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BilbaoTurismo
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Climate & Biota of the Early Paleogene

12th – 20th June
Bilbao 2006

POST-CONFERENCE FIELD EXCURSION GUIDEBOOK:

ZUMAIA SECTION

Climate & Biota of the Early Paleogene

Bilbao 2006

Organized by:

Department of Stratigraphy and Paleontology

Faculty of Science and Technology

University of the Basque Country

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ZUMAIARI ESKAINIA
etxean dauzkazuen altxor geologikoak
ezagutzen lagunduko duelakoan.
Zumaiako Udalbatzari esker onez,
emandako laguntzagatik.

DEDICATED TO ZUMAIA
in the hope it will contribute to discover
the geological treasures you have at home.
Many thanks to the Zumaia Town Council
for your kind support.

Aitzgorri Lurmuturreko Daniarreko kareharri gorriska eta zuriskaz dago eraikita Zumaiako San Pedro eliza.

The San Pedro church in Zumaia was built using the reddish and whitish Danian limestones that crop out in the Aitzgorri headland.

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5 THE PALEOCENE/EOCENE BOUNDARY INTERVAL

THE PALEOCENE AND LOWER EOCENE OF THE ZUMAIA SECTION (BASQUE BASIN)

Fig. 1.- Axel von Hillebrandt

CONTRIBUTION OF PROFESSOR AXEL VON HILLEBRANDT´S PIONEERING RESEARCHES TO PROMOTE ZUMAIA AS A GLOBAL REFERENCE SECTION

(*X. Orue-Etxebarria, E. Apellaniz*)

The lower Paleogene Zumaia section is one of the best-known and most studied land-based succesion in the world. This is partly due to the fact that this section contains a continuous and expanded record of the Paleocene, is easily accessible and is beautifully exposed in sea-cliffs. Furthermore, the position of the section between the Tethys and the North Sea makes it an important link between these two classical regions. But the success of Zumaia as a global reference section is also a result of the pioneering work on planktic foraminifera carried out by Professor von Hillebrandt more than 40 years ago.

Professor Axel von Hillebrandt was born on April 12th, 1933 in Tilsit (East Prusia) where he completed his secondary education studies in 1952. Later on, he undertook his university career choosing Geology at the universities of München and Freiburg between 1952 and 1957. This last year he obtained his Geology Diploma by the Technic University in München with the Diploma Thesis entitled *"Die Schichten der Gosau und des Alttertiärs im Nordem und Nordwestern des Untersberges (Becken von Reichenhall)"*.

In the summer of 1958 he took part in a scientific-mountaineering expedition to the southern Pyrenees (Ordesa Valley, Monte Perdido). Availing himself of that opportunity, he had his first contact with the cliffs in Zumaia, where he had come around to take sediment samples for micropaleontologic studies. The reason of this visit was the indication by his Diploma Guide Professor, Dr. P. Schmidt-Thomé. The latter, deeply inpressed by the superb outcrops in Zumaia, most probably knew the area from his dear friend J. Gómez de Llarena, local geologist that had already published some papers on the Cretaceus and Paleogene of Gipuzkua.

It was that very year, 1958, being camped next to the road going from Zumaia to Getaria with his friend Dr. Herm, geologist who worked in the upper Cretaceous between

Deba and the Aitzgorri headland at the same time Dr. Hillebrandt was researching in Zumaia, that he had the opportunity to meet Joaquín Gómez de Llarena, who came to visit them together with Professor Kockel (Marburg). Already in 1960 he took his Doctorate exam, with a Thesis entitled *"Das Paleozän und tiefere Untereozän im Becken von Reichenhall und Salzburg".*

Between 1960 and 1962 he held the position of Assistant Lecturer at the Institute of Paleontology and Historic Geology of the University of München and from October 1963 the same position at the Institute of Geology and Paleontology of the Technic University of Berlin. Also in 1963 he got married (was widowed in 1990), later on having two daughters, Barbara and Ingrid.

In 1961, 1962 and 1964 he visited the Southern Pyrenees and Zumaia with the aim of measuring stratigraphic profiles and taking samples for his micropaleontologic studies. The scientific results of these excursions in the Pyrenees were published in *Eclogae Geol.* Helv., 55/2, 1962 and later on translated into Spanish in *Notas y Comunicaciones del Inst. Geol. Y Min. de España, 73, 1964.*

After several articles mostly published in german journals, in 1965 appeared the work by Dr. Hillebrandt which granted Zumaia international renown in the Geology (Micropaleontology) field, entitled *"Foraminiferen-Stratigraphie im Alttertiär von Zumaya (Provinz Guipúzcoa, NW-Spanien) und ein Vergleich mit anderen Tethys-Gebieten.- Bayer. Akad. Wiss., math.-nat. Kl., 123, S.1-62, 3 Prof., 3 Kartenskizzen, 5 Tab., München".* Moreover, this work constituted the subject of his Thesis for qualification at the Technic University of Berlin and it also included the results of the studies carried out in Muskiz, Province of Navarre.

From 1965 onwards, professor Hillebrandt extened his micropaleontologic studies to the South of the Iberian peninsula, a fact that was later on reflected in several articles published in the *Rev. Española de Micropaleontología, Bull. De la Société Géologique de France,* etc., where a correlation between the biozones of planktic foraminifera and larger foraminifera was proposed. In 1967, using his excellent mountaineer and climber skills, which enabled him to sample even in the most adverse situations, he took part in a mountaineering and scientific expedition to the Central Mountain Range in Peru.

From 1966 to 1968 he was Professor in the Geology School of the University of Santiago de Chile and in 1969 he was appointed Professor of the Technic University of Berlin. His paleontologic research widened to include groups as varied as planktic foraminifera, larger foraminifera and finally ammonites. Between 1971 and 1997 he made several visits to Chile, Peru and Argentina in order to deal with his scientific studies, mostly dedicated to ammonites from the Pliensbachian of South America.

Throughout his long career he has supervised a number of Theses, both Diploma Theses and Doctoral Theses, carried out by students of the Insitute of Geology and Paleontology of the Technic University in Berlin. He has also been the Doctoral Thesis supervisor of J. Martínez (1977) and J. Cremades (1982), from the University of Granada, as well as of Xabier Orue-Etxebarria (1982), from the University of the Basque Country.

In the management area, both university and scientific, he has been part of several committees and international groups, having held, among others, the position of UNESCO's Consultant for Micropaleontology in the School of Geology of Central America

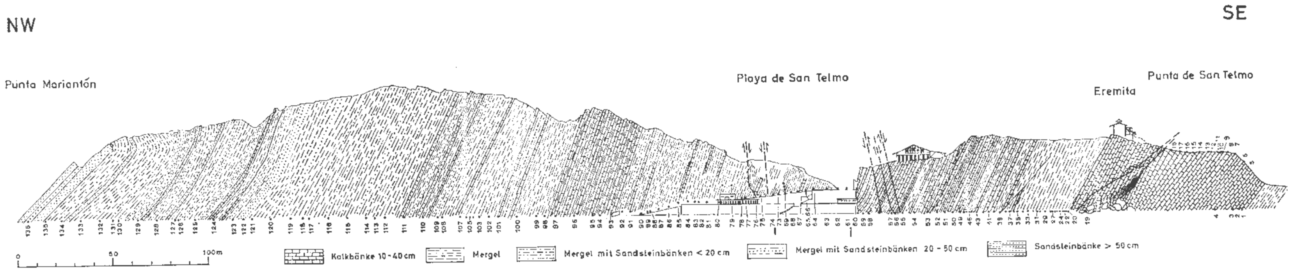


Fig. 2.- Zumaia cross section from Hillebrandt (1965).

at the University of Costa Rica. He was also Director of the Institute of Geology and Paleontology at the Technic University of Berlin from 1984 to 1987, Chairman of the German Subcommission on Jurassic Stratigraphy between 1988 and 1995, Chairman of "Paläontologische Gesellschaft" from 1995 to 1997, and finally retired in 1998.

It is clear that, in a summary like this, it is impossible to exhaustively describe all his works, which are close to a hundred, although it is necessary to mention that, even after his retirement he continues developing an active research, mainly dedicated to study ammonites from the lower Jurassic in South America. This fact has been reflected in the publication of monographies on ammonites from the Hettangian, Sinemurian and Pliensbachian.

Finally, we would like to highlight professor Hillebrandt's human aspect. Those of us who have been lucky enough to meet him, either while he stayed in Elche (Alicante) and Monte Perdido (Huesca) or during his participation in the 2nd International Conference Global Bioevents, held in Bilbao in 1987, appreciate his kindness in sharing with us his knowledge and vast experience. In our particular case, we must underline his behaviour as scientific director, as he always managed to find some time for discussing and contributed to the subject of Xabier´s Thesis. Furthermore, we will never forget his generosity when he hosted us in his home in Berlin in 1980, during the period in which we corrected the Doctoral Thesis. Since then, not only is he a master, but also a dear friend.

Dr. Hillebrandt, mila esker eta besarkada bero-bero bat.
(thank you and a huge hug. // viele Danke un liebe Grüβe.)

On behalf of the Organizing Committee, your friends
Xabier Orue-Etxebarria and Estibaliz Apellaniz.

The Paleocene and Lower Eocene of the Zumaia Section (Basque Basin)
Field-Trip Guidebook

1.- INTRODUCTION

1.1. The Upper Maastrichtian and lower Paleogene in the Pyrenean Basin (*G. Bernaola, J.I. Baceta, V. Pujalte*)

The Pyrenees are an unusual geologic area due to the thickness and cartographic extension of the Mesozoic and Cenozoic sediments. It was configured as an emerged area during the Mid Cenozoic (between 30-40Ma) as a result of the tectonic convergence undergone by the European and Iberian plates (Pujalte *et al.* 2002). Until the uplift of the Pyrenees the Pyrenean area constituted an open marine basin opened westward to the Proto-Bay of Biscay and North Atlantic. However, although the Pyrenean Basin was opened to the Atlantic Ocean it shared with the Tethys similar climatic conditions and biotic communities.

In the Pyrenean domain the upper Maastrichtian, Paleocene and Ilerdian (Lower Ypresian) are made up by an up to 500 m thick sedimentary succession representative of a wide-range of sedimentary environments, from continental to deep marine. The outcrops of these successions are extensive in both sides of the Pyrenean chain and in the Basque Basin, and their maping and facies analysis have allowed the elaboration of precise paleogeographic reconstructions (Fig. 3).

The upper Maastrichtian, Paleocene and Ilerdian (MPI) succession was deposited in a relative tectonic calm phase that precedes the principal orogenic phase (Puigdefàbregas y Souquet, 1986;

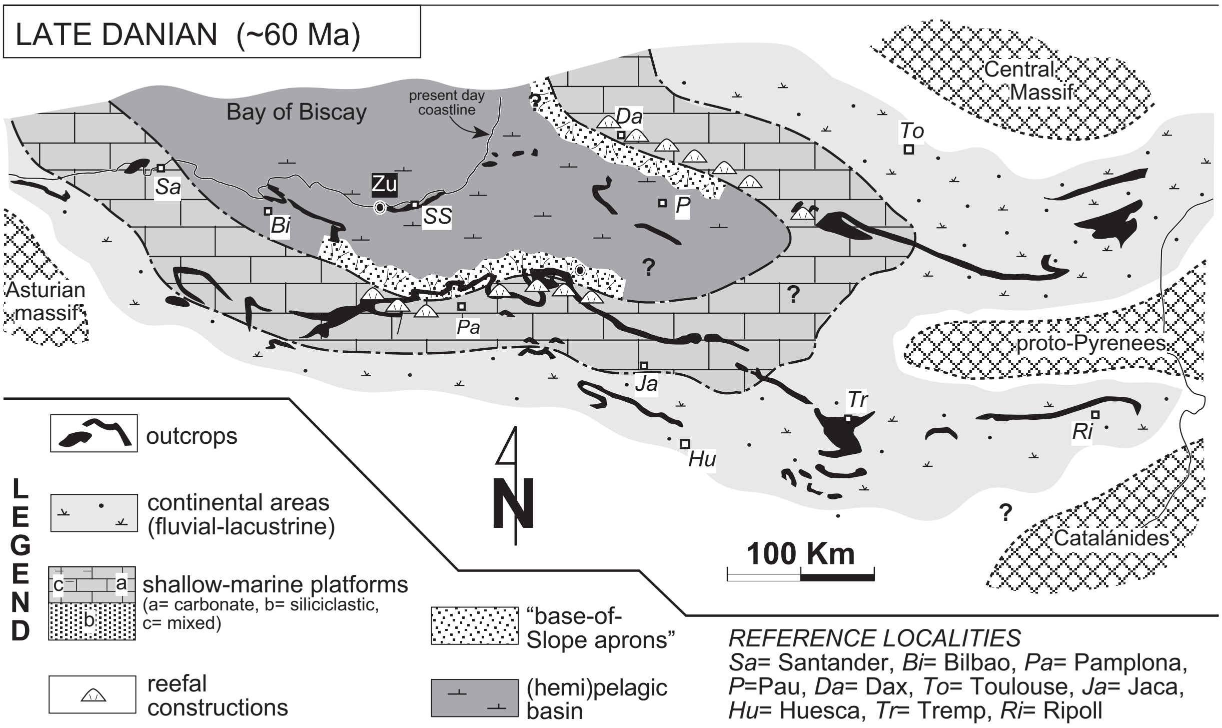


Fig. 3.- Late Danian (60 Ma) paleogeography of the Pyrenean domain, with indication of outcrops location and reference localities (after Baceta *et al.*, 2004).

