i Stockholm Förhandlingar, 122(1), 106-107.

Molina, E., Angori, E., Arenillas, I., Brinkhuis, H., Crouch, E.M., Luterbacher, H., Monechi, S., Schmitz, B., 2003. Correlation between the Paleocene/Eocene boundary and the Ilerdian at Campo, Spain. Revue de Micropaléontologie, 46(2), 95-109.

Murru, M., Ferrera, C., Da Pelo, S; Ibba, A., 2003. The Palaeocene-Middle Eocene deposits of Sardinia (Italy) and their palaeoclimatic significance. Comptes Rendus Geoscience, 335 (2): 227-238.

Orue-Etxebarria, X., Apellaniz, E., Baceta, J.I., Coccioni, R., Di Leo, R., Dinarès-Turell, J., Galeotti, S., Monechi, S., Núñez-Betelu, K., Parés, J.M., Payros, A., Pujalte, V., Samsó, J.M., Serra-Kiel, J., Schmitz, B., Tosquella, J., 1996. Physical and biostratigraphic analysis of two prospective Paleocene–Eocene Boundary stratotypes in the intermediate-deep water Basque Basin, western Pyrenees: The Trabakua Pass and Ermua sections. Neues Jahrbuch für Geologie und Paläontologie Abhandlungen, 201, 179-242.

Orue-Etxebarria, X., Pujalte, V., Bernaola, G., Apellaniz, E., Baceta, J.I., Payros, A., Núñez-Betelu, K., Serra-Kiel, J., Tosquella, J., 2001. Did the Late Paleocene thermal maximum affect the evolution of larger foraminifers? Evidence from calcareous plakton of the Campo Section (Pyrenees, Spain). Marine Micropaleontology, 41, 45-71.

Ozgen-Erdem, N., Akyazi, M., Karabasoglu, A., 2007. Biostratigraphic interpretation and systematics of Alveolina assemblages from the Ilerdian-Cuisian limestones of Southern Eskisehir, Central Turkey. Journal of Asian Earth Sciences, 29 (5-6), 911-927

Payros, A., Pujalte, V., Baceta, J.I., Bernaola, G., Orue-Etxebarria, X., Apellaniz, E., Caballero, F., Ferràndez, C., 2000. Lithostratigraphy and sequence stratigraphy of the upper Thanetian to middle Ilerdian strata of the Campo section (southern Pyrenees, Spain): revision and new data. Revista de la Sociedad Geológica de España, 13(2), 213-226.

Plaziat, J.C., 1975. L'Ilerdien à l'intérieur du Paléogène languedocien; ses relations avec le Sparnacien, l'Ilerdien sudpyrénéen, l'Yprésien et le Paléocène. Bulletin de la Société Géologique de France, 17(2), 168-182.

Plaziat, J.C., 1981, Late Cretaceous to late Eocene palaeogeographic evolution of Southwest Europe. Palaeogeography, Palaeoclimatology, Palaeoecology, 36, 263-320.

Pomerol, Ch., 1975. La signification de l'Ilerdien et l'intérêt de cet étage dans la stratigraphie du Paléogène mésogéen. B. S. G. F., XVII, 213-217

Puigdefàbregas, C., Souquet, P., 1986. Tectosedimentary cycles and depositional sequences of the Mesozoic and Tertiary from the Pyrenees. Tectonophysics, 129, 173-203.

Pujalte, V., Orue-Etxebarria, X., Schmitz, B., Tosquella, J., Baceta, J.I., Payros, A., Bernaola, G., Caballero, F., Apellaniz, E., 2003a. Basal Ilerdian (earliest Eocene) turnover of larger foraminifera: age constraints based on calcareous plankton and $\delta^{13}C$ isotopic profiles from new southern Pyrenean sections (Spain). In: Wing, S.L., Gingerich, P.D., Schmitz, B., Thomas, E. (eds.). Causes and Consequences of Globally Warm Climates in the Early Paleogene. Geological Society of America Special Paper, 369, 205-221.

Pujalte, V., Dinarès-Turell, J., Bernaola, G., Baceta, J.I., Payros, A., 2003b. A reappraisal of the position of chron C25n in the Campo section (Huesca province, south-central Pyrenees). Geogaceta, 34, 155-158.

