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Correlation of the Thanetian-Ilerdian turnover of larger foraminifera and the Paleocene-Eocene thermal maximum: confirming evidence from the Campo area (Pyrenees, Spain)

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ABSTRACT

It has long been known that a major larger foraminifera turnover (LFT) occurred at the boundary between the Thanetian and Ilerdian stages, but its possible correlation with the Paleocene-Eocene thermal maximum (PETM) was unsuspected until the work of Baceta (1996), and has been controversial ever since. After summarizing the history of this controversy, we present information from three new sections that conclusively resolve the issue, all of them placed less than 2 km to the east of the classical Campo section in the southern Pyrenees. In these three sections, an up to 7 meter-thick intercalation of continental deposits rich in pedogenic carbonate nodules is sandwiched between uppermost Thanetian and lowermost Ilerdian shallow marine carbonates. The $\delta^{13}C$ composition of 42 pedogenic nodules collected from two of these sections (San Martín and La Cinglera) ranges between –11.4 and -14.3‰ and averages –12.9‰, values that conclusively represent the PETM and for the first time are recorded in sections where the LFT is clearly represented. Further, a high-resolution lithological correlation between Campo and the three new sections across the P-E interval unquestionably demonstrates that the lowermost marine beds with autochthonous specimens of *Alveolina vredenburgi* (a tell-tale of the LFT) are laterally interfingered –and are therefore coeval- with the nodule-bearing PETM continental deposits. On the basis of the new evidence, the temporal coincidence of the PETM and the LFT can no longer be doubted.

KEYWORDS **PETM. Paleocene-Eocene boundary. Larger foraminifera. Isotopes. Campo section.**

INTRODUCTION

The Paleocene-Eocene thermal maximum (PETM; also called the initial Eocene thermal maximum, IETM) was a dramatic global environmental perturbation that occurred at the Paleocene–Eocene (P–E) boundary, ca. 55 Ma. The fingerprint of the event is a 2–6‰ negative carbon isotopic excursion (CIE), first identified at ODP Site 690 in the Antarctic Ocean by Stott et al. (1990) and subsequently recognized at many other deep sea sites and land sections (e.g., Schmitz et al., 1996; Bains et al., 1999; Thomas et al., 1999; Zachos et al., 2001, 2003). The CIE developed during less than 10 ky, requiring the addition of large quantities of isotopically light carbon to the atmosphere–ocean reservoir. The resulting greenhouse effect caused exceptionally warm conditions, which lasted for ca. 150 ky.

The origin of this dramatic warming in climate is not yet well understood, the leading hypothesis being a massive dissociation of seafloor methane hydrates (Dickens et al., 1995; Thomas et al., 2002; Svensen et al., 2004). Whatever its cause, it has become increasingly clear that the PETM caused a major ecological impact. In the oceans, deep-sea benthic foraminifera experienced their greatest extinction during the last 90 Ma (Thomas, 1990, 1998; Kennett and Stott, 1991; Speijer et al., 1996; Thomas and Shackleton, 1996), planktonic foraminifera and calcareous nannoplankton experienced transient diversifications (Kelly et al., 1996; Aubry, 1998; Monechi et al., 2000; Kelly, 2002; Raffi et al., 2005; Gibbs et al., 2006), and *Apectodinium* dinoflagellates bloomed worldwide in shelf areas (Crouch et al., 2001). On land, archaic mammals were replaced by modern groups, including the earliest true primates (Clyde and Gingerich, 1998; Bowen et al., 2002; Gingerich, 2003), while the floras experienced important changes (Jaramillo, 2006; Wing et al., 2005; Smith et al., 2007). Not surprisingly, the Paleogene Subcommission of the International Commission on Stratigraphy agreed to use the base of the CIE as the primary correlation criterion for the P-E boundary (Luterbacher et al., 2000, 2004).

Long before that, however, an important turnover of larger foraminifera had been discovered around the P-E boundary in the shallow marine carbonate platforms of the northern Tethys (Schaub, 1951; Hottinger, 1960), and used to define the base of the Ilerdian Stage (Hottinger and Schaub, 1960). It was even suggested that this larger foraminiferal turnover (LFT) was an ideal event to mark the base of the Eocene Epoch (Pomerol, 1975). Yet, it was also generally believed that the LFT was older than the PETM and, because of that, this premonitory suggestion was neither accepted in the seventies, nor reconsidered during later discussions on the placement of the Eocene lower boundary.

Several recent studies (see below) have claimed that the LFT and the PETM were synchronous events and, consequently, that the thermal event also had a great impact on shallow-water marine communities. These studies, however, have so far attracted little attraction, and have even been questioned by some authors.

The main objective of this paper is to provide new evidence that corroborates the correlation between these two dramatic events. However, in order to place the new data in a proper context, we will first discuss at some length the reasons for the earlier miscorrelation between the LFT, the calcareous plankton biostratigraphic scales and the PETM, and also will try to explain the reluctance of some specialists to accept the alternative correlation here conclusively confirmed.

HISTORICAL BACKGROUND

The initial paradigm

The history of the LFT-PETM controversy is yet another corroboration of Kuhn's theory about scientific progress. As it is well known, Kuhn (1970) proposed that, rather than in a steady and cumulative manner, science advances through crisis and scientific revolutions, during which an accepted paradigm is rejected and a new one is elaborated. Our starting premise is that the correlation of the LFT with the calcareous nannofossils and planktonic foraminifera stratigraphic scales elaborated during the 60's and 70's (here called the classic correlation) became generally accepted, thus qualifying as a paradigm.

The emergence of a paradigm, according to Kuhn (1970), is preceded by a period of random observations, which in the present case were made by Schaub (1951) in the Swiss Alps and by Hottinger (1960) in the Spanish Pyrenees. Their observations led them to realize that lower Paleogene sections from the northern Tethyan margin contained an association of larger foraminifera intermediate between those of the Thanetian and Cuisian strata of the classical stratotype sections of Britain and the Paris basin. To accommodate that finding in the time scale Schaub (1951) initially proposed a two-fold subdivision ("Paleocene with Nummulites" and "Early Ypresian"), but subsequently Hottinger and Schaub (1960) defined a new formal stage encompassing both subdivisions, which they named "Ilerdian". The base of this new stage was marked by the first occurrence of *Nummulites fraasi* and of *Alveolina vredenburgi* (initially named *A. cucumiformis,* see Hottinger et al., 1998).

Further studies made it evident that the appearance of these Ilerdian organisms implied a significant turnover in the evolution of larger foraminifera, as summarized by Hottinger (1998, p. 61-62): *"the biotic turnover* (was characterized by) *the diversification of species belonging to a restricted number of successful genera at the beginning of the Ilerdian stage* (…). *The diversification was linked to a considerable increase of shell size and adult dimorphism* (…). *The turnover event can be easily recognized by the microfacies documenting the rise of larger-sized and distinctly dimorph foraminifera in the field with a hand lens: all the well-known limestones with larger porcelaneous form, i.e. orbitolitids, alveolinids, lacazinids and/or the lamellar-perforate, involute nummulitids (s.str.)."* More documentation about the LFT can be found in Tambareau (1994), Pujalte et al. (2003a) and Scheibner and Speijer (this issue).

Planktic foraminifera and calcareous nannofossils are customarily scarce, if at all preserved, in the shallow-water sections where the LFT is best recorded, a circumstance that made it difficult to precisely establish the position of this event (and therefore the base of the Ilerdian Stage) in calcareous plankton biostratigraphic scales. Because of that, the correlation between these scales had to be based on data from a comparatively short number of sections in which open and shallow marine biotas coexisted, particularly on the information obtained from the south Pyrenean Campo section, the para-stratotype of the Ilerdian Stage (e.g., Schaub, 1966, 1973; Tambareau and Villate, 1974; Hillebrandt, 1975; Kapellos and Schaub, 1975). According to these and similar studies, summarized by Cavelier and Pomerol (1986), the base of the Ilerdian Stage would be situated at or near the bottom of the zone P5, in the lower part of the calcareous nannoplankton zone NP9 (= *Discoaster multiradiatus*), and just below geomagnetic chron C25n (Fig. 1A). This correlation was adopted in most subsequently published chronostratigraphic schemes containing the P-E boundary, including those produced by the IGCP 286 Working Group on Paleogene Shallow benthos of the Tethys (Serra-Kiel et al., 1998), and was further corroborated in more recent studies of the Campo section (Molina et al, 1992, 2000).