Vergés, 1993; Baceta, 1996; Pujalte *et al.*, 2002; Muñoz, 2002). This interval is, moreover, characterised by a general transgressive trend, that together with the tectonic quiscence and a predominantly warm and semiarid climatic regimen (Medus & Colombo, 1991; Schmitz & Pujalte, 2003) brought out a substantial reduction of the siliciclastic input from the adjacent emerged areas and favoured the development of an extensive shallow marine carbonate platform system during the Paleocene and Ilerdian (Fig. 4). These platforms exceeded 50 Km in width and at some intervals developed bioconstructed margins (coralgal reefs) that coexisted with bioclastic shoals and, landward, with protected lagoons, carbonate beach systems and wide tidal plains. The pioneer studies about the Paleocene and lower Ypresian platforms of the Pyrenean basin were done by Mangin (1959-1960), von Hillebrandt (1964) and van de Velde (1967). From the later publications deserve to be stressed Garrido and Rios (1972), Plaziat (1981), Eichenser (1988), Betzler (1989), Barnolas *et al.* (1990), van der Hurk (1990), Luterbacher *et al.* (1991), Pluchery (1995), Serra-Kiel *et al.* (1994), Baceta (1996), Baceta *et al.* (2005).

The carbonates of the abovementioned platforms and the underlying upper Maastrichtian materials grade landwards to continental or transitional deposits, that have been classically named "Garum" or "Garumnian facies" (Leymerie, 1863; Rosell *et al.*, 2001), or Trem Group (Cuevas, 1992).

Seawards, the MPI platform systems evolved to slope areas characterised by a discontinuous sedimentary record and accumulation of breccias, conglomerates and carbonatic turbidites derived from the adjacent platforms (Plaziat, 1975; Baceta *et al.*, 1991; Baceta, 1996; Pujalte *et al.*, 1998a, b). Overall, the slope zones demarcate a central deep basin area (Basque Basin), with an estimated water depth of about 1000 m (Pujalte *et al.*, 1998) whose deposits have been principally preserved in the Basque Arc (Bizkaia and Gipuzkoa) and in the North Pyrenean Zone (Fig. 3). In the Basque Basin the sedimentation during the MPI was predominantly hemipelagic, and as a result laterally extensive successions composed of an alternation of marls, marly limestones and limestones were deposited. Some of the sections that contains these facies (mainly Zumaia, Gipuzkoa) are global reference sections for this time interval as they have been the subject of multiple studies about late Cretaceous ammonites and inoceramids (e.g. Wiedmann, 1988; Ward *et al.*, 1991; Ward & Kennedy, 1993), late Maastrichtian-lower Paleogene calcareous plankton (e.g. von Hillebrandt, 1965; Orue-Etxebarria, 1983; Arenillas, 1996; Bernaola, 2002), magnetostratigraphy and cyclostratigraphy (Roggenthen, 1976; Ten Kate & Sprenger, 1993; Pujalte *et al.*, 1995; Dinarès-Turell *et al.*, 2002, 2003) and furthermore because they offer excellent sections of some significant chronoestratigraphic boundaries such as the Cretaceous/Paleogene (Lamolda *et al.* 1988; Lamolda & Gorostidi, 1992; Apellaniz et. al, 1997; Apellaniz, 1998; Arenillas *et al.*, 1998) Danian/Selandian (Schmitz *et al.*, 1998), Selandian/Thanetian or Paleocene/Eocene (Canudo *et al.* 1995; Schmitz *et al.*, 1997; Baceta *et al.* 2000; Dinarès-Turell *et al.*, 2002; Orue-Etxebarria *et al.*, 2004).

1.2 The Zumaia section (*J.I. Baceta, V. Pujalte, G. Bernaola, X. Orue-Etxebarria, E. Apellaniz*)

The Zumaia section, located in the coast of the Gipuzkoa province halfway between the cities of Bilbao and San Sebastian, is an essentially continuous lower Santonian to lower Eocene succession that crops out along a series of sea-cliffs and beaches. The MPI of the Zumaia section (latitude/longitude 42°18.00'N /2°15.30'W) is one of the best-known and most studied worldwide. In fact the section contains a complete and very expanded record of the Paleocene, is easily accessible and is beautifully exposed at the Aitzgorri headland seacliffs and the contiguous Itzurun beach (Fig. 5). Furthermore, the position of the section between the Tethys and the North Sea makes it an important link between these two paleogeographical domains.

Due to its easy accessibility and superb quality of exposure, the Zumaia section already attracted the attention of pioneer workers in the region (e.g., Gómez de Llarena, 1946). It was later the sub-

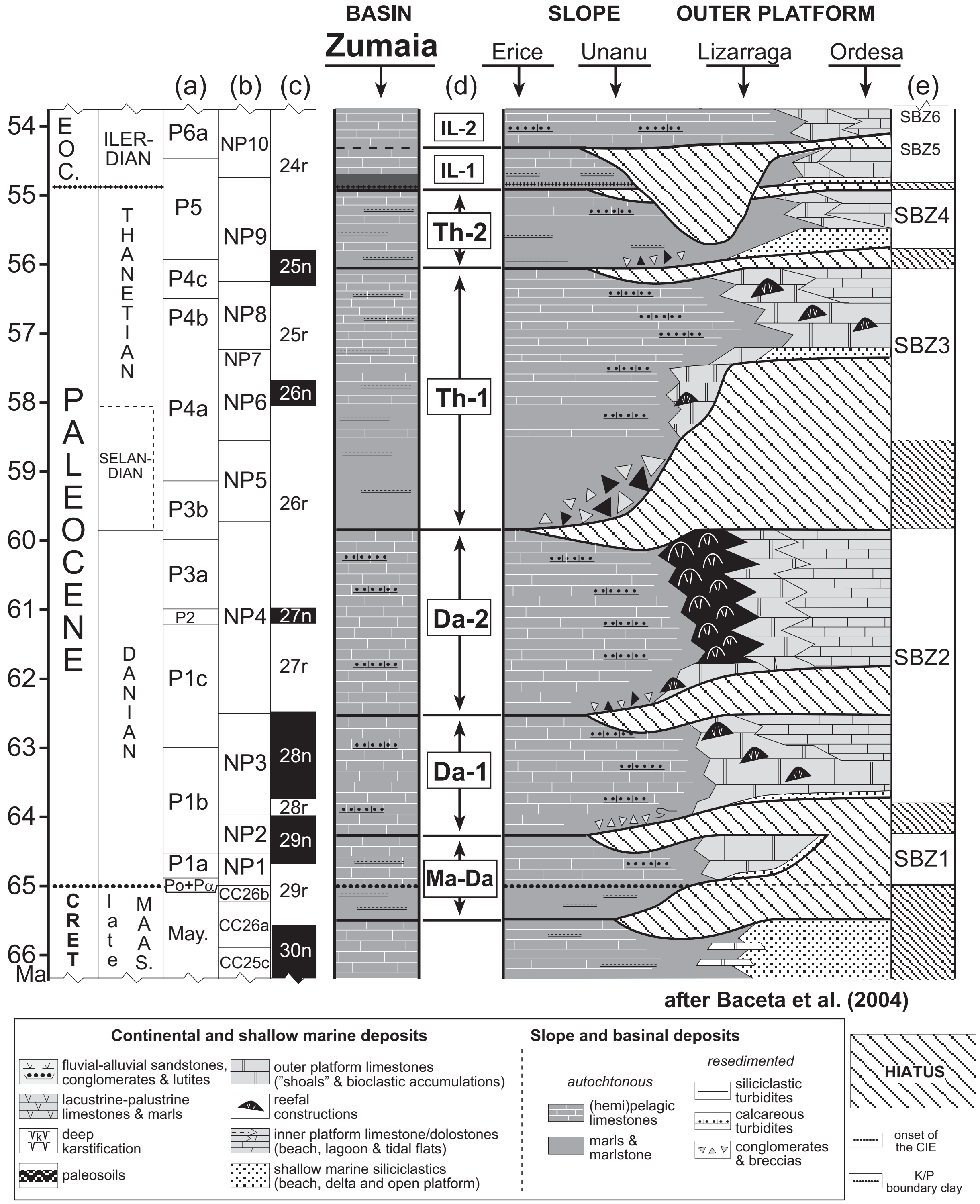


Fig. 4.- Biochronostratigraphic framework of the Paleocene-lower Ilerdian succession of the SW Pyrenees, showing depositional sequences and main facies. (modified from Baceta *et al.*, 2004).

ject of general studies about planktic foraminifers (Hillebrandt, 1965), calcareous nannofossils (Kapellos, 1974; Van Vliet, 1982) or sequence stratigraphy (Baceta, 1996; Pujalte *et al.*, 1998a), to mention but a few. Several other papers have focused on the analysis of the Cretaceous/Paleogene (K/P) and the Paleocene/Eocene (P/E) boundaries including Alvarez *et al.* (1982), Smit and Ten Kate

(1982), Wiedman (1986), Canudo and Molina (1992), Gorostidi (1993), Kunht and Kaminski (1993), Ward and Kennedy (1993), Canudo *et al.* (1995), Ortiz (1995), Schmitz *et al.* (1997, 1998), Knox (1998), Apellaniz (1999), Adatte *et al.* (2000), Bernaola (2002), Orue-etxebarria *et al.*, (2004).

During the last decade the Zumaia section was considered a suitable section as the boundary stratotype for the K/P and P/E boundaries and now it has been proposed for the definition of the Selandian and Danian stages (see www.stratigraphy.org/gssp.htm). The Zumaia section, in effect, complies with most of the requirements listed by the International Commission on Stratigraphy (ICS, see Remane *et al.*, 1996), namely: (a) it posses an adequate thickness of exposed sediments, since the Paleocene interval in Zumaia is in fact but a small segment of a near-continuous section spanning from the Campanian to the lowest Lutetian; (b) no biostratigraphic gap or condensation are recognizable; (c) a high sedimentation rate, Zumaia being in fact one of the most expansive open marine successions of the Paleocene so far reported; (d) identifiable geochemical and geomagnetic signals (Schmitz *et al.*, 1997 and below), a clear proof of minor diagenetic alteration; (e) abundant, well diversified open marine fossils, including planktic foraminifers and calcareous nannofossils; (f) good exposure and easy accessibility.

The lower Paleogene sediments of Zumaia were deposited at an estimated water depth of about 1000 m (Pujalte *et al.* 1995, 1998). The upper Maastrichtian, Paleocene and Ilerdian at Zumaia are represented by rythmic alternations of hemipelagic limestones and marlstones, plus some intercalations of thin bedded carbonate, siliciclastic and mixed turbidites. This lithologic cyclicity has been attributed to orbital forcing by different authors (Ten Kate and Sprenger, 1993; Dinarès-Turell *et al.* 2002, 2003).