Pujalte, V. Schmitz, B., 2005: Revisión de la estratigrafía del Grupo Tremp («Garumniense», Cuenca de Tremp-Graus, Pirineos meridionales). Geogaceta, 38, 79-82.

Pujalte, V., Schmitz, B., 2006 : Revised lithostratigraphy of the Tremp Group. In: Baceta, J.I., Pujalte, V., and Caballero, F (eds.). Paleocene and Early Eocene facies and events: a Basin-Platform-Coastal Plain Transect (south-central and western Pyrenees). Post-conference field trip Guidebook, CBEP 2006. 63-69.

Pujalte, V., Schmitz, B., Baceta, J.I., Orue-Etxebarria, X., Bernaola, G., Dinarès-Turell, J., Payros, A., Apellaniz, E., Caballero, F., 2009. Correlation of the Thanetian Ilerdian turnover of larger foraminifera and the Paleocene-Eocene thermal maximum: confirming evidence from the Campo area (Pyrenees, Spain). Geologica Acta, 7(1-2), 161-175.

Rasser M.W., Scheibner C., Mutti M., 2005. A paleoenvironmental standard section for Early Ilerdian tropical carbonate factories (Corbieres, France; Pyrenees, Spain). Facies 51(1-4), 227-242.

Remane, J., Bassett, M.G., Cowie, J.W., Gohrbandt, K.H., Lane, H.R., Michelsen, O., Naiwen, W., 1996. Revised guidelines for the establishment of global chronostratigraphic standards by the International Commission of Stratigraphy (ICS). Episodes, 19, 77-81.

Robador, A., Zamorano, M., in press. Mapa Geológico de España. Scale 1:50.000, 2nd series, nº. 212, Campo. Madrid, Instituto Geológico y Minero de España, 1 fold. map.

Robador, A., 2005. El Paleoceno e Ilerdiense inferior del Pirineo occidental: Estratigrafía y sedimentología. Doctoral thesis, Universidad del País Vasco, 456 p.

Röhl, U., Bralower, T.J., Norris, R.D., Wefer, G., 2000. New chronology for the late Paleocene thermal maximum and its environmental implications. Geology, 28, 927-930.

Röhl, U., Westerhold, T., Bralower, T.J., Zachos, J.C., 2007. On the duration of the Paleocene-Eocene thermal maximum (PETM). Geochemistry, Geophysics, Geosystems, 8 (12), 1-13.

Rosell, J., 1994. Mapa Geológico de España. Scale 1:50.000, 2nd series, nº. 252, Tremp. Madrid, Instituto Tecnológico Geominero de España, 63 pp., 1 fold. map.

Salvador, A., (ed.), 1994. International Stratigraphic Guide (2nd edition). A guide to stratigraphic classification, terminology and procedure. Trondheim (Norway), International Union of Geological Sciences and The Geological Society of America, 214 p.

Schaub, H., 1969. L'Ilerdien; état actuel du problème. Colloque sur l'Eocene, Paris, 1968. Mèmoirs Bulletin Recherches Geologiques et Minieres, 69, 259-261.

Schaub, H., 1973. La sección de Campo (Prov. de Huesca). In: Libro-Guía del XIII Coloquio Europeo de Micropaleontología (España, 1973) (ed. Enadimsa), 139-158.

Schaub, H., 1973. La sección de Campo (Prov. de Huesca). In: Libro-Guía del XIII Coloquio Europeo de Micropaleontología (España, 1973) (ed. Enadimsa), 139-158.

Schaub, H.,1992. The Campo section (NE Spain), a Tethyan parastratotype of the Cuisian. Neues Jahrbuch für Geologie und Paläontologie Abhandlungen, 186, 63-70.

Scheibner, C., Rasser, M.W., Mutti, M., 2007. The Campo section (Pyrenees, Spain) revisited: Implications for changing benthic carbonate assemblages across the Paleocene–Eocene boundary. Palaeogeography, Palaeoclimatology, Palaeoecology, 248, 145-168.

Scheibner, C., Speijer, R.P., 2009: Recalibration of the Tethyan shallow-benthic zonation across the Paleocene-Eocene boundary: the Egyptian record. Geologica Acta, 7(1-2), 195-214.