Because of such paradigmatic correlation, when data from ODP boreholes and open marine outcrop sections conclusively demonstrated that the onset of the CIE occurred during magnetochron C24r, and in mid-zone P5 (Berggren et al., 1995), the LFT was considered an event older than, and unrelated with, the PETM. This was explicitly stated by Hottinger (1998, p. 61): "*According to these correlations, the biotic turnover of shallow benthic communities* [the LFT] *seems not to correlate with any possible Paleocene-Eocene boundaries based on planktic or deeper benthic organisms* [the PETM benthic extinction event] *as proposed so far.*"

Crisis of the paradigm

To our knowledge, Baceta (1996) was the first author ever to place the base of the Ilerdian Stage in the middle part of zone P5 and within magnetochron C24r (Fig. 1B), a correlation subsequently publicized in two International Meeting dealing with the P-E boundary (Pujalte et al., 1998, 2000). However, the correlation proposed by these authors caused little stir in the scientific community, a fact that at the time was ascribed to the limited visibility of PhD reports and Meeting proceedings.

However, the alternative correlation advanced by these authors was strongly reinforced after the re-study of the

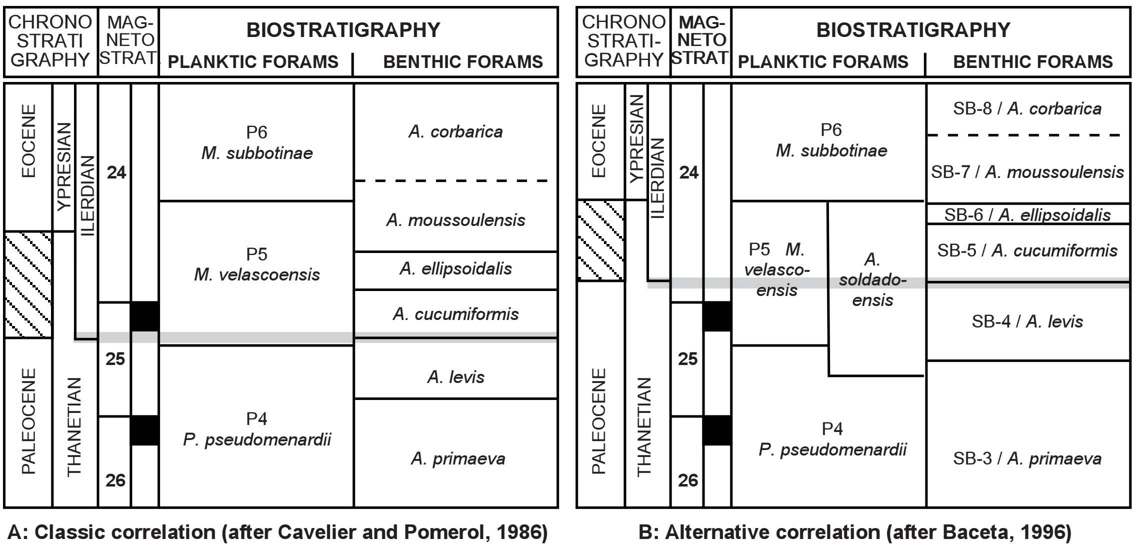


FIGURE 1 **Classic and alternative chrono-biostratigraphic schemes across the Thanetian-Ypresian interval. The position of the base of Ilerdian Stage (= *A. levis-A. cuccumiformis* boundary) is shaded in grey in both schemes.**

uppermost Paleocene and lowermost Eocene interval of the Campo section by Orue-Etxebarria et al. (2001). The results of that study, published in a major and widely circulated journal (Marine Micropaleontology), included robust biostratigraphic data on calcareous plankton that indicated that at Campo the Thanetian-Ilerdian boundary was indeed situated in the middle of Zone P5 and near the NP9-NP10 zonal boundary (*sensu* Bybell and Self-Trail, 1995). Further, Orue-Etxebarria et al. (2001) carried out a correlation between Campo and two deep-marine sections in the Basque basin (Ermua and Trabakua pass), in which the benthic extinction event and the onset of the CIE had previously been identified (Fig. 2). It was on the basis of such correlation that Orue-Etxebaria et al. (2001) explicitly proposed that the LFT and the PETM were coeval. Such proposition was further reinforced after the study of two outer platform sections in the southern Pyrenees (Urrobi and Mintxate, Fig. 2A), which included data on larger and planktonic foraminifera, calcareous nannofossils and stable isotopes from bulk-rock samples (Pujalte et al., 2003a), and again at Campo, with new calcareous nannofossil and magnetostratigraphic data (Pujalte et al., 2003b).

Another important breakthrough was made by Schmitz and Pujalte (2003, 2007), based on numerous isotopic analyses of carbonate soil nodules. They reported that in many sections of the Tremp-Graus basin, to the east of the study area, the PETM was recorded by two comparatively thin but laterally extensive lithological units, the 4 m thick Claret Conglomerate below and the 20 m thick Yellowish Soils above, both situated beneath marine Ilerdian deposits. In these sections, the $\delta^{13}C$ composition of nodules from a 200 m thick "Garumnian" red bed sequence underlying the Claret Conglomerate typically ranged between –6‰ and –9‰. Instead, nodules from the Yellowish Soils always produced very negative $\delta^{13}C$ values (between –11‰ and –14‰, averaging –13‰). This represent a shift of about –6‰, equivalent in magnitude to the one observed in soil nodule profiles from well-known Paleocene-Eocene localities in Wyoming and China (Koch et al., 1992, 2003; Bowen et al., 2001, 2002). Claret Conglomerate soil nodules, which occur at rare clay intercalations within the unit, or at its upper sandy part, yielded isotopic values slightly less negative than those of the Yellowish Soils, varying between –9‰ and –13‰ and averaging 10.5‰. Schmitz and Pujalte (2007) therefore concluded that the Yellowish Soils were developed during peak PETM conditions, whereas the Claret Conglomerate was formed during the earliest part of the event.

More recently, Scheibner et al. (2005) and Scheibner and Speijer (this issue) reported results from a platform-to-basin transect in the southern Galala Mountains of Egypt, also concluding that the LFT closely correlates with the PETM. That conclusion entailed that the LFT-PETM synchronism is not just a local feature, adding a strong support to the alternative correlation worked out in the Pyrenees.

Yet, despite the growing evidence in favour of the alternative correlation, the implications of the LFT-PETM synchronism have been so far largely overlooked. Thus, while the ecological impact of the PETM on groups as diverse as deep-water benthic foramifera or continental biotas has been widely publicized (see above), very few papers other than those of Orue-Etxebarria et al. (2001), Pujalte et al. (2003a), Scheibner et al. (2005, 2007) or Scheibner and Speijer (this issue) have so far explicitly mentioned a possible influence of the PETM in the evolution of larger foraminifera. This is remarkable, considering that larger foraminifera were major rock builders in the extensive early Paleogene carbonate platforms of the Tethys, and their turnover implies that the shallow-marine ecosystem was severely altered by the thermal event. Such overlooking can, in part, be attributed to the vast amount of new literature published nowadays, which makes it difficult to divulge new ideas among the scientific community, and also to a current bias to base paleoecological and paleoceanographic studies on open marine microfossils, a bias enhanced by the pre-eminence of ODP-based investigations.

In addition to this, however, some scepticism may have arisen from the fact that some specialists in larger foraminifera have been reluctant to accept the LFT-PETM synchronicity, as shown by the following two examples: "*The question arises whether the evolution of the larger foraminifera during the Thanetian and Ilerdian* [the LFT] *is affected by the abiotic event currently used to define the Paleocene-Eocene boundary* [the PETM]" (Drobne, 2003). Or, more explicitly, "*The time-equivalence of the radiation event among the larger foraminifera* [the LFT] *with the PETM (Orue-Etxebarria et al., 2001) cannot be confirmed at Campo section*" (Molina et al., 2003).

A devil's advocate approach to the databases in which the PETM-LFT correlation is based indicate that, although their combined evidence is compelling, the information provided separately by individual Pyrenean or Egyptian sections so far studied might not be without ambiguities. Thus, while in deep-water marine sections the location of the PETM is well established, the position of the LFT relies upon resedimented larger foraminifera. Conversely, in shallow-water marine sections where the LFT is well constrained the isotopic signature is either obliterated or unclear. Also, in the continental sections of the Tremp-Graus basin, the base of the marine Ilerdian sequence either overlies the Yellowish Soils or is situated a few meters above their top (Schmitz and Pujalte, 2003,