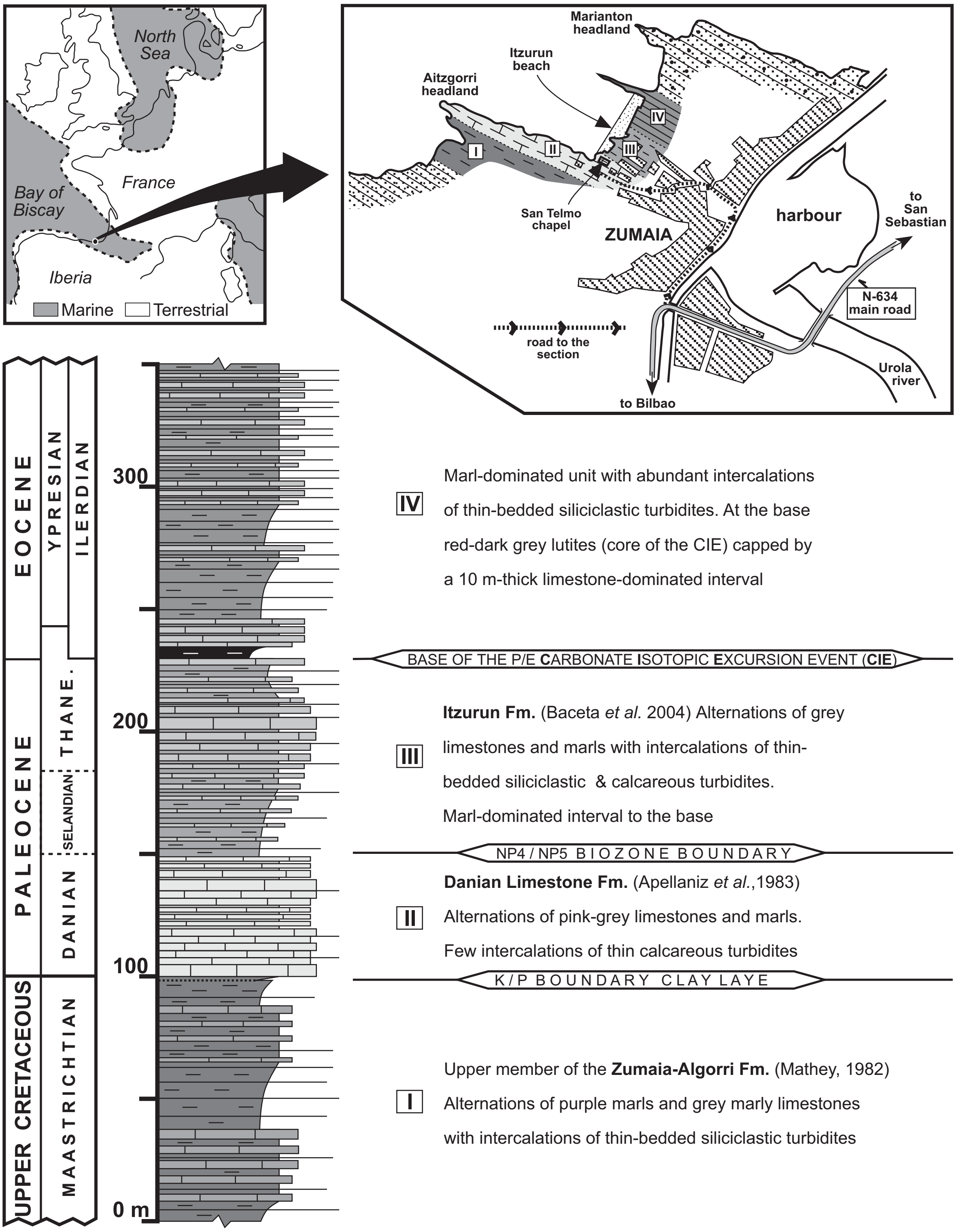
The MPI succession at Zumaia is arranged into four main lithological units that are easy to recognise in the field: The Zumaia-Algorri Fm., the Danian Limestones Fm., the Itzurun Fm. and The Eocene Flysch (Fig. 5).

I) Zumaia-Algorri Fm. (Mathey, 1982). This unit reaches a thicknes of ~200 m and encompasses the Middle-Upper Maastrichtian. The upper member, which includes the upper Maastrichtian, is characterised by alternations of purple marls and grey marly limestones with intercalations of thin-bedded siliciclastic turbidites. The Cretaceous/Paleogene (K/P) clay marks the boundary between the Zumaia-Algorri Fm. and the overlying Danian Limestones Fm.

II) Danian Limestone Fm. (Apellaniz *et al.*,1983). This is composed of a 49 m-thick alternation of pink-grey limestones and marls with few intercalations of thin calcareous turbidites. According to Pujalte *et al.* (1998) this unit represented an stage of low sedimentation rates across the whole Basque Basin. Its boundary with the overlying Itzurun Fm. is defined by a sharp drop in the total carbonate content from 75% to less than 25%.

III) Itzurun Fm (Baceta *et al.*, 2004). It is a 80 m thick alternation of grey limestones and marls with intercalations of thin-bedded siliciclastic and calcareous turbidites. At the base of this formation there is a marl-dominated interval. The carbonate content increases upwards, reaching its maximum at the middle-upper part of the unit, and then decreases progressively to research a minimum at the P/E boundary. The turbidite content is higher in the upper part of the Itzurun Fm.

IV) Eocene Flysch: This interval is a marl-dominated unit with abundant intercalations of thin-bedded siliciclastic turbidites. At the base of this unit there are is a 4 m thick distinct interval of red-dark grey lutites that represents the important negative $\delta^{13}C$ shift of the P/E boundary.



IV Marl-dominated unit with abundant intercalations of thin-bedded siliciclastic turbidites. At the base red-dark grey lutites (core of the CIE) capped by a 10 m-thick limestone-dominated interval

III **Itzurun Fm.** (Baceta *et al.* 2004) Alternations of grey limestones and marls with intercalations of thin-bedded siliciclastic & calcareous turbidites. Marl-dominated interval to the base

II **Danian Limestone Fm.** (Apellaniz *et al.*,1983) Alternations of pink-grey limestones and marls. Few intercalations of thin calcareous turbidites

I Upper member of the **Zumaia-Algorri Fm.** (Mathey, 1982) Alternations of purple marls and grey marly limestones with intercalations of thin-bedded siliciclastic turbidites

Fig. 5.- Generalised early Paleogene paleogeography of Western Europe and geologic map of the Upper Cretaceous-Lower Paleogene outcrops in the Zumaia area. Synthetic lithologic section of the uppermost Cretaceous-lower Paleogene inteval of the Zumaia section, showing four main stratigraphic units.

2.- THE CRETACEOUS/PALEOGENE (K/P) BOUNDARY

2.1 The K/P boundary and the Pyrenean Basin (*G. Bernaola, E. Apellaniz, X. Orue-Etxebarria*)

The K/P boundary is one of the best and most studied mass extinction events in the geological record. Although it has been the subject of multiple high resolution analyses over the last three decades, there are still some unresolved aspects related to the type and causes of the extinction. In general, two basic extinction patterns have been proposed, a gradual one starting below or just above the boundary and an abrupt or mass extinction. The identification of the extinction pattern is essential to the determination of its causes. According to Pospichal (1996b), a gradual or stepwise extinction might have resulted from long term climatic and/or sea-level changes (Clemens *et al.*, 1981; Hallam, 1987) or from an extended period of intense volcanism that modified the composition of the atmosphere (McLean, 1982, 1985; Officer & Drake, 1983, 1985; Officer *et al.* 1987), whereas an abrupt mass extinction might have resulted from a catastrophic event such as an impact of a large asteroid (Alvarez *et al.*, 1980, 1984) or an intense volcanism (Courtillot *et al.*, 1988).

Presently, there is no doubt about the presence of a extraterrestrial body impact in connection with the K/P boundary, but total agreement does not exist among the authors to determine if there was one or several impacts (Keller *et al.*, 2004 and references therein) and if the extraterrestrial cause was the unique and/or principal reason of the biotic crisis. A mix of coincident causes such as an intense period of volcanism and long term climatic and/or sea-level changes could have given rise to a gradual extinction that could have reached its climax with one or several meteoritic impacts.

In the Pyrenean Basin the K/P boundary is preserved in numerous deep marine sections distributed in three major bands of outcrops: the Bilbao area, the San Sebastian area, and the Pamplona area (Fig. 3). In addition to these, scattered outcrops also occur in southern Aquitaine (Pau area). Some of the Basque sections have been the focus of palaeontological, geochemical, magnetostratigraphic and sedimentological studies including many detailed micropalaeontologic analyses. Indeed, several of them are already classic, notably those of Zumaia (Percival and Fischer, 1977), Bidart (Perch-Nielsen, 1979; Delacotte *et al.*, 1985), Hendaia (Seyve, 1990) or Sopelana I (Orue-etxebarria, 1982; Lamolda *et al.*, 1983) as well as the lesser known from Monte Urko (Orue-Etxebarria *et al.*, 1991).

In Zumaia section as in all deep-marine sections of the Pyrenean basin, the K/P boundary separates two different groups of rocks. The rocks below the boundary are mostly reddish to purplish marls and marlstones (Upper Member of the Zumaia-Algorri Formation, Mathey, 1986, also referred to as "Maastrichtian marls"), whereas the rocks above it correspond to a group of grey and pink pelagic limestones and marlstones that are called "Danian Limestones" (Apellaniz *et al.*, 1983) (Figs. 6a, 6c). Sequence stratigraphic analyses have established that the K/P boundary is included within a single depositional sequence (Ma-Da Depositional Sequence in Figure 4, see also Baceta, 1996 and Pujalte *et al.*, 1997). In fact, the K/P boundary is included within the transgressive systems tract (TST) and in most sections corresponds to a period of very low sedimentation rates (condensed interval). At Zumia the TST spans the upper 2 m of the Cretaceous and the lower 3 metres of the "Danian Limestones".

2.2. Lithostratigraphy of the K/P boundary interval *(J.I. Baceta, V. Pujalte)*

Due to the dominant hemipelagic nature of the sedimentation in the Basque Basin, the lithostratigraphy of the various K/P boundary sections, including Zumaia, is quite similar.

Our observations demonstrate that the composite detailed stratigraphy of the K/P boundary in the Basque Basin includes 5 well-defined lithologic intervals (Fig. 7), some of which may be locally missing due to tectonic deformation during the Alpine Orogeny. From base to top, the lithologic succession is as follows.

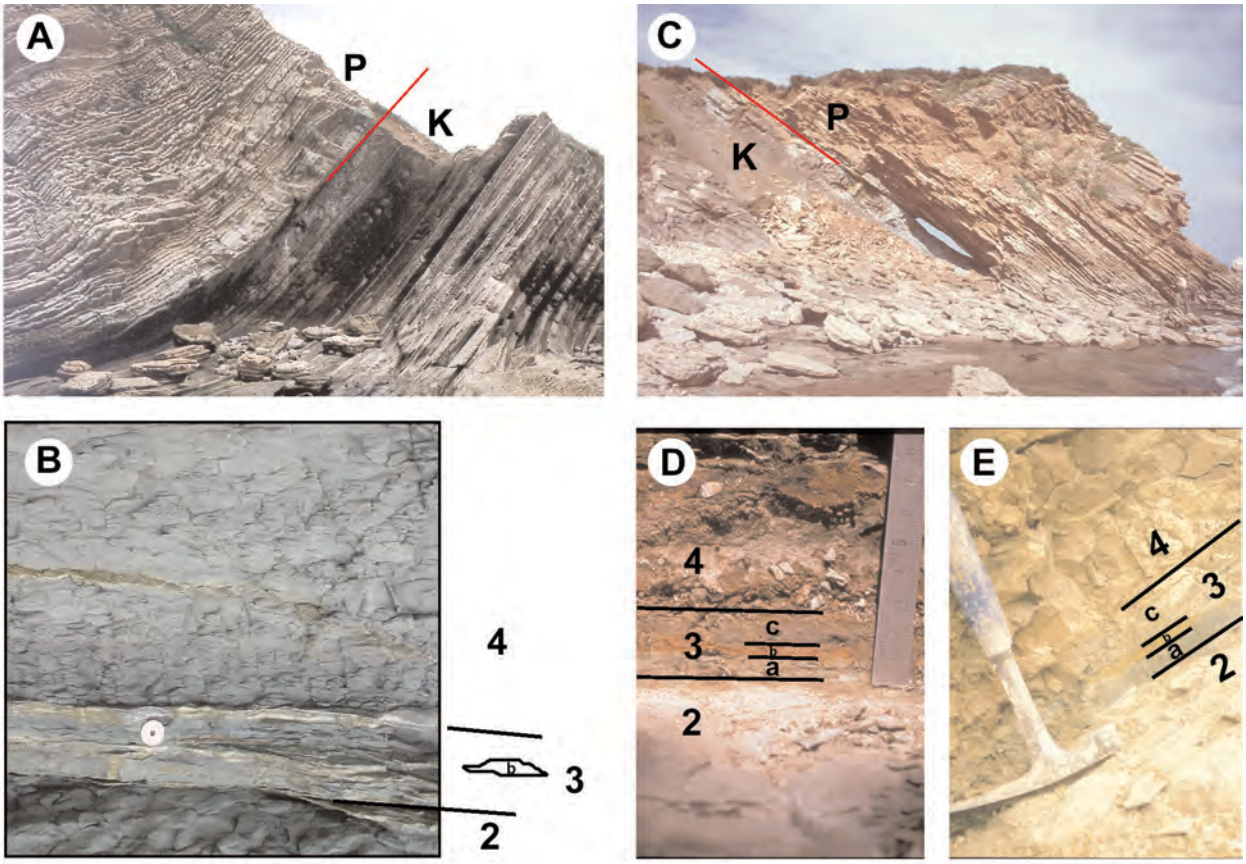


Fig. 6.- Field photographs showing different aspects of the K/P boundary transition in several Basque Basin sections. (A-C) General view of the K/P boundary at Zumaia and Hendaia sections. (B,D,E) Detailed view of the K/P boundary clay in Zumaia, Urrutxua and Bidart sections.