Schmitz, B., Asaro, F., Molina, E., Monechi, S., von Salis, K., Speijer, R.P., 1997. High-resolution iridium, $\delta^{13}C$, $\delta^{18}O$, foraminifera and nannofossil profiles across the latest Paleocene benthic extinction event at Zumaya, Spain. Palaeogeography, Palaeoclimatology, Palaeoecology, 133, 49-68.

Schmitz, B., Pujalte, V., Núñez-Betelu, K., 2001. Climate and sea-level perturbations during the Initial Eocene Thermal Maximum: evidence from siliciclastic units in the Basque Basin (Ermua, Zumaia and Trabakua Pass), northern Spain. Palaeogeography, Palaeoecology, Palaeoclimatology 165, 299-320.

Schmitz, B., Pujalte, V., 2003. Sea-level, humidity, and land-erosion records across the initial Eocene thermal maximum from a continental-marine transect in northern Spain. Geology, 31, 689-692.

Schmitz, B., Pujalte, V., 2007. Abrupt increase in seasonal extreme precipitation at the Paleocene-Eocene boundary. Geology, 35, 215–218.

Serra-Kiel, J., Canudo, J.I., Dinarés, J., Molina, E., Ortiz, N., Pascual, J.O. Samsó, J.M., Tosquella, J., 1994. Cronoestratigrafía de los sedimentos marinos del Terciario inferior de la Cuenca de Graus-Tremp (Zona Central Surpirenaica). Revista Sociedad Geológica de España, 7(3-4), 273-297.

Serra-Kiel, J., Hottinger, L., Caus, E., Drobne, K., Ferràndez, C., Jauhri, A.K., Less, G., Pavlovec, R., Pignatti, J., Samsó, J.M., Schaub, H., Sirel, E., Strougo, A., Tambareau, Y., Tosquella, J., Zakrevskaya, E., 1998. Larger foraminiferal biostratigraphy of the Tethyan Paleocene and Eocene. Bulletin de la Société Géologique de France, 169, 281-299.

Tambareau, Y., 1972. Thanétien supérieur et Ilerdien inférieur des Petites Pyrénées, du Plantaurel et des Chaînons audois. Doctoral thesis. Universite de Toulouse, 377 pp.

Tambareau, Y., Villatte, J., 1974. Le passage Thanétien-Ilerdien dans la región de Campo: comparison avec les Petites Pyrénées. Bulletin Histoire Naturelle de Toulouse 110, 340-361.

Tambareau, Y., Feist, M., Gruas-Cavagnetto, C., Murru M., 1989. Characterization of the continental Ilerdian in the Western Mediterranean area. Comptes Rendus de l'Ácademie des Sciences, 308(7), 689-695.

Thiry, M., Aubry, M-P., Dupuis, C., Sinha, A., Stott, L.D., Berggren, W.A., 2006. The Sparnacian deposits of the Paris Basin: d13C isotope stratigraphy. Stratigraphy, 3, (2), 119-138.

Tosquella, J., Serra-Kiel, J., Ferràndez-Cañadell, C., Samsó, J.M., 1996 (pub. 1998). Las biozonas de nummulítidos del Paleoceno Superior-Eoceno Inferior de la Cuenca Pirenaica. Acta Geologica Hispanica, 31(1-3), 23-36 (with English Abridged version).

Vergés, J., Martínez, A., 1988. Corte compensado del Pirineo oriental: Geometría de las cuencas de antepaís y edades de emplazamiento de los mantos de corrimiento. Acta Geologica Hispanica, 23, 95-106.

Wright, V.P. and Marriott, S. B. 1996. A quantitative approach to soil occurrence in alluvial deposits and its application to the Old Red Sandstone of Britain. Journal of the Geological Society, London, 153, 907-913.

Zachos, J.C., Wara, M.W., Bohaty, S., Delaney, M.L., Petrizzo, M.R., Brill, A., Bralower, T.J., Premoli Silva, I., 2003. A transient rise in tropical sea surface temperature during the Paleocene–Eocene Thermal Maximum. Science, 302, 1551-1554.

**Manuscript received March 2008;
revision accepted May 2008;
published Online November 2008.**