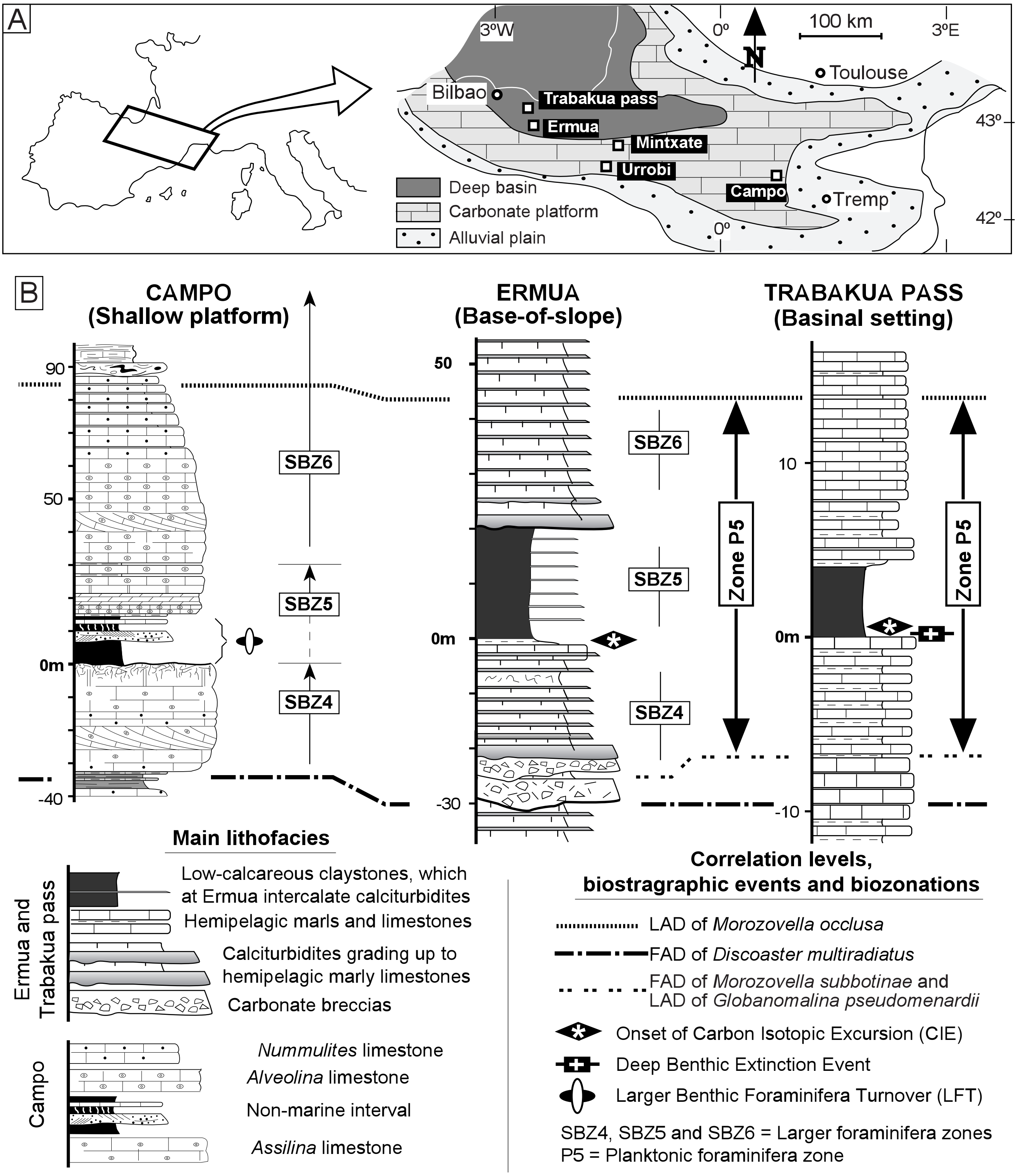


FIGURE 2 **2A Paleocene paleogeography of the Pyrenean area and location of sections cited in the text. B: Correlation of the Campo, Ermua and Trabakua pass sections, showing the location of the onset of the CIE, the Benthic extinction event and the LFT (modified from Orue-Etxebarria et al., 2001).**

2007). Strictly speaking, therefore, the data from the Tremp-Graus basin demonstrate that the PETM preceded the LFT, but do not prove the synchronicity of both events. Clearly, the proposed temporal concurrence of the LFT and the PETM hinges on a biostratigraphic correlation between sections of different settings, and that might be one of the reasons for the reluctance of some specialists to accept it.

NEW DATA CONFIRMING THE LFT-PETM CORRELATION

Materials and methods

In an attempt to resolve the controversy surrounding the LFT-PETM correlation, further fieldwork has been carried out across the P-E boundary interval in a comparatively small area to the east of the Campo section (Fig. 3). This area was selected because earlier studies (e.g., Eichenseer and Luterbacher, 1992; Serra-Kiel et al., 1994; Robador, 2005) had demonstrated that the predominantly shallow-marine succession cropped out at Campo interfingers rapidly to the east with continental "Garumnian" deposits, in which the position of the PETM is now well constrained (Schmitz and Pujalte, 2003, 2007; see above).

The new study has included: 1) a re-appraisal of the Campo section and a high-resolution analysis of three new sections, La Cinglera, San Martín and Bacamorta (Figs. 3, 4, and 5); 2) a detailed field mapping of the area, using vertical air photos at scales ranging from 1:15000 to 1:2500, which has allowed a high-resolution correlation of the four sections and of the intervening outcrops; 3) a reconstruction of the sedimentary architecture of the studied continental-marine transect across the P-E interval (Fig. 6); 4) analysis of the stable isotopic composition of 42 soil carbonate nodules from 22 samples collected at La Cinglera and San Martín which, together with earlier isotopic data from Campo, have provided crucial information to age date key intervals of the succession (Fig. 4); and, 5) isotopic analyses of three bulk-rock samples and one pilot analysis of a gastropod shell from San Martín. For details of the analytical procedure, see Schmitz et al. (2001) and Schmitz and Pujalte (2003).

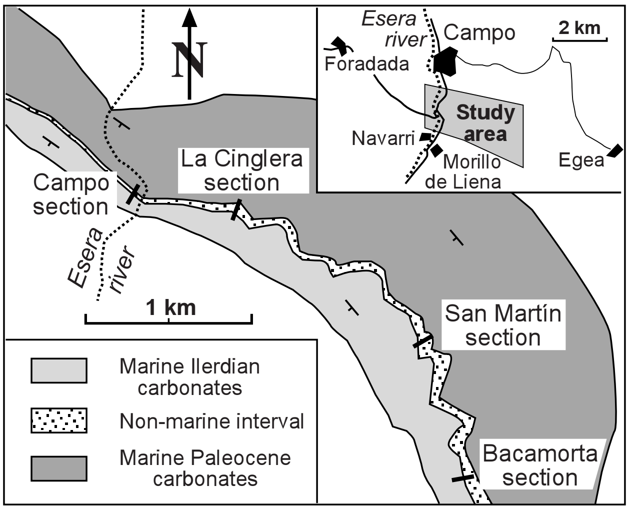


FIGURE 3 | **Location of the studied sections on a simplified geological map of the study area. For paleogeographical context of the Campo area see Fig. 2A.**

Results

Stratigraphy and depositional settings

Eight lithological units of variable thickness have been recognized in the studied sections and in the reconstructed transect (Figs. 4 and 6), which for the purposes of this paper are numbered from 1 to 8. They are described and interpreted below in ascending stratigraphic order.