- ***Interval 1*** (upper Maastrichtian, *Abathomphalus mayaroensis* biozone; *Micula murus* and *Micula prinsii* biozones). It includes 20-30 cm thick couplets of compact, grey and reddish marls and marlslones. In Zumaia and other sections such as Urrutxua, and Herrera there are intercalations of thin, mixed carbonate-siliciclastic turbidites with silt/fine sand sized detritic quartz (50-65 %) and tests of planktic foraminifers (10-20%) embedded in a matrix of micritic mud.
- ***Interval 2*** (uppermost Maastrichtian, uppermost part of the *A. mayaroensis* biozone and *M. prinsii* biozone). It is mainly characterized by a gradual but rapid decrease in the carbonate content. Calcareous plankton is abundant in this interval.
- ***Interval 3*** (K/P boundary clay). It includes a clay level that varies in thickness between 1 and 8 cm. These differences are partly the result of interbed sliding, as demonstrated by the presence of calcite slickensides in some sections . Three layers can be differentiated within this clay bed (Fig. 6d, 6e,7). The lowest one (a) is a grey or yellowish silty bed that usually contains planktic foraminifer tests with indications of dissolution. The intermediate layer (b) is a reddish to brownish siltstone where microspherules with Ni-rich spinel crystals are significantly present. The uppermost layer (c) includes thinly laminated, dark grey siltstones. The highest iridium anomaly occurs at the contact between the upper two layers. At Zumaia the boundary clay is not laterally continuous due to the presence of calcite slickensides and it is only is visible in small patches (Fig. 6b). At Zumaia and other Basque Basin K/P sections features of impact deposits ocurr: iridium, Ni-rich spinels and soot (Rocchia *et al.*, 1996). The strongest peak of iridium has been found at Urrutxua section with almost 29 ppm. In the sediments

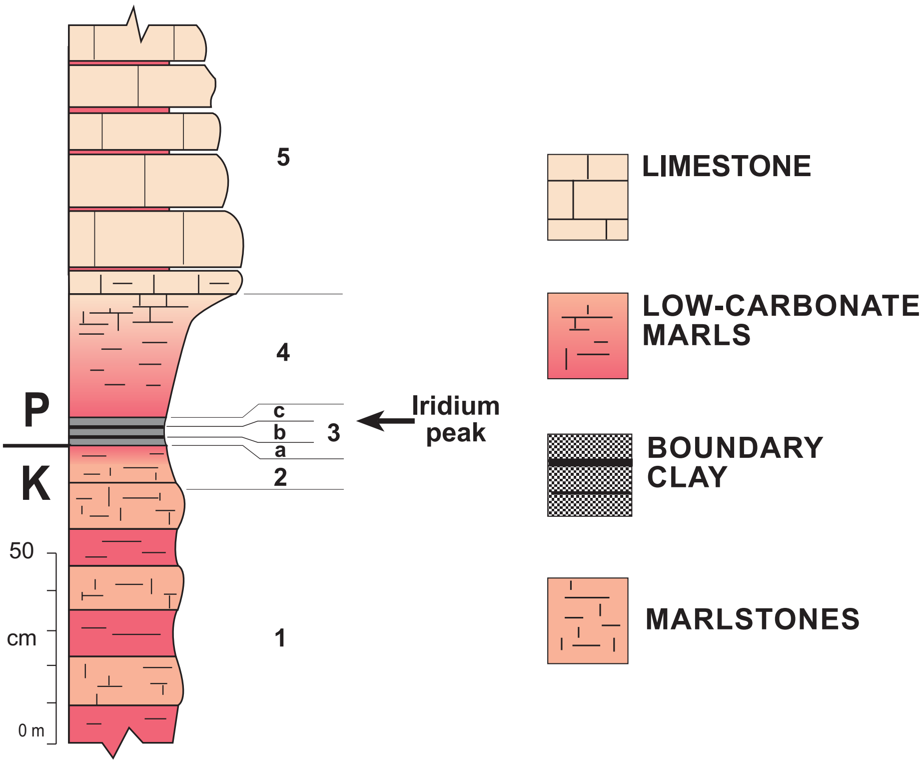


Fig. 7.- Composite detailed stratigraphy of the K/P boundary at the Basque Basin including 1 to 5 lithological intervals.

of the Sopelana section, there is a strong concentration of soot in the first centimetres of the Danian. All these data support the existence of, at least, one impact at the K/P boundary, coinciding with the extinction of the calcareous plankton.

- ***Interval 4*** (lowermost Danian, *Gb. cretacea* and *Civisina euskalherriensis* biozones; *Biantholithus sparsus* biozone). It corresponds to a succession of grey marls that display an upward increase in carbonate content, becoming marlstones in the upper part. In Zumaia the fist 4 cm of this interval are of a darker colour.

- ***Interval 5*** (lower Danian, *Parasubbotina pseudobulloides* biozone; *Biantholithus sparsus* biozone). It is mostly made up of hemipelagic limestones with minor intercalations of hemipelagic marls. In the Zumaia section there are also a few turbidites of mixed composition. These turbidites still yield a significant number of Cretaceous calcareous nannofossils and planktic foraminifers, although Danian taxa are already the major components of the assemblages. They also contain glauconite (5-10% of the total) both as small rounded detritic grains, and authigenic, filling or partly replacing the tests of the planktic foraminifers.

2.3. Calcareous nannofossil turnover across the K/P boundary (G. Bernaola)

It the K/P boundary transition of Zumaia section a total of 88 calcareous nannofossil species has been recognised: 68 cretaceous, 6 Persistent and 14 Paleocene (Fig. 8). According to preservation criteria proposed by Roth and Thierstein (1972) all the samples yielded moderatelly preserved calcareous nannofossil assemblages. The total abundance is variable, relatively high in the upper Maastrichtian samples, from 3,27 to 7,1 spp./f.v. (a mean of 5,74 spp./f.v.), and much lower in the lower Danian, ranging from 0,65 to 2,38 spp./f.v. (a mean of 1,04 spp./f.v.)(Fig. 9).

The upper Maastrichtian assemblage is characterised by high total abundance and species richness, a mean of 5,75 spp./f.v. and 39-50 species per sample (S-H of about 2,8). The assemblage is dom-

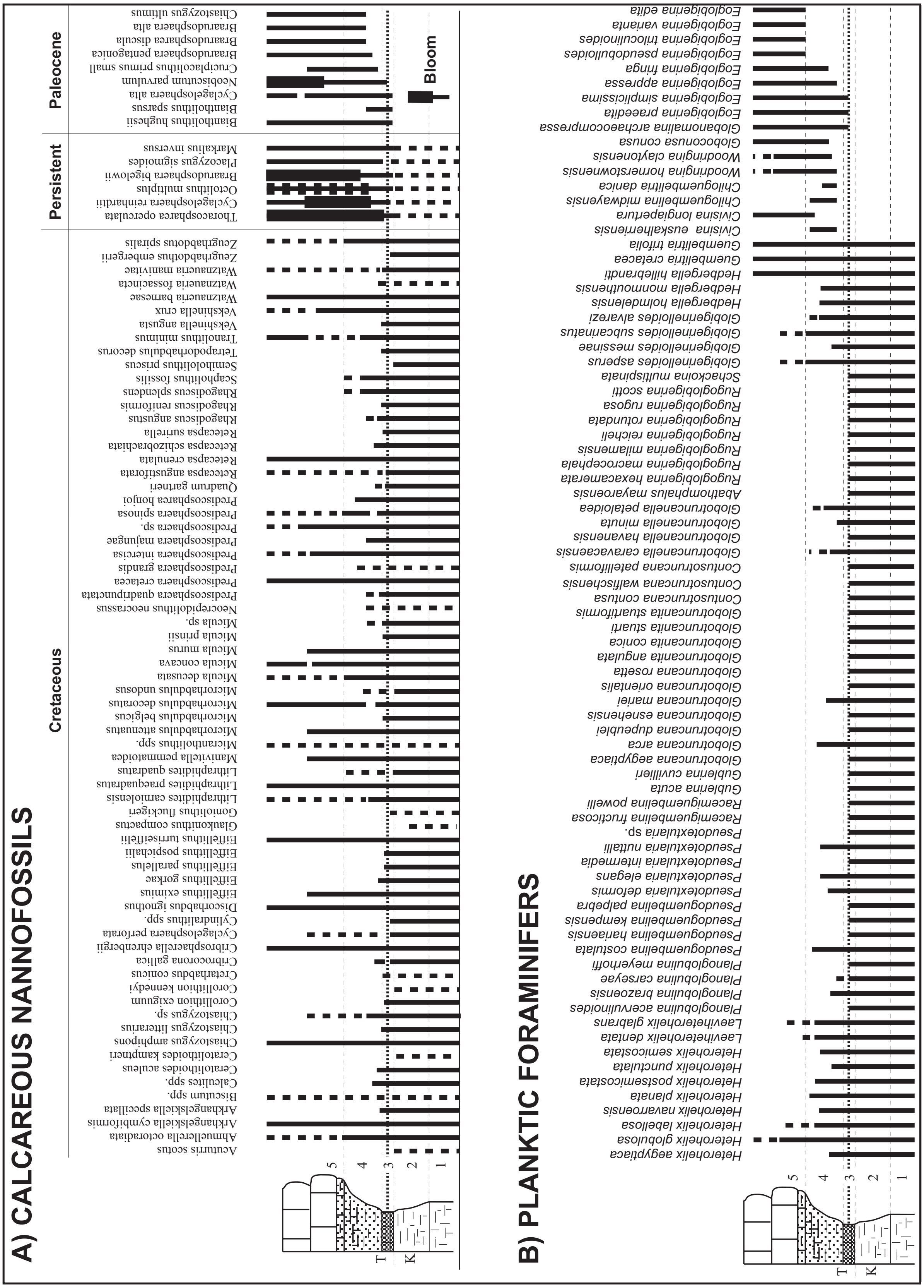


Fig. 8.- Biostratigraphic distribution of calcareous nannofossil and planktonic foraminifers for the uppermost Cretaceous and lowermost Paleogene from the Zumaia section.

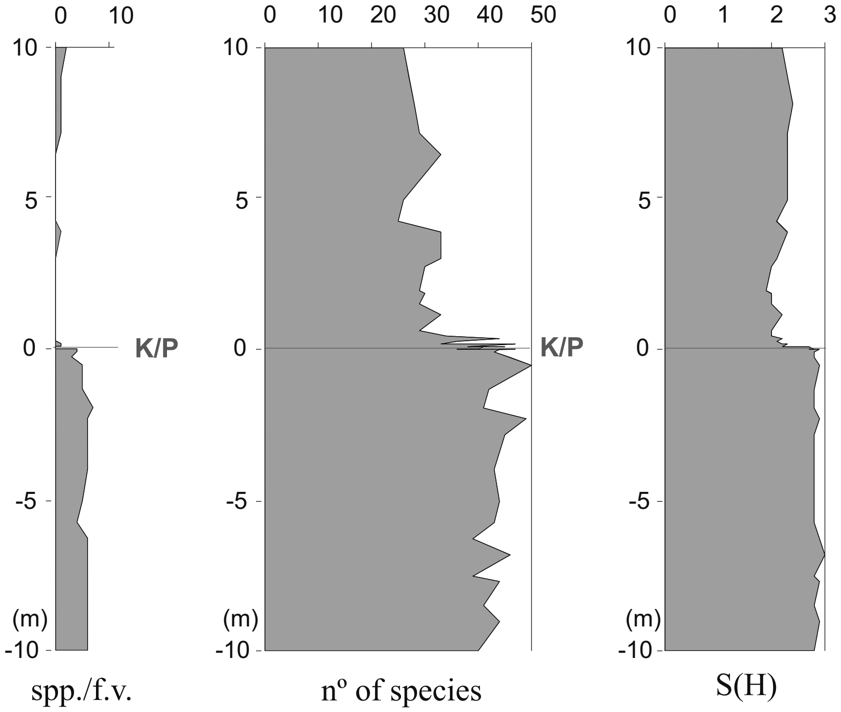


Fig. 9.- Calcareous nannofossil total abundance and species richness (spp./f.v., S-H) across the K/P boundary interval at Zumaia section.

inated by more than 99% of cretaceous species the persistent ones being very rare (Fig. 10). The most abundant species, that represent 70-80% of the total assemblage, are *W. barnease*, *Cr. ehrenbergii*, *M. decussata*, *Pr. cretacea*, *Pr. intercisa*, *Retecapsa* spp., *Ze. spiralis*, *Ark. cymbiformis*, *Eiff. turriseifelii* and *Ah. regularis*. The remaining 20-30% is composed by those species that never reach the 2-3%, such as *M. attenuatus*, *Cy. serratus*, *Chiastozygus spp.*, *Pr. spinosa*, *M. murus* and *M. prinsii*, the latest the marker of the base of the *Micula prinsii* biozone. The thickness of the *M. prinsii* biozone, which characterises the terminal Maastrichtian in complete sections, changes among the Basque Basin sections. It is 6 m thick at Zumaia but it is just few centimetres in Monte Urko section (Gorostidi and Lamolda, 1991), and up to 10 m in Bidart (Bernaola 2002). All the sections have a similar sedimentation rate and the difference in the thickness of this biozone in different sections is due to a taphonomic control.

In connection with the lithological change between horizon 2 and 3 an important turnover in calcareous nannofossil assemblages is registered. The calcareous nannofossil total abundance suddenly drops from 4,76 to 0,77 spp/f.v. (a reduction of the 84%) and the diversity also shows a marked decrease (Fig. 9). In connection with this lithological change a slight increase in the relative abundance of *Thoracosphaera* spp. is also observed, from very rare to about 3% of the total assemblage (Fig. 11). However, the calcareous nannofossil assemblage is very similar to that found in the underlying sediments and the Cretaceous species are the dominant group (Fig. 10). During the last two decades the massive occurrence or increased frequency of *Thoracosphaera* spp. has been widely used to define the K/P boundary.

Three centimetres above the lithological change the relative abundance of *Thoracosphaera* spp. increases abruptly, reaching the 30% of the total assemblage. At Zumaia section *Thorocospharea* spp. is not the only persistent species that blooms above the K/P boundary. Following the bloom of

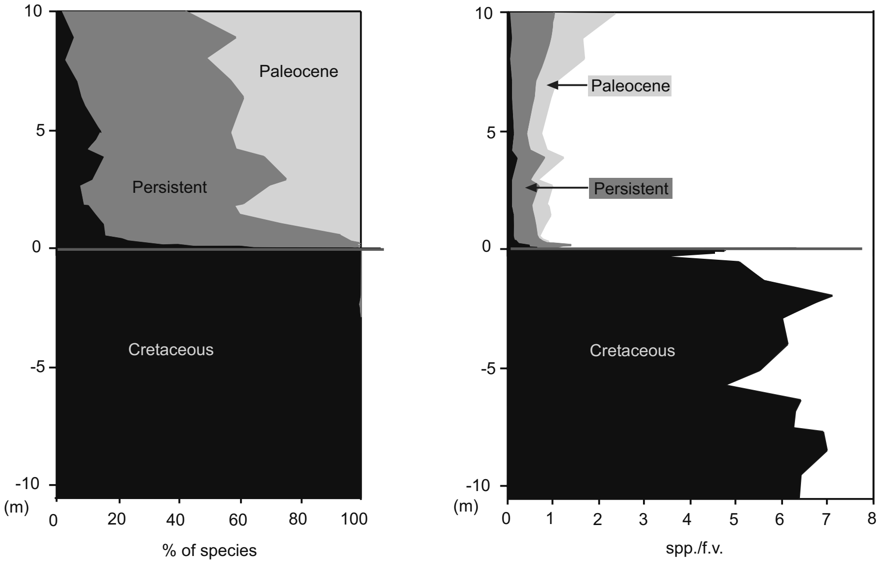


Fig. 10.- Changes in relative (%) and absolute (spp./f.v.) abundance of Creataceous, Persistent and Paleocene species across the K/P boundary interval at Zumaia section.

Thorocospharea spp. the blooms of other persistent species such as *Cy. reinhardtii*, *Oct. multiplus*, *Br. bigelowii* are registered (Fig. 11). The group of the persistent species is completed by *Pl. sigmoides* and *M. inversus* that never reach 5% of the total assemblage.

12 cm above the lithological change that marks the K/P boundary (boundary between intervals 2 and 3) the assemblage is dominated by persistent species. This calcareous nannofossil assemblage dominace change is very fast (Fig. 10).

After an abrupt decrease across the first 25 cm of the Paleocene, the relative abundance of cretaceous species decreases progressively, being rare 1m above the K/P boundary. At Zumaia the dominace of the persistent species extends up to 7 m above the K/P boundary, where the new Paleocene species become the dominant species (Fig. 10).

Several new Paleocene species have their First Occurrence (FO) in the first meter of the Paleocene (Fig. 12). *B. sparsus* and *B. hughesii* appear just above the Ir anomaly, followed by the FO of *Br. discula, Br. alta*, *Cy. alta, Ch. ultimus Cr. primus* small and *Ne. parvulum*. *Ne. romeinii*, very abundant in low latitude sections such as El Kef, is not present in Zumaia section. Following the blooms of the persistent species the blooms of the some diminute Paleocene species, such as *Ne. parvulum Fu. petalosa*, *Pr. dimorphosus* and *Pr. martinii,* are registered (Fig. 11).

The understanding of the calcareous nannofossil extinction model and the subsequent recovery over the K/P transition largely depends on the interpretation of the Cretaceous species found in Danian sediments. In the present study the persistent species are the unique forms that are considered confidently survivors. It is not possible to prove or disprove unequivocally the survivorship of most vanishing species by means of the criteria proposed by previous authors. Some of the Cretaceous species could have survived for a short time but they were not able to adapt to the hostile environment and changes in ocean circulation related to the K/P event(s) and did not serve as ancestors to Paleogene species.

After the K/P event(s) the persistent species suddenly become the dominant group of the non-reworked assemblage. Their total abundance, however, is low along the first meters of the Danian, suggesting a persistence of low nannofossil carbonate production for at least the first million years.

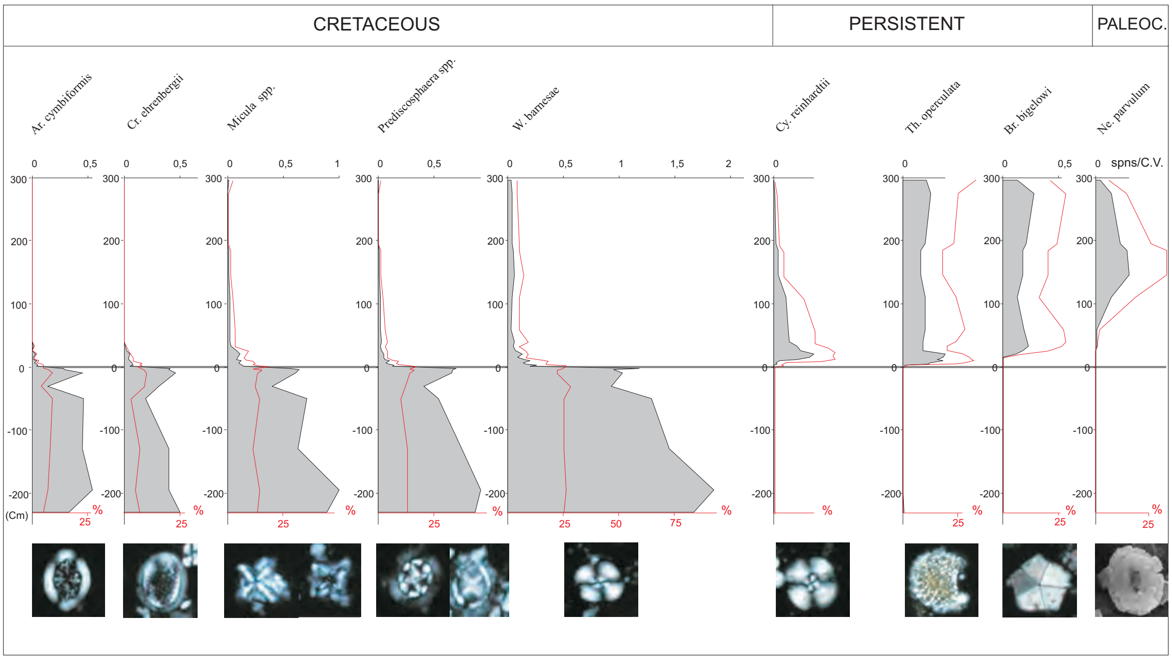


Fig. 11.- Relative (red lines) and absolute (grey shadow) abundances of selected calcareous nannofossil taxa across the K/P boundary interval at Zumaia section.

In agreement with Bown (2005) we consider the persistent species, mainly r-selected, opportunistic taxa, which took up progressively the ecologic niches that were previously occupied by dominant Cretaceous K-selected, specialist species, undergoing in many cases a significant geographical expansion. However, their low total abundance suggests that they survived in a still hostile ecosystem and/or they changed their life-cycle strategies i.e. as naked coccolithophorids.

The persistent species keep their dominance in the calcareous nannofossil assemblages up to 7 m above the boundary. The beginning of an increase in the calcareous nannoplankton total abundance recorded one meter above the K/P boundary together with the increase in abundance of the cited new-Paleocene species mark the beginning of the calcareous nannoplankton communities' recovery. However, the relatively high abundance of persistent taxa higher up in the succession indicates that the total recovery of the calcareous nannoplankton communities did not take place until the upper part of Danian.

*2.4. **Planktic foraminifera turnover across the K/P boundary*** *(E. Apellaniz, X. Orue-Etxebarria, H. Luterbacher)*

The interpretation of the planktic foraminifer turnover across the K/P boundary varies between the researchers. For instance, some authors perceive a gradual decrease of the Cretaceous species (Keller, 1988), whereas others argue in favour of a sharp extinction event (Smit, 1982).

The Zumaia section, together with other Basque Basin K/P boundary sections, has become essential to shed light on this controversy.

In this section, where the planktic foraminifer to total foraminifer assemblage ratio is higher than 90%, the planktic foraminifera assemblage preservation is good and the total abundance high. In the upper Maastrichtian the planktic foraminifers abundance is between 50,000 and 90,000 per gram of dry processed sample in the fraction 100 to 500 µm. Lower Paleocene samples are much poorer and their content of the 40-100 µm fraction increases progressively upward from the base of the Paleocene varying between 2,500 and 11,000 planktic foraminifers.

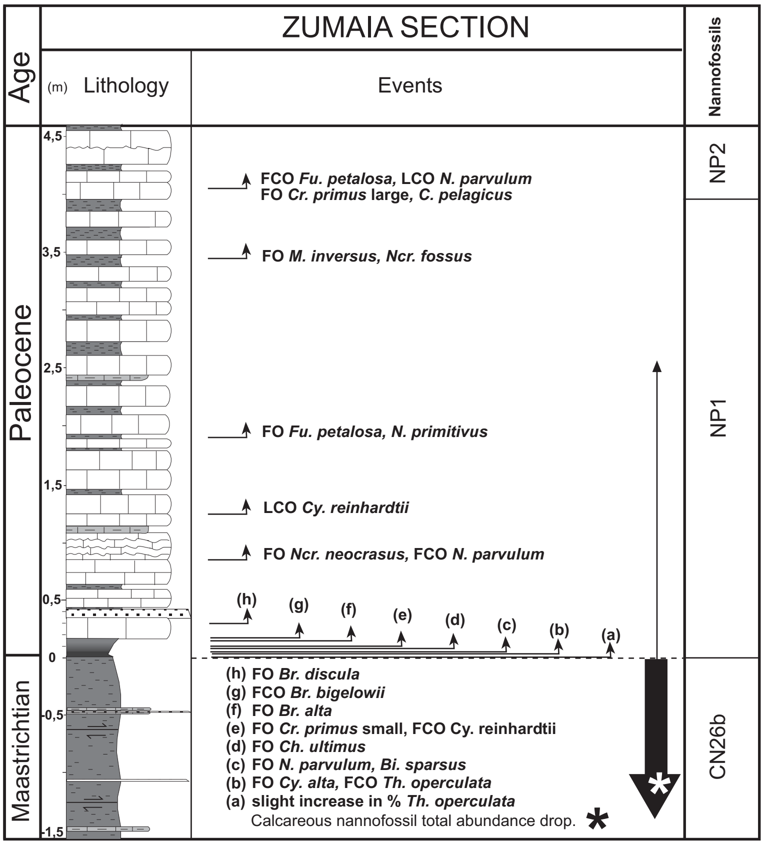


Fig. 12.- Selected calcareous nannofossil main biovents across the K/P boundary interval at Zumaia section.

At the K/P transition we recognize a total of 79 planktic foraminifer species. Of these, 63 are Cretaceous and the remaining 16 are present only in the Paleocene samples (Fig. 8). Despite of such a high diversity, we have not found some species characteristic of lower latitudes such as *Plummerita hantkeninoides* and *Gll. prairiehillensis* present for instance at the Beloc (Leroux *et al.*, 1995) and El Kef sections (Orue-Etxebarria, 1997).