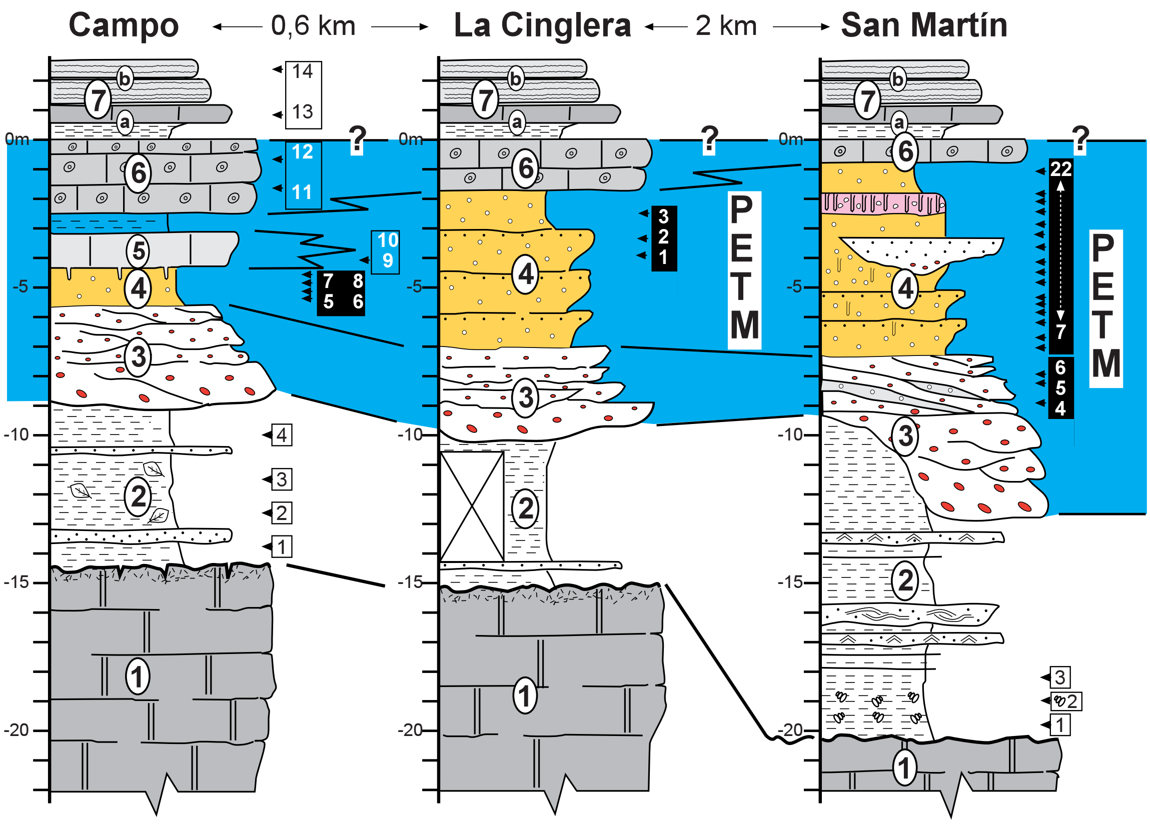
Unit 1

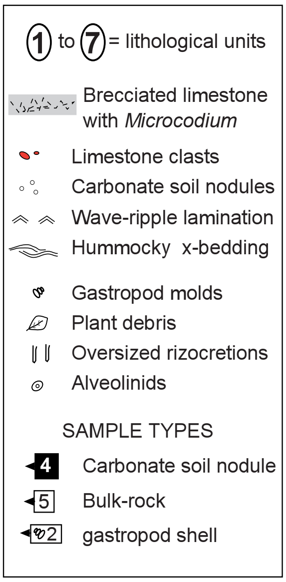
This unit encompasses the upper part of the shallow marine Paleocene carbonate sequence and is composed of thick-bedded bioclastic sandy packstones and grainstones rich in red coralline algae, with subordinate coral and bryozoan remains (Payros et al., 2000). In addition, larger foraminifera have been reported in beds situated a few meters below the top of the unit, including *Alveolina (G.) levis, Assilina yvettae* and *Operculina azilenzis* (Tambareau and Villate, 1974; Eichenseer and Luterbacher, 1992; Serra-Kiel et al., 1994; Robador, 2005; Scheibner et al., 2007). This is a typical association of the late Thanetian SBZ4 (shallow benthic foraminifera zone) of Serra-Kiel et al. (1998).

The upper boundary of unit 1 is a sharp erosional surface that, to the east of Campo, cut down progressively across the succession, having eroded away at least 5 m of marine carbonates in the San Martín and Bacamorta sections (Figs. 4 and 6). At Campo, the uppermost 50 cm of the marine marine limestones are brecciated below this surface and contain abundant *Microcodium* remains, a clear proof of subaerial exposure, as noted by previous authors (Molina et al., 1992; Serra-Kiel et al., 1994; Payros et al., 2000; Scheibner et al., 2007). However, the location of magnetochron C25n and of *Discoaster multiradiatus* in beds situated respectively at –44/–49 m and at –35 m below the top of the unit demonstrates that the hiatus associated with this surface is of small magnitude (Pujalte et al., 2003b).

Unit 2

This unit ranges in thickness between 5 and 10 m, and is mostly composed of grey, weakly calcareous mudstones, with lesser intercalations of calcareous sandstones. The mudstones are massive to crudely laminated and usually devoid of fossils. At Campo, however, small plant leaves and coaly fragments are concentrated at some





Isotopes Campo section
(data from Schmitz & Pujalte, 2003)

Sample	$\delta^{13}C$‰	$\delta^{18}O$‰
☐ 1 -->	-6,10	-5.37
☐ 2 -->	-6,68	-5.27
☐ 3 -->	-7,44	-2.89
☐ 4 -->	-5.68	-5.67
■ 5 -->	-8.09 // -8.75	-5.64 // -6,21
■ 6 -->	-9.41 // -8.86	-5.90 // -5,93
■ 7 -->	-9.24 // -9.75	-5.75 // -6,01
■ 8 -->	-9.83 // -10.95	-6.04 // -5,93
☐ 9 -->	-6,27	-2.89
☐ 10 -->	-4.27	-0.96
☐ 11 -->	+1.38	-3.13
☐ 12 -->	+1.83	-3.23
☐ 13 -->	+1.98	+1.06
☐ 14 -->	-0.25	-3.87

Isotopes La Cinglera section

Sample	$\delta^{13}C$‰	$\delta^{18}O$‰
■ 1 -->	-11.10	-6.16
■ 2 -->	-11.51 // -11.59	-6.21 // -6,27
■ 3 -->	-11,42 // -11,69	-6,18 // -6,25

Isotopes San Martín section

Sample	$\delta^{13}C$‰	$\delta^{18}O$‰
☐ 1 -->	-8.70	-6.20
2 -->	-5,18	-5.30
☐ 3 -->	-8.91 // -8.60	-6.13 // -5,94
■ 4 -->	-11.44 // -11.80	-6,05 // -6,24
■ 5 -->	-12.92 // -11.65	-6,21 // -5,52
■ 6 -->	-13.18 // -13.50	-6.05 // -6,19
■ 7 -->	-13,54 // -11,09	-6.13 // -6,10
■ 8 -->	-11,97 // -14,00	-6.43 // -6,52
■ 9 -->	-14,30 // -14,16	-6.55 // -6,45
■ 10 -->	-12,86 // -13,04	-6.07 // -6,01
■ 11 -->	-13,28 // -12,86	-5.88 // -6,19
■ 12 -->	-11.37	-6.40
■ 13 -->	-13,87 // -14,06	-6.33 // -6,40
■ 14 -->	-13,16 // -12,77	-6.16 // -6,35
■ 15 -->	-11,89 // -13,97	-6.43 // -6,23
■ 16 -->	-10,95 // -11,82	-6.25 // -6,14
■ 17 -->	-12,61 // -13,01	-5.73 // -5,97
■ 18 -->	-13,67 // -13,80	-5.94 // -6,04
■ 19 -->	-13,67 // -13,89	-5.94 // -5,81
■ 20 -->	-13,06 // -13,23	-5.90 // -5,86
■ 21 -->	-13,50 // -13,81	-5.79 // -5,98
■ 22 -->	-14,30 // -14,16	-6.55 // -6,45

FIGURE 4 **Lithologs of the Campo, La Cinglera and San Martín sections, showing the different units recognized and the location and type of samples used for isotopic analysis. The position of the PETM interval is highlighted. Boxes include key to symbols and isotopic values. For most samples, two nodules were analysed. Precision of isotopic analyses is ca ± 0.02 ‰ for $\delta^{13}C$ and ± 0.04‰ for $\delta^{18}O$. Isotopic values are given as ‰ relative to PDB (cf. Schmitz et al., 2001).**