All the species mentioned in the final part of the Maastrichtian reach theK/P boundary (Fig. 8) and no gradual extinction as that suggested by Keller (1988) has been observed. The source of the problem relies on the fact that if a sufficient amount of sample is not analyzed it is likely that some of the rare species may not appear and can be considered as extinct. At least 40 of the 63 identified species are rare and contribute each to less than 1% of the total number of specimens in the sample. Although no species become extinct in the last three metres below the boundary, there is a decrease in population size for some species probably due to environmental changes previous to the extinction at the K/P boundary. This is the case of *Gta. stuarti*, *Gta. stuartiformis*, *C. contusa*, *C. patelliformis*, *C. walfischensis*, *Gll. subcarinatus*, *A. mayaroensis*, *Sh. multispinata* and most rugoglobigerinids, specially *R. rugosa* and *R. scotti* which change from a relative proportion of more than 1% to less than 1%.

Of all the Cretaceous species recognized, 33 (52%) become extinct at the K/P boundary. The species most affected by the extinction event are the large-sized, complexly ornamented ones belonging to genera *Pseudotextularia*, *Racemiguembelina*, *Abathomphalus*, *Globotruncana*, *Globotruncanita*, and *Contusotruncana* which are characteristic of tropical and subtropical environments. All the species that disappear except *Pst. elegans* and *Pst. nuttalli* show a relative proportion smaller than 1%. Furthermore, the total of specimens included in the species that become extinct represent approximately 20% of the total.

With regard to Paleocene samples, we think that most of the Cretaceous species found in these samples are surviving ones. We do not agree with Smit (1982, 1990) and Olsson *et al.* (1999) who suggested that only one or very few species survived the K/P boundary extinction. Our opinion is based mainly in the following observations: (1) it should be pointed out that not only in the sections of this study but also in Caravaca, Brazos River, El Kef, etc., upper Maastrichtian foraminifers consistently occur at the base of the Paleocene; it would be intriguing to think that reworking processes would work in the same way for the same intervals in all these sections; (2) it is still more significant to realize that most of the Cretaceous species found in the base of the Paleocene are always the same in all the sections and that, furthermore, their stratigraphic distribution or temporal extinction at the beginning of the Paleocene is always very similar in all of them; (3) on the other hand, if they are reworked specimens it is difficult to explain why most of the species are the small-sized and thin-walled ones and not those with a solid test and thick-walled; and (4) finally, it is important to indicate that most of the planktic foraminifers considered to be surviving species are smaller in the basal Paleocene samples than they were in the latest Cretaceous ones.

The 30 surviving species progressively disappear along the *Civisina euskalherriensis* and *P. pseudobulloides biozones*. In the lower Paleocene samples we have recognized 16 Paleocene species, three that first appear in the *Gb. cretacea* biozone, nine that first occur through the *Civisina euskalherriensis* biozone and the other four that first appear in the *P. pseudobulloides* biozone.

These results partially agree with Smit's idea in the sense that all the species of the end of the Maastrichtian reach the K/P boundary, where they suffer a sudden extinction. On the other hand, there are a high number of species surviving the extinction which mainly affected the less cosmopolitan species.

The extinction of more than half of the species of the end of the Cretaceous was probably caused by the impact of, at least, one extraterrestrial bolide. This statement is reinforced by the relationship between the moment of the extinction and the appearance of the impact markers. Larger-sized species (i. e., *G. arca* or *Pst. deformis*), or even some small ones, initially survived to the environmental stress conditions at the beginning of the Paleocene by becoming smaller, but they later suffered a terminal extinction since they were not able to further adapt themselves to the new stressful conditions created by the impact. On the other hand, some small-sized surviving species, which were able to endure stressful environments, suffered a phyletic extinction evolving towards forms that would occupy the ecological niches left empty after the boundary extinction. Based on data we now have (Apellaniz *et al.*, 2002) both from the Basque Basin and elsewhere, we can affirm that all or almost all the species characteristic of the beginning of the Paleogene did evolve from surviving Cretaceous forms, but from a number of species notably higher than that suggested up to now (Fig. 13-15).

2.5. Evolution of the early Paleogene planktic foraminifera *(E. Apellaniz, X. Orue-Etxebarria, H. Luterbacher)*

Evolutionary trends in planktic foraminifera across the K/P boundary are very well documented in numerous surface sections of the Basque Country (Western Pyrenees). Apellaniz *et al.* 2002 proposed a phylogeny of the early Paleogene planktic foraminifera which was somewhat more complex than those suggested by previous authors. They favour a polyphyletic radiation which starts after the faunal crash at the Cretaceous/Paleogene boundary from the surviving populations of *Hedbergella* spp., *Heterohelix* (s.l.) spp. and *Guembelitria* spp. which show considerable morphologic variability (Figs. 13-15). These morphotypes which lived in the K/P 'Strangelove'ocean (Zachos *et al.*, 1989) underwent important and maybe in part heterochronous morphologic changes in their attempt to occupy ecological niches emptied by the mass extinction at the end of the Cretaceous.

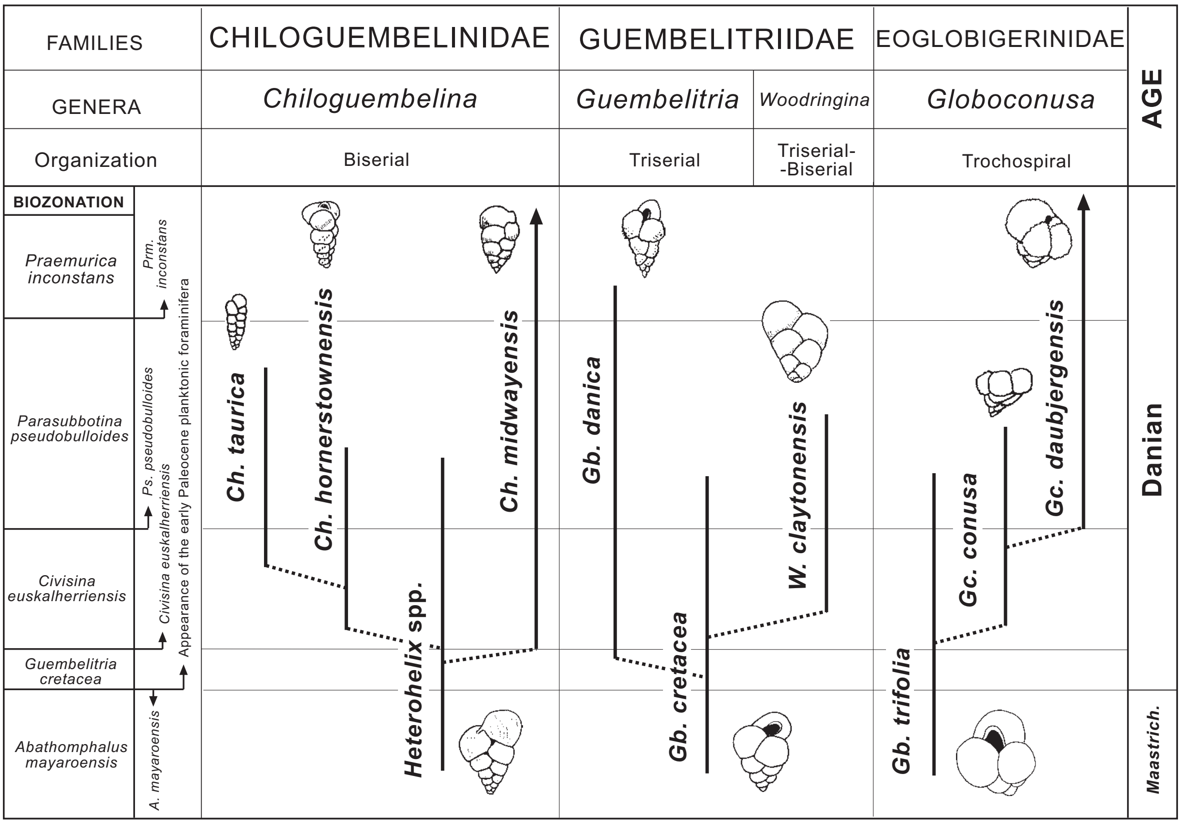


Fig. 13.- Evolution of the early Paleogene planktic foraminifera with perforate wall (except species of the genus *Parvularugoglobigerina*) originating fom *Heterohelix* spp., *Guembelitria cretacea* and *Guembelitria trifolia*.

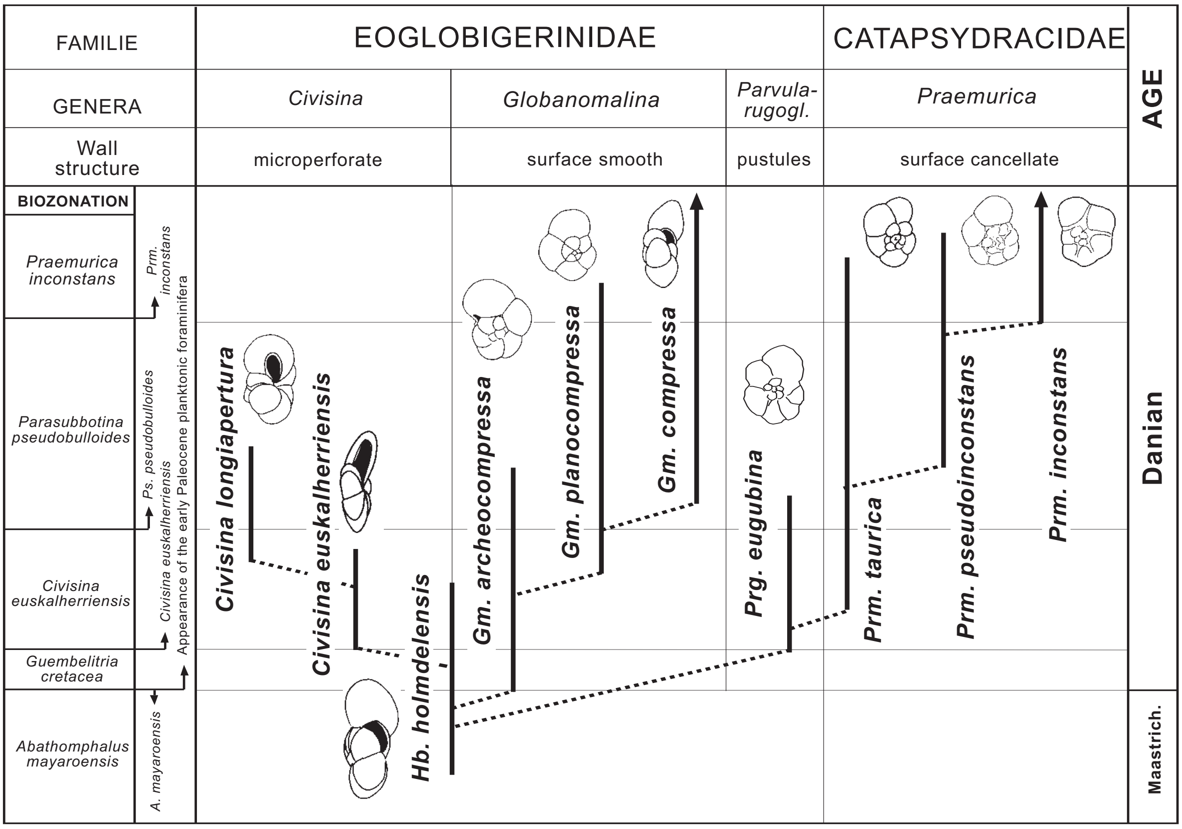


Fig. 14.- Evolution of the early Paleogene planktic foraminifera (genera *Parvularugoglobigerina*, *Globanomalina* and *Praemurica*) originated from *Hedbergella holmdelensis*.