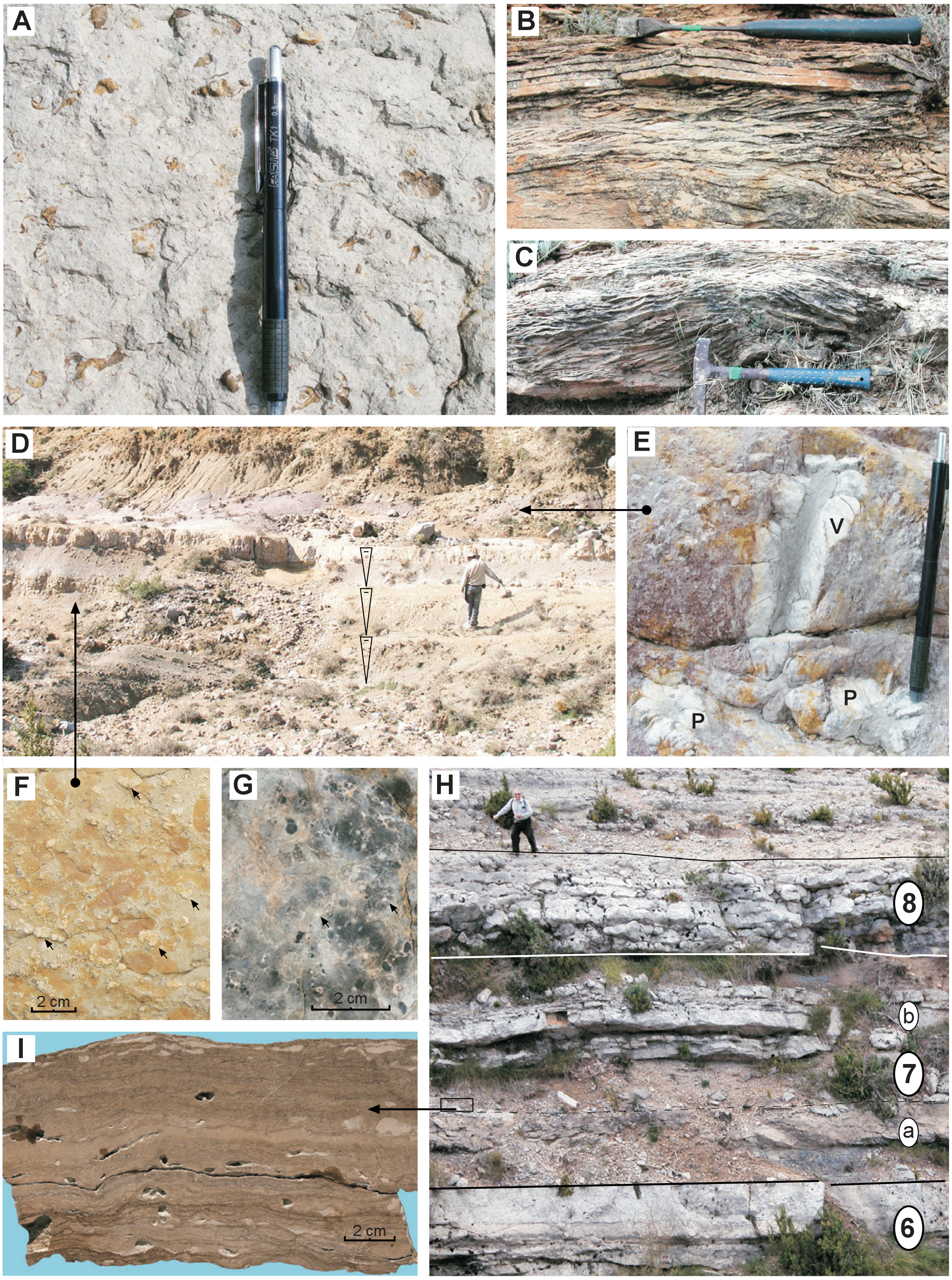


FIGURE 5 **Field photos of representative facies and outcrops. A) Grey weakly calcareous mudstones with gastropod remains (unit 2, San Martín section); B, C) Calcareous sandstones respectively exhibiting cross-lamination and hummocky cross-bedding (unit 2, San Martín section). D) small-scale coarsening-up sequences of yellowish mudstones grading up to fine-grained sandstones, overlain by an interval of purple mudstones with oversized rhizocretions (unit 4, San Martín section). E) close-up of the oversized rhizocretions of the purple interval of unit 4 as seen in vertical (v) and plan-view sections (p) (San Martín section). F) Close-up of a mudstone of unit 4, showing mottling and numerous small-sized carbonate soil nodules (arrowed); mudstones of yellowish colour like the one shown form the bulk of unit 4 in all sections but Campo. G) Close-up of the polished surface of a mudstone of unit 4 from Campo, similar in appearance to that of F, but dominantly grey in colour. H) Lower part of the marine Ilerdian succession at La Cinglera, showing units 6 to 8; note the brown colour of subunit 7a, which contrast with the grey colour of all remaining strata. I) Polished slab of a limestone of subunit 7b showing microbial laminations and evaporite nodules.**

levels, and rare amber remains have been observed. In addition, at San Martín the grey mudstones contain abundant gastropods and occasional bivalves, generally just as inner moulds (Fig. 5A). These moulds are a clear sign of post-depositional dissolution of carbonate, and open the possibility that other carbonate fossils might be missing in this unit. Neither *Microcodium* nor other proofs of subaerial exposure such as soil nodules, roots or rhizocretions have been observed. The intercalated calcareous sandstones occur as laterally extensive beds up to 65 cm thick, which often exhibit cross laminations and, less frequently, hummocky cross-stratification (Figs. 5B and C). All these features suggest a non-marine shallow aquatic environment, generally of low-energy, but occasionally affected by storms.

Unit 3

This unit is composed of channel-like bodies, 2 to 5 m thick, made up of cross-bedded and parallel-laminated conglomerates and pebbly sandstones, which are usually arranged in fining-up sequences that may contain mudstone intercalations in their upper parts (Fig. 4). The lateral continuity of these channel bodies is difficult to assess in the field, because of poor exposure. However, a correlation between nearby outcrops and sections suggests that individual channels may extend laterally for at least 100 m, and tend to become laterally amalgamated into more extensive composite packages (Fig. 6). Most clasts of the conglomerates are either derived from Mesozoic limestones or from the upper Cretaceous Aren sandstone, and vary between pebble and cobble size, rarely exceeding 5 cm in diameter. Unit 3 is almost devoid of authochtonous fossils, the exception being occasional plant remains included in the sandstones. Soil carbonate nodules have been found within mudstone intercalations at the top of a channel sequence at San Martín (Fig. 4). These features suggest that unit 3 was deposited within braided fluvial channels. Indeed, as discussed below, unit 3 is considered to be the distal correlative of the Claret Conglomerate, an alluvial megafan accumulation developed at the onset of the PETM (Schmitz and Pujalte, 2007).

Unit 4

This unit shows somewhat different features in Campo than in the other three studied sections. Unit 4 is up to 7 m thick in Bacamorta, San Martín and La Cinglera, and is mostly made up of unfossiliferous silty mudstones and fine-grained calcareous sandstones, usually arranged in small scale, coarsening-up sequences (Fig. 5D). Besides, it includes intercalations of channelized cross-bedded pebbly sandstones, not unlike those of unit 3, but of smaller dimensions (Figs. 4 and 6). Most mudstones and fine-grained sandstones of these three sections exhibit a characteristic yellowish-brownish mottling in weathered exposures, and contains abundant, randomly dispersed pedogenic carbonate soil nodules, rarely larger than 1 cm in diameter (Figs. 5D and F). However, in the upper part of the San Martín section, there also exists a 1 m thick interval of purple mudstones with abundant carbonate nodules that are up to 3 cm in diameter (Fig. 5D). These nodules indicate well-drained soil conditions. In addition, carbonate rhizocretions are common at several levels, a clear proof of roots of higher plants (Klappa, 1980) and of subaerial exposure. Most rhizocretions are less than 1 cm across, although they reach 6 cm across in the purple level of San Martín (Fig. 5E).

In contrast, unit 4 is only ca. 1.5 m thick in the Campo section and is represented by mottled calcareous mudstones that, except for some minor irregular yellowish patches, are grey in colour (Figs. 4 and 5G). Further, carbonate nodules are much less abundant at Campo than in the other sections and also of smaller in size, usually <0.5 cm in diameter. Despite these differences, field data undoubtedly demonstrate the lateral continuity of unit 4 from Bacamorta to Campo (Fig. 6). However, the predominance of grey colours indicates more reduced conditions at Campo than elsewhere, and suggest that the soils of this section may have been waterlogged for at least part of the year. Such situation would also explain the relative scarcity of pedogenic carbonate nodules at Campo, and their small size.

All these features indicate that the bulk of unit 4 represents floodplain and overbank deposits of a fluvial system. Field evidence further demonstrates that unit 4 interfingers laterally with marine deposits (see below), which implied that unit 4 was accumulated in a coastal alluvial plain. Campo was the section nearer to the paleoshoreline, a fact that probably account for the observed differences in soil development.

Unit 5

This unit is represented by a mere 1.3 m thick bed of non-fossiliferous grey micritic limestone. Unit 5 only occurs in the Campo section, pinching out rapidly to the SE within unit 4 (Figs. 4 and 6). Its lower boundary is conformable and gradational, with thick rhizocretions protruding from its base and going down nearly 1 m into the underlying unit 4 (Fig. 4). The limestone contains silt-sized quartz grains and rare fragments of charophyte stems, the latter suggesting a fresh or brackish-water depositional environment. Thin-sections of the limestone reveal incipient pedogenetic features, such as poorly developed circumgranular cracks and glaebulae (Payros et al., 2000), suggestive of palustrine conditions, a possibility reinforced by its basal rhizocretions. Palustrine carbon-

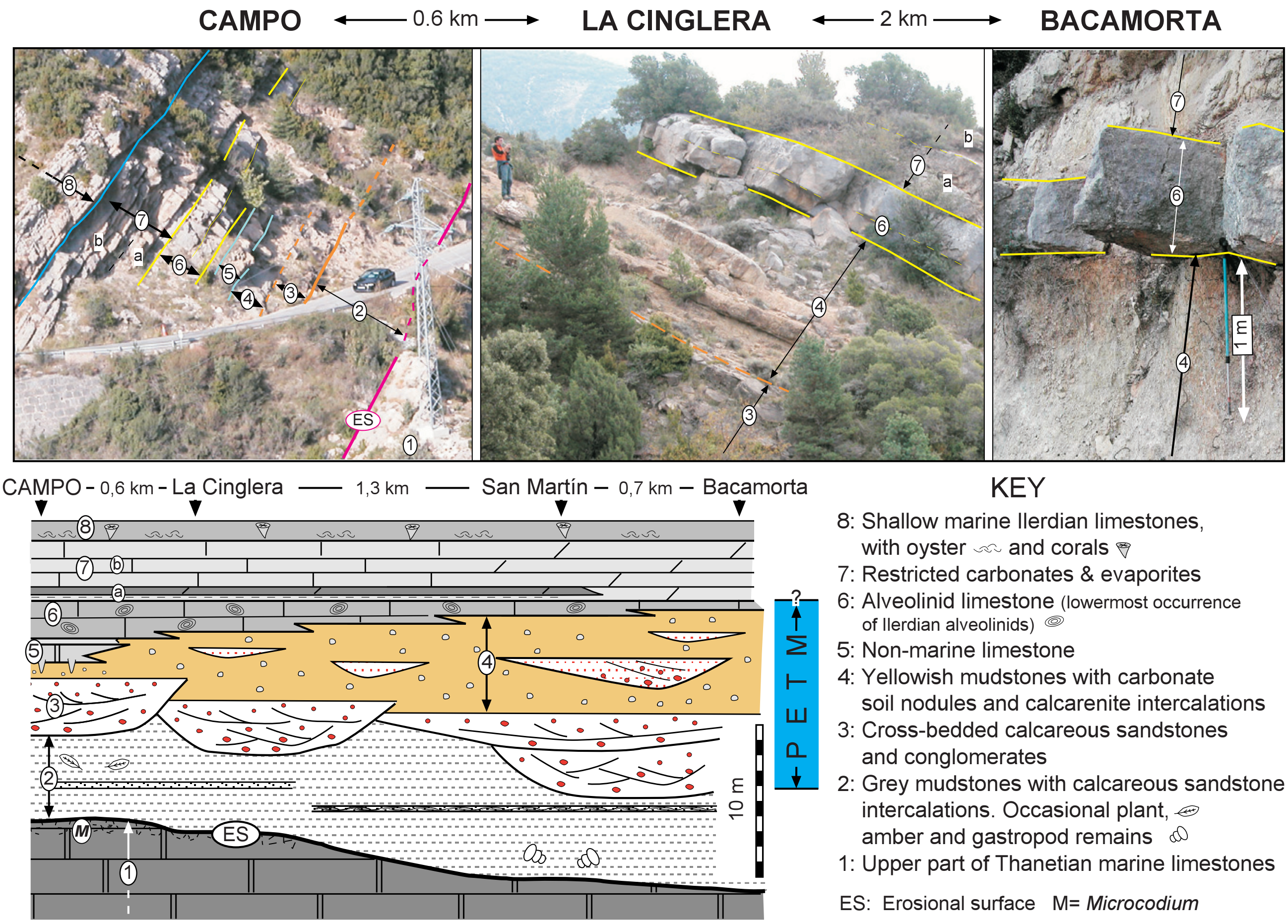


FIGURE 6 | **Above, field photos of the Campo, La Cinglera and Bacamorta sections, with indication of lithological units 1 to 8 described in the text. Below, reconstructed transect showing the sedimentary architecture of the studied interval. For this reconstruction, the boundary between units 6 and 7 has been used as datum.**

ates are the result of exposure and pedogenesis of lacustrine mud, and this implies that unit 5 was deposited in a zone with oscillating water-table and/or seasonally inundated. Such zone might either correspond to the margin of a lake or lagoon, or even to a floodplain. Indeed, unit 5 was interpreted by Eichenseer and Luterbacher (1992) as a calcimorph floodplain paleosoil.

Unit 6

This unit is lithologically very similar in all studied sections, being exclusively made up of thick-bedded, fine-grained (wackstone-packstone) limestones. This is the lowest unit of the succession containing Ilerdian alveolinids, which belong to the early Ilerdian *A. vredenburgi* or SBZ5 zone (e.g., Schaub, 1973; Eichenseer and Luterbacher, 1992; Serra-Kiel et al., 1994; Robador 2005). Indeed, the base of the Ilerdian Stage, for which Campo is the parastratotype, was originally placed at the base of this unit (Schaub, 1966 and 1973). Alveolinid specimens occur randomly dispersed throughout the unit, although they tend to be more abundant in the upper parts of the strata. Such random distribution argues against resedimentation and, together with the fine-grained character of the limestones, indicates a low energy environment, probably a sheltered marine embayment. Alternatively, unit 6 was attributed to a lagoon by Eichenseer and Luterbacher (1992) and to an open water inner platform by Scheibner et al. (2007).

Despite its lithological homogeneity, unit 6 decreases in both number of beds and in composite thickness toward the southeast, the unit having 3 beds at Campo that amount to 2.6 m, and just one bed, 0.6 m thick, at Bacamorta (Figs. 4 and 6). Any explanation of these changes must take into account the following field observations: 1) the thickness reduction of unit 6 is concurrent with a thickness increase of units 4+5; 2) there is neither evidence of onlap of unit 6 onto unit 4, nor of unconformity between units 6 and 7; and 3) no measurable thickness change has been observed in the overlying units 7 and 8. Under these constraints, two alternative scenarios can be envisaged, which are discussed below.

Scenario A, unit 6 is everywhere younger than units 4+5. This required that i) subsidence was greater in Bacamorta than in Campo during deposition of units 4+5; ii) subsidence was lesser in Bacamorta than in Campo during deposition of unit 6; iii) no differential subsidence occurred between Bacamorta and Campo during deposition of units 7 and 8.

Scenario B, units 6 and 4 are coeval, the southeastward thinning of the former being the result of its lateral interfingering with the latter. The only assumption needed in this second scenario is that no significant differential subsidence occurred between Bacamorta and Campo during the deposition of units 4 to 8. Therefore, based on the Occam's Razor principle, scenario B is considered to be far more likely than scenario A (Figs. 4 and 6).

Unit 7

This unit is named after a set of lithologically varied strata, collectively about 5 m thick, sandwiched between units 6 and 8 (Figs. 5H and 6). In the Bacamorta section unit 7 is entirely made up of secondary dolomites. Elsewhere, two subunits can be recognized:

Subunit 7a comprises a lower marly interval (65 cm) and an upper calcareous bed (70 cm). The latter is made up of featureless, neomorphised limestone or fine grained dolomite and has a characteristic light brown colour in weathered surfaces (Fig. 5H), which is in striking contrast with all other grey coloured strata of units 6, 7 and 8. This has made it possible to trace subunit 7a from Campo to San Martín without any noticeable change of facies or thickness (Fig. 6).

Subunit 7b (ca. 4 m) is composed of fine-grained, laminated limestones and dolomites. Polished slabs and thin sections reveal thin, subtle and irregular laminations clearly related to microbial mats, which often appear disrupted by synsedimentary nodules (Fig. 5I). Most of these nodules are now replaced by sparite or chert but, according to Eichenseer and Luterbacher (1992), were originally made of evaporite minerals, mainly gypsum and celestine. Further, Schbeiner et al. (2007) report ostracods and rare miliolids and benthic foraminifera in this subunit. All these authors interpreted the depositional setting of unit 7 as a semiarid tidal flat, an interpretation subscribed here.

Unit 8

This unit, the youngest of the studied interval, is up to 4 m thick and consist of medium-bedded, nodular-like limestones (Fig. 5H). As noted by Eichenseer and Luterbacher (1992) and by Schbeiner et al. (2007), the unit is floored by abundant oyster remains and contains small corals, gastropods and encrusting coralline algae, a fossil association clearly indicative of shallow open-marine conditions. Unit 8, therefore, is a testimony of the marine re-flooding of the previous tidal flat.

Stable isotopes

Reconstructions of past isotopic variations in water-laid carbonate deposits should ideally be based on analyses of well-preserved, monospecific calcite tests or shells, but in the outcropped and indurated lower Paleogene rocks of the Pyrenees such materials cannot be obtained. Alternatively, bulk-rock samples and pedogenic carbonate nodules were used in this study.

High-quality $\delta^{13}C$ data can be extracted from fine-grained marine limestone lithified during early diagenesis (e.g., Schmitz et al., 2001). However, the bulk-rock samples analysed by Schmitz and Pujalte (2003) and in this study were either coarse-grained, marly or clayey samples, of low quality for isotopic work.

Soil carbonate nodules, on the other hand, can be successfully used to track past changes in the carbon isotopic composition of the atmosphere for two main reasons: one, the carbon in these nodules is derived from atmospheric CO_2 and from the organic decomposition and root respiration (Cerling et al., 1991; Bowen et al., 2001); and two, soil nodules usually grow within fine-grained sediments and thus represent relatively closed systems with respect to carbon isotopes. Hence, nodules were used whenever possible for isotopic analyses in this study.