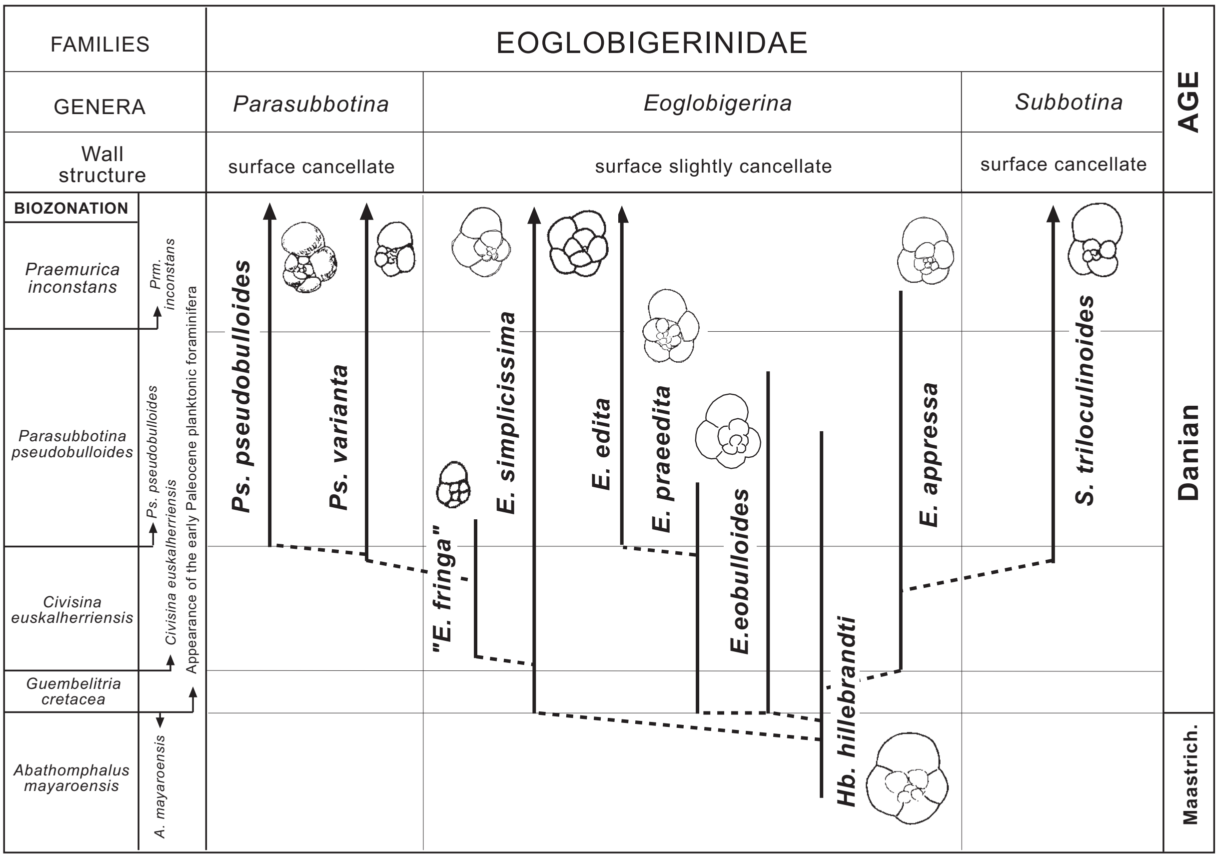


Fig. 15.- Evolution of the early Paleogene planktic foraminifera with cancellate or weakly cancellate wall (genera *Parasubbotina*, *Eoglobigerina* and *Subbotina* originating from *Hedbergella hillebrandti*.

Quite a few morphotypes developed during this rapid adaptive radiation were not successful as demonstrated by the many very short-lived species.

The proposed polyphyletic radiation is based on the recognition of a considerable number of Cretaceous species which survive into the early Paleocene as shown by the study of sections in different regions and depositional environments (Orue-Etxebarria, 1997; Apellaniz *et al.*, 1997; Apellaniz, 1999).

A total of nine different evolutionary lineages are recognized within the early Paleocene planktic foraminifera (Figs. 13-15). The new genus *Civisina* with its type-species *C. euskalherriensis* is characterised by the appearance of a very characteristic apperture which extends onto the peripheral margin of the youngest chamber (Fig. 14). Besides its type-species, it includes also *C. longiapertura* which is kept separate from *Parvularugoglobigerina eugubina*.

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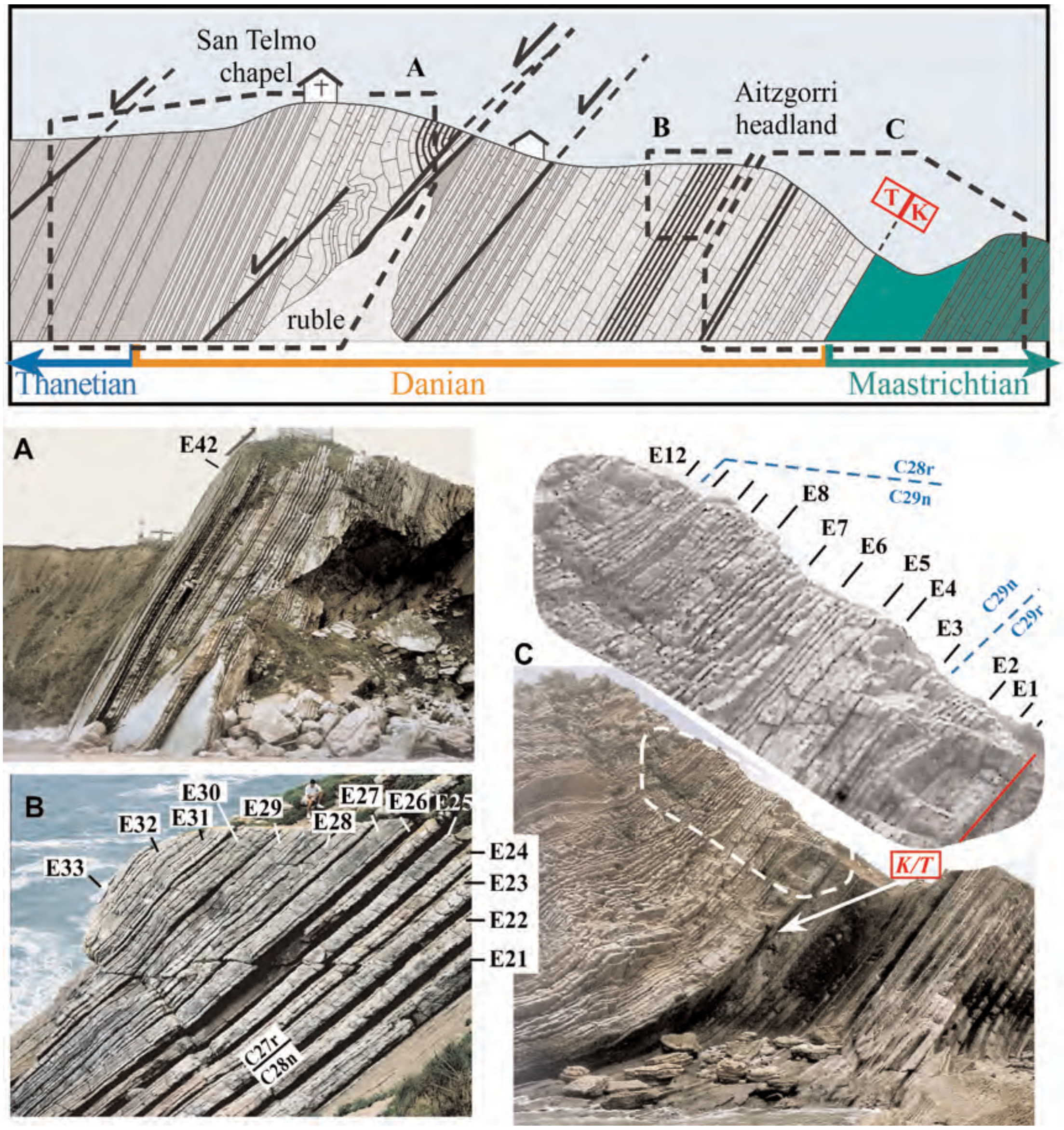
Fig. 16.- Schematic view of the Aitzgorri headland outcrop and location of field photographs (A, B, C) showing different aspects of the studied Danian section. The position of the K/P boundary, the magnetic chron boundaries and some of the ~110 ky eccentricity cyles are shown. (see also Fig. 17).

3. CYCLO-, MAGNETO- AND CALCAREOUS PLANKTON BIOSTRATIGRAPHY OF THE DANIAN LIMESTONE

3.1. Magnetostratigraphy of the Daniam Limestone (*J. Dinarès-Turell, J.I. Baceta, V. Pujalte*)

The intensity of the Natural remanent magnetization (NRM) of the 1992 samples collected for the magnetostratigraphic analysis of the Danian of the Zumaia section (Figs. 16-17) was generally around 1×10^{-3} A/m although it was lower for some of the whitish-coloured carbonate-rich beds. Three magnetization components can be recognized in most samples upon demagnetization, in addition to a viscous magnetization removed below 100-150°C. Component H conforms the ChRM and has either normal or reverse polarity in bedding-corrected coordinates. Before bedding-correction, component H has a steep inclination not compatible with a Paleocene to recent geomagnetic field direction, whereas the mean ChRM inclination after bedding correction is near the expected Paleocene inclination for the site paleolatitude.

The ChRM declination, inclination and VGP latitude together with relevant biostratigraphic and lithologic information for the studied section is shown in Fig. 17. Taken the K/P boundary, known to occur in the upper part of chron C29r, as the starting point, the succession of magnetozones at



Zumaia can be straightforwardly correlated to the standard GPTS. Six reversal boundaries are unambiguously identified at a bed level. Within chron C26r, a few samples with positive VGPs that delineate two short normal zones that may represent cryptochrons are identified (Fig. 17).

3.2. Cyclostratigraphy of the Danian Limestone (J. Dinarès-Turell, J.I. Baceta, V. Pujalte)

In order to evaluate sedimentary-forcing mechanisms from the lower Paleocene strata, a lithologic coded series has been generated at intervals of 1 cm and studied with spectral analysis. A bed-by-bed lithologic log was created in the field and the GSA Rock Colour Chart (based on the Munsell® colour system of Hue/Value/Chroma codes) was used to estimate the colour of all limestone beds. Marlstone beds are mostly reddish in colour. Limestone beds can be grouped in three colour categories (reddish, pinkish and whitish), which have a clear relation to carbonate content (red colour decreases with relative high carbonate content). In addition to the bedding stacking information in terms of limestone/marl lithology, the arbitrary coding for the lithologic series takes into account the colour of the beds as follows. Limestone beds have been given a value of 90, 60 or 40 depending whether they were whitish, pinkish or reddish in colour respectively. Marlstone beds have been given a value of 10 or 20 depending on the field appreciation of their relative carbonate content. A value of 20 has been also assigned to corresponding marly bed partitions within crowded bundles. Consequently, the evenly spaced (1cm) constructed lithologic series ranges from 10 to 90 and represents an approximation to a carbonate proxy record.

Standard (Fourier) spectra were calculated for the lithologic-coded series. For that use and for filtering purposes, the software package "AnalySeries" has been employed. The Continuous Wavelet Transform (hereafter CWT) has also been applied to analyze the time evolution of the spectral power and to test its statistical significance. For nonstationary time series, the CWT can be particularly useful to unravel and document periodicities that might be present at different times (i.e. localization of periodicities), related, for instance, to variations of the sedimentation rate or to detect (temporal) discontinuities of the accumulation. The CWT has been chosen as it is proven superior to the windowed Fourier transform, given its inherent multi-scale character.

The stacking of hemipelagic marlstone and limestone lithologies at Zumaia defines a cyclic (or rhythmic) bedding pattern (Figs. 16, 18). The basic noticeable cycles are the *bedding couplets*, which consist of alternating pairs of marl and limestone beds in different relative proportions or by a vertical variation in carbonate content within marl or limestone intervals. Thickness of these couplets ranges from 15 to 45 cm (average of about 25 cm). The limestone/marl couplets can in turn be grouped into *bundles* usually containing 4 to 6 couplets. A whole range of bundles has been observed. These include *open* bundles where usually all the constituent couplets retain their marly and lime portions (i.e. bundles E6-E8 in Fig. 16C), and *crowded* bundles which, in contrast, almost lack recognizable the marly partitions between limestones (i.e. bundles E23-E29 in Fig. 16B). The thickness of the bundles diminishes from open (maximum thickness of ~2 m) to crowded ones (minimum thickness of ~60 cm) with an average of about 1.2 m. The precession (~21 kyr average periodicity) and eccentricity (~100 kyr) origin for the couplets and bundles is roughly confirmed when the duration of magnetozones and biostratigraphic intervals retrieved in the Early Paleocene segment are compared to standard time-scales.