Two bulk-rock samples collected from unit 2 at San Martín gave $\delta^{13}C$ values of ca. –8.7‰, and the pilot analysis of one gastropod carbonate shell from the same unit was somewhat less negative (–5.2‰), which are similar to the ones earlier obtained at Campo from the same unit by Schmitz and Pujalte (2003) (Fig. 4). These are typical background values for lower Paleogene non-marine carbonates but, for the reasons above, should be interpreted with caution. However, less than 10 km to the east of the study area, in the Rin section, unit 2 contains well-developed soil carbonate nodules that produced comparable $\delta^{13}C$ values (Schmitz and Pujalte, 2007; see fig. DR-4 from Data Repository item 2007048, available at www.geosociety.org/pubs/ft2007.htm). Consequently, it can be safely stated that unit 2 pre-dates the PETM.

Twenty two samples of carbonate nodules were collected from different horizons at both the San Martín and

La Cinglera sections, where they are plentiful in unit 4 and sparser in unit 3 (Fig. 4). Two nodules per sample were generally analized, their $\delta^{13}C$ values being always very negative, varying from -11.0 to –14.3‰, and averaging –12.9‰. These are nearly the same values provided by PETM soil nodules in continental sections to the east of the study area by Schmitz and Pujalte (2003, 2007) and leave little doubt that both units developed during the PETM. Because of this fact, and taking also into account their sedimentary features and their stratigraphic and paleogeographical position, it is evident that units 3 and 4 are but the distal equivalents of, respectively, the Claret Conglomerate and the Yellowish Soils of the eastern sections.

DISCUSSION

Before the present study, the search for the PETM in the area had only been attempted in the classical Campo section, with ambiguous results. Molina et al. (2000 and 2003), for instance, suggested three possible alternative levels for the event, two of them situated at 165 m and 85 m above the base of the marine Ilerdian succession, the third one at the base of the marine Ilerdian succession itself. Orue-Etxebarria et al. (2001) and Scheibner et al. (2007) placed it even lower, within the non-marine interval represented by units 2 to 5, but they did not pinpoint its precise position. Finally, Schmitz and Pujalte (2003) found soil nodules in the upper part of the non-marine succession (unit 4 of this paper) with $\delta^{13}C$ values between –8 and –11‰ (Fig. 4), which they considered a somewhat diagenetically altered signal of the PETM.

Bulk-rock samples from unit 5, however, produced less $\delta^{13}C$ negative values (–4.3‰ and –6.27‰, Fig. 4), whereas those of units 6 and 7 mostly provided positive ones, the shift from negative to positive values clearly reflecting the passage from a fresh-water to a marine environment. On that evidence, Schmitz and Pujalte (2003) assigned units 5, 6 and 7 to the post-PETM interval. The new field and isotopic data here presented force a reassessment of these interpretations.

Firstly, the new isotopic data from pedogenic carbonate nodules from La Cinglera and San Martín permit to unequivocally place the onset of the PETM at the base of unit 3. This finding is valuable in itself because the base of the PETM has now been agreed as the global criterion for the P-E boundary (Luterbacher et al., 2000, 2004). The explanation for the ambiguous isotopic values earlier reported from unit 4 at Campo remains speculative. One possible explanation is that the soils of unit 4 were more waterlogged there than in the other sections, making the isotopic $\delta^{13}C$ values somewhat less negative. This issue, however, is no longer crucial, since the new field data allow a direct correlation between Campo, La Cinglera and San Martín (Fig. 6).

However, contrary to earlier reports, the field evidence now demonstrates that in all probability units 5 and 6 were also developed during the PETM, since both units are laterally interfingered with unit 4 (= Yellowish soils; Figs. 5 and 6, and above). This fact is crucial for the main purpose of this paper. In effect, the alveolinids of unit 6 are the earliest known autochthonous representatives of Ilerdian larger foraminifera and, for the first time, they have been found to occur in beds obviously deposited during the thermal event. The synchronism of the LFT and the PETM is thus conclusively demonstrated.

Since the bulk-rock isotopic values of unit 6 and 7 are similar (Fig. 4), one may further speculate whether or not the latter was also accumulated, totally or in part, during the PETM. This question, however, cannot be answered with the data so far available. Therefore, the position of the top of the PETM in the study area remains unfixed, pending future field and/or isotopic information.

CONCLUSIONS

1) The new data presented in this paper confirm that the P-E boundary (as recently defined by the International Commission on Stratigraphy) coincides in the study area with the base of unit 3, which is the distal equivalent of the Claret Conglomerate of Schmitz and Pujalte (2007). Earlier suggestions that this boundary might be located at the top erosional surface of unit 1, at the base of the marine Ilerdian beds, or higher up in the succession, must therefore be abandoned.

2) More important for the main purpose of this paper, the synchronism of the LFT and the PETM can no longer be doubted. In addition to the large amount of data earlier provided by Orue-Etxebarria et al (2001), Pujalte et al. (2003a), Scheibner et al. (2005) and Scheibner and Speijer (this issue), the present paper demonstrates that in the study area Ilerdian marine beds containing the oldest-known autochthonous *A. wredenburgi* (a proof of the LFT) are laterally interfingered - and are therefore coeval-with the PETM Yellowish Soils of unit 4.

3) Larger foraminifera were major rock-building organisms in tropical and subtropical shallow seas of the early Paleogene, and the confirmation that their Ilerdian turnover was coeval with the PETM is a clear proof of the impact of the thermal event on shallow marine benthic communities. The expanded southern Pyrenean sections may provide key information to better understand the mechanisms of such impact. However, before endeavour-

ing such analysis, more data will be necessary to confidently establish the end of the PETM in the Ilerdian marine succession of the study area.

4) The long-held classical miscorrelation between larger foraminifera and calcareous plankton stratigraphic scales must be rejected, as anticipated by Baceta (1996) more than ten years ago. The long delay in abandoning this erroneous correlation is a clear example of the difficulties to replace established paradigms.

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REFERENCES

Aubry, M.P., 1998. Early Paleogene calcareous nannoplankton evolution: A tale of climatic deterioration. In: Aubry, M.-P., Lucas, S., Berggren, W.A. (eds.). Late Paleocene-early Eocene climatic and biotic events in the marine and terrestrial record. New York, Columbia University Press, 158-203.

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/Euskal Herriko Unibertsitatea, 372 pp.

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

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. SEPM (Society for Sedimentary Geology), Special Publication, 54, 129-212.

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. Ann Arbor, 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, Ys., Wang Y., 2002. Mammalian Dispersal at the Paleocene/Eocene Boundary. Science, 295, 2062-2065.

Bybell, L.M., Self-Trail, J.M., 1995. Evolutionary, biostratigraphic and taxonomic study of calcareous nannofossils from a continuous Paleocene/Eocene boundary section in New Jersey. US Geological Survey Professional Paper 1554, 1-36.

Cavelier, C., Pomerol, Ch., 1986. Stratigraphy of the Paleogene. Bulletin de la Société géologique de France, 8, 255-265.

Cerling, T.E., Solomon, D.K., Quade, J., Bowman, J.R.., 1991. On the isotopic composition of carbon in soil carbon dioxide. Geochimica and Cosmochimica Acta, 55, 3403-3405.

Clyde, W.C., Gingerich, P. D., 1998. Mammalian community response to the latest Paleocene thermal maximum: An isotaphonomic study from the northern Bighorn basin, Wyoming. Geology, 26, 1011-1014.

Crouch, E.M., Heilmann-Clausen, C., Brinkhuis, H., Morgans, H.E.G., Rogers, K.M., Egger, H., Schmitz, B., 2001. Global dinoflagellate event associated with the late Paleocene thermal maximum. Geology, 29, 315-318.

Dickens, G.R., O'Neil, J.R., Rea, D.K., Owen, R.M., 1995. Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of the Paleocene. Paleoceanography, 10, 965-971.

Drobne, K., 2003. Larger foraminifera at the Paleocene/Eocene Boundary on the Adriatic Carbonate Platform. Abstract and program book of the Symposium on the Paleogene. Leuven, Belgium, p. 18.

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.

Gibbs, S.J., Bown, P.R., Sessa, J.A., Bralower, T.J., Wilson, P.A., 2006. Nannoplankton Extinction and Origination Across the Paleocene-Eocene Thermal Maximum. Science, 314, 1770-1773.

Gingerich, P.D., 2003. Mammalian responses to climate change at the Paleocene–Eocene boundary: Polecat Bench record in the northern 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. The Geological Society of America, Special Paper, 369, 463-478.

Hillebrandt, A. von, 1975. Corrélation entre les biozones de grans Foraminifères et de Foraminifères planctoniques de l'Ilerdien. Bulletin de la Société Géologique de France, 17(2), 162-167.

Hottinger, L., 1960. Recherches sur les Alvéolines du Paléocène et de l'Eocène. Schweizerische Palaeontologische Abhandlungen, 75/76, 1-243.

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.

Hottinger, L., Sameeni, S.J., Butt, A.A., 1998. Emendation of *Alveolina vredenburgi* Davies and Pinfold, 1937 from the Surghar range, Pakistan. In: L. Hottinger, K. Drobne (eds.). Paleogene Shallow benthos of the Tethys, 2. Ljubljana, Dela-Opera Sazu, 155-163.

Jaramillo, C., Rueda, M.J, Mora, G., 2006. Cenozoic Plant Diversity in the Neotropics. Science, 311, 1893-1896.

Kapellos, C., Schaub, H., 1975. L'Ilerdien dans les Alpes, dans les Pyrénées et en Crimée: corrélation de zones à grands foraminifères et à nannoplancton. Bulletin de la Societé géologique de France, 17, 148-161.

Kelly, D.C., 2002. Response of Antarctic (ODP) planktonic foraminifera to the Paleocene-Eocene thermal maximum: Faunal evidence for ocean/climate change. Paleoceanography, 17, 1071.

Kelly, D.C., Bralower, T.J., Zachos, J.C., Premoli Silva, I., Thomas, T., 1996. Rapid diversification of planktonic foraminifera in the tropical Pacific (ODP Site 865) during the late Paleocene thermal maximum. Geology, 24, 423-426.

Kennett, J.P., Stott, L.D., 1991. Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Palaeocene. Nature, 353, 225-229.

Klappa, C. F., 1980. Rhizoliths in terrestrial carbonates: classification, recognition genesis and significance. Sedimentology, 27, 613-629.

Koch, P.L., Zachos, J.C., Gingerich, P.D., 1992. Correlation between isotope records in marine and continental carbon reservoirs near the Palaeocene/Eocene boundary. Nature, 358, 319-322.

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. The Geological Society of America, Special Paper, 369, 49-64.

Kuhn, T.S., 1970. The structure of scientific revolutions. Chicago, University Chicago Press, 210 pp.

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.

Molina, E., Canudo, J.I., Guernet, C., McDougall, K., Ortiz, N., Pascual, J.O., Parés, J.O., Samso, 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., Angori E., Arenillas I., Monechi S., Schmitz B., 2000. Integrated stratigraphy across the Paleocene/Eocene boundary at Campo, Spain. GFF - The Geological Society of Sweden (Geologiska Föreningen), 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.

Monechi, S., Angori, E., von Salis, K., 2000. Calcareous nannofossil turnover around the Paleocene/Eocene transition at Alamedilla (southern Spain). Bulletin de la Société Géologique de France, 171, 477-489.

Orue-Etxebarria, X., Pujalte, V., Bernaola, G., Apellaniz, E., Baceta, J.I., Payros, A., Núnez-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.

Payros, A., Pujalte, V., Baceta, J.I., Bernaola, G., Orue-Etxebarria, X., Apellaniz, E., Caballero, F., Ferrandez, 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.

Pomerol, Ch., 1975. La signification de l'Ilerdien et l'intérêt de cet étage dans la stratigraphie du Paléogène mésogéen. Bulletin de la Societé Géologique de France XVII, 213-217.

Pujalte, V., Schmitz, B., Baceta, J.I., Orue-Etxebarria, V., Nuñez-Betelu, K., Payros, A., Serra-Kiel, J., 1998. An Early Ilerdian transient switch-off in shallow and deep-water carbonate deposition in the western Pyrenees, Spain: origin and relevance for the Paleocene/Eocene boundary debate. Strata 9, 111-112.

Pujalte, V., Baceta, J.I., Payros, A., Orue-Etxebarria, X., Schmitz, B., 2000. Upper Paleocene-lower Eocene strata of the western Pyrenees, Spain: a shelf-to-basin correlation. GFF - The Geological Society of Sweden (Geologiska Föreningen), 122(1), 129-130.

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 d13C 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. The 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.

Raffi, I., Backman, J., Pälike, H., 2005. Changes in calcareous nannofossil assemblages across the Paleocene/Eocene transition from the paleo-equatorial Pacific Ocean. Palaeogeography, Palaeoclimatology, Palaeoecology, 226, 93-126.

Robador, A., 2005. El Paleoceno e Ilerdiense inferior del Pirineo occidental: Estratigrafía y sedimentología. Doctoral thesis. Universidad del País Vasco/Euskal Herriko Unibertsitatea, 456 pp.

Schaub, H., 1951. Stratigraphie und Paläontologie des Schlierenflysches mit besonderer Berücksichtigung der paleocaenen und untereocaenen Nummuliten und Assilinen. Schweizerische paläontologische Abhandlungen, 68, 1-222.

Schaub, H., 1966. Über die Grossforaminiferen im Untereocaen von Campo (Ober-Aragonien). Eclogae Geologicae Helvetiae, 59, 355-377.

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

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.

Scheibner, C., Speijer, R.P., Marzouk, A.M., 2005. Larger foraminiferal turnover during the Paleocene/Eocene thermal maximum and paleoclimatic control on the evolution of platform ecosystems. Geology, 33, 493-496.

Scheibner, C., Rasser M.W.M., 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.

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.

Schmitz, B., Speijer, R.P., Aubry, M.-P., 1996. Latest Paleocene benthic extinction event on the southern Tethyan shelf (Egypt): Foraminiferal stable isotopic (δ^{13}C, δ^{18}O) records. Geology, 24, 347-350.

Schmitz, B., Pujalte, V., Núnez-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.

Serra-Kiel, J., Canudo, J.I., Dinares, 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 de la Sociedad Geológica de España, 7 (3-4), 273-297.

Serra-Kiel, J., Hottinger, L., Caus, E., Drobne, K., Ferrandez, C., Jaurhi, A.K., Less, G., Pavlovec, R., Pignatti, J., Samso, 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 Societé géologique de France, 169, 281-299.

Smith, F.A., Wing, S.L., Freeman, K.H., 2007. Magnitude of the carbon isotope excursion at the Paleocene–Eocene thermal maximum: The role of plant community change. Earth and Planetary Science Letters, 262, 50-65.

Speijer, R.P., Van der Zwaan, G.J., Schmitz, B., 1996. The impact of Paleocene/Eocene boundary events on middle neritic benthic foraminiferal assemblages from Egypt. Marine Micropaleontology, 28, 99-132.

Stott, L.D., Kennett, J.P., Shackleton, N.J., Corfield, R.C., 1990. The evolution of Antarctic surface waters during the Paleogene: Inferences from the stable isotopic composition of planktonic foraminifers, ODP leg 113. In: Barker, P.F., Kennett, J.P., et al. (eds.). Proceedings of the Ocean Drilling Program, Scientific Results, 113, 849-863.

Svensen, H., Planke, S., Malthe-Sorenssen, A., Jamtveit, B., Myklebust, R., Eidem, T.R., Rey, S.S., 2004. Release of methane from a volcanic basin as a mechanism for initial Eocene global warming. Nature, 429, 542-545.

Tambareau, Y., 1994. Paleocene/Eocene boundary in the platform deposits of the northern Pyrenees. Bulletin de la Société belge de Géologie, 103, 293-299.

Tambareau Y., Villate, J., 1974. Le passage Thanétien-Ilerdien dans la región de Campo: comparaison avec les Petites Pyrénées. Bulletin Societé Histoire Naturel, Toulouse, 110, 340-361.

Thomas, E., 1990. Late Cretaceous-early Eocene mass extinction in deep sea. In: Sharpton, V.L., Ward, P.D. (eds.). Global Catastrophes in Earth History. Geological Society of America, Special Publication, 247, 481-495.

Thomas, E., 1998. Biogeography of the late Paleocene benthic foraminiferal extinction. In: Aubry, M.-P., Lucas, S., Berggren, W.A. (eds.). Late Paleocene-Early Eocene climatic and biotic events in the marine and terrestrial records. New York, Columbia University Press, 214-243.

Thomas, E., Shackleton N.J., 1996. The Paleocene-Eocene benthic foraminiferal extinction and stable isotope anomalies. In: Knox, R.W.O'B., Corfield, R.M., Dunay, R.E. (eds.). Correlation of the early Paleogene in Northwest Europe. Geological Society of London, Special Publication, 101, 401-441.

Thomas, D.J., Bralower, T.J., Zachos, J.C., 1999. New evidence for subtropical warming during the late Paleocene thermal maximum: Stable isotopes from Deep Sea Drilling Project Site 527, Walvis Ridge. Paleoceanography, 14, 561-570.

Thomas, D.J., Zachos, J.C., Bralower, T.J. Thomas, E., Bohaty, S., 2002. Warming the fuel for the fire: evidence for the termal dissociation of methane hydrate during the Paleocene-Eocene thermal maximum. Geology, 30, 1067-1070.

Wing, S.L., Harrington, G.J., Smith, F.A., Bloch, J.I., Boyer, D.M., Freeman, K.H., 2005. Transient floral change and rapid global warming at the Paleocene-Eocene boundary. Science, 310, 993-996.

Zachos, J., Pagani, M., Sloan, L., Thomas, E., Billups, K., 2001. Trends, rhythms, and aberrations in global climate 65 Ma to present. Science, 292, 686-693.

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.

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