Spectral analysis of the coded series was first performed in the depth domain (Fig. 19). The results using the Blackman-Tukey method, which is based on the standard Fourier transform, reveal spectral peaks in the frequency band at 21-26-31 cm and 83-122 cm that correspond to the average thickness of the basic lithologic couplet and bundle cycles respectively (Fig. 19A). The filtered record of the 1.22 m component follows well the stacking pattern of bundles. The difference in bundle thickness is probably indicative of varying sedimentation rates along the section. The global wavelet spec-

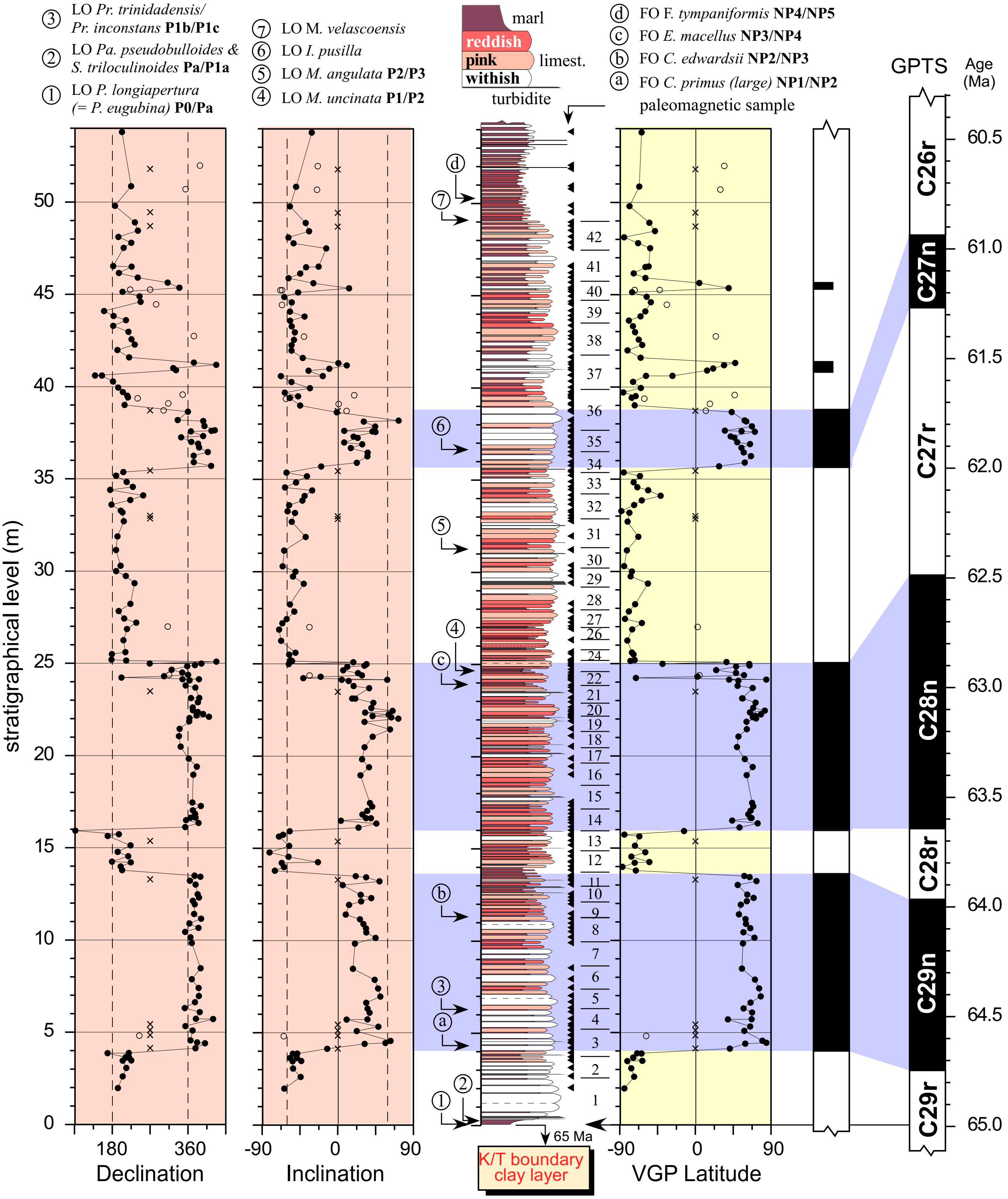


Fig. 17. Stratigraphic variations of declination, inclination of the ChRM vectors and virtual geomagnetic pole (VGP) latitude plotted on a lithologic log where main biostratigraphic events and position of paleomagnetic samples are indicated. Open circles denote unreliable data and crosses mark the position of samples that have provided no data. Correlation to the GPTS is provided.

trum (Fig. 19B) shows significant peaks at periods of 30 cm and 1.22 m that are comparable to the ones detected with the Blackman-Tukey method that correspond to the ~21/kyr precession and ~100 kyr

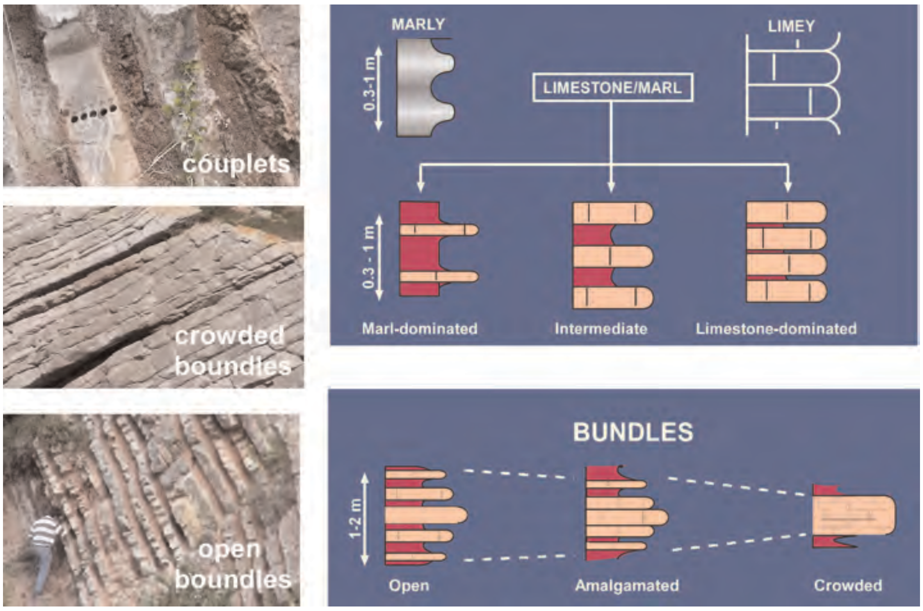


Fig. 18.- Key features of the lithologic stacking pattern for the Paleocene Zumaia section.

eccentricity cycles respectively. In addition, a longer period peak at 4.4 m is also obvious. This period most likely corresponds to the 404-kyr eccentricity cycle. Recent long-term numerical simulations of the orbits of the major planets in our solar system (Varadi *et al.*, 2003) has enabled an accurate Earth's orbital eccentricity time series (Va03_R7 solution) to be computed back to 110 Ma with restricted accuracy beyond the past 50 Ma. More recently Laskar *et al.* (2004) have also provided a full integrated target orbital solution.

For orbitally tuning "floating" geologic records, one should ideally use the single and more stable long eccentricity cycle of 404-kyr period. The corresponding peak of this cycle in the depth domain does not seem to be clearly present and persistent in the Zumaia lithologic coded series (see above). However, we note that the strongest localization of this peak in the 33-43 m interval (Fig. 19) includes two obvious thick whitish (carbonate-rich) bundles (E35 and E36) that conform a maxima in the corresponding 4.4 m filter output as previously noted. Another interval of high amplitude in the 4.4 m filter output (and significance of this period in the local wavelet) occurs towards the base of the section where several carbonate-rich bundles exist. The time span between these two intervals can roughly be estimated to be about 2.8-3.2 Ma (by counting the ~100-kyr eccentricity bundles or the number of maxima in the 4.4 m filter output and assumed to correspond to the 404-kyr eccentricity cycle). This is consistent with the long-term eccentricity cycle of ~2.4 Myr period which defines modulation nodal patterns of ~2.0 Myr and ~2.8 Myr. When examining the Va03_R7 eccentricity solution for the time interval 61-66 Ma, it is found that two such nodes of low eccentricity amplitude exits at ~62.1 Ma and ~64.9 Ma resulting in a 2.8 Myr modulation cycle (Fig. 20) (note that low-amplitude ~100-kyr eccentricity results in a high- amplitude 404-kyr eccentricity). A direct tuning of the Zumaia section to the Va03_R7 eccentricity solution can therefore be attempted by matching the intervals of high significance of the 4.4 m period to the nodes of the ~2.8 modulation cycle as the starting point. The node towards the base of the section is less obvious and could also be marred by long-term effects of the K/P boundary (i.e., nannofossil abundance does not recover until 12 m above the K/P boundary). The obvious carbonate-rich bundles E35 and E36 from the upper interval appear more suitable for correlation. The phase relationship between the sedimentary record and the astronomical target needs first to be established. We assume that relatively carbonate-rich intervals correspond to eccentricity minima as usually found in

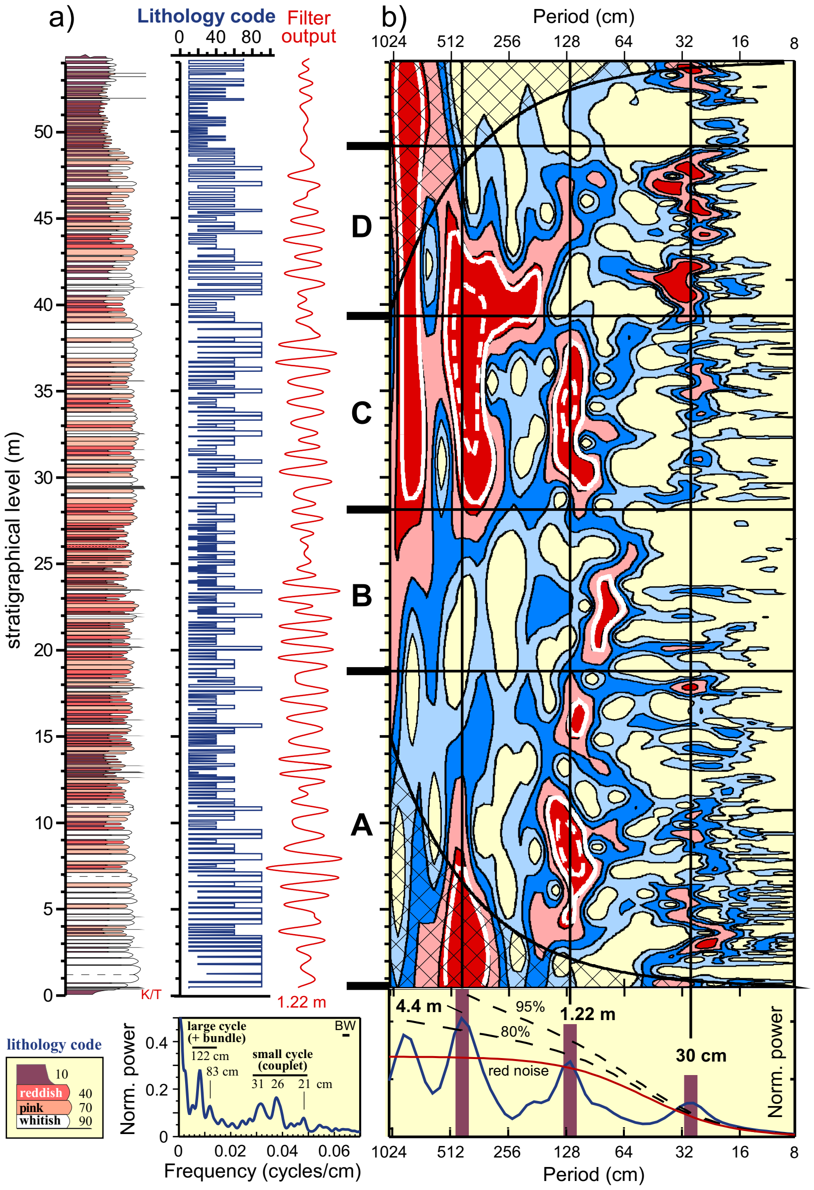


Fig. 19.- (a) Lithologic column, coded lithologic series and filter outputs (Gaussian filter centred on 1.22 m (0.8197 ±0.4m^{-1} bandpass) and 4.4 m (0.2272 ±0.05 m^{-1} bandpass) on a stratigraphis scale. At the bottom is shown the power spectra (Blackman-Tukey method with a Barlett window) of the coded lithologic series where the bandwidth (BW) and the period of relevant peaks are given. (b) local wavelet power spectrum of the coded lithologic series using a Morlet wavelet. The colored contours are at normalized variances of 2, 5, 10 and 20. The thick white contours enclose regions of greater that 95% (dashed) and 80% (continuous) confidences for a red-noise process with a lag-1 autoregressive process plotted only for periods greater than 64 cm. Cross-hatched regions on either end indicate the "cone of influence" where edge effects become important (see Torrence and Compo, 1998). The global wavelet spectrum for the entire series is shown at the bottom.

marine pelagic records. Consequently, the center of bundles E35 and E36 are correlated to consecutive eccentricity minima that define a distinct minimum in the 404-kyr eccentricity cycle around 62.35 Ma (Fig. 20) as a start in our tuning. The tuning procedure continues by matching successive bundles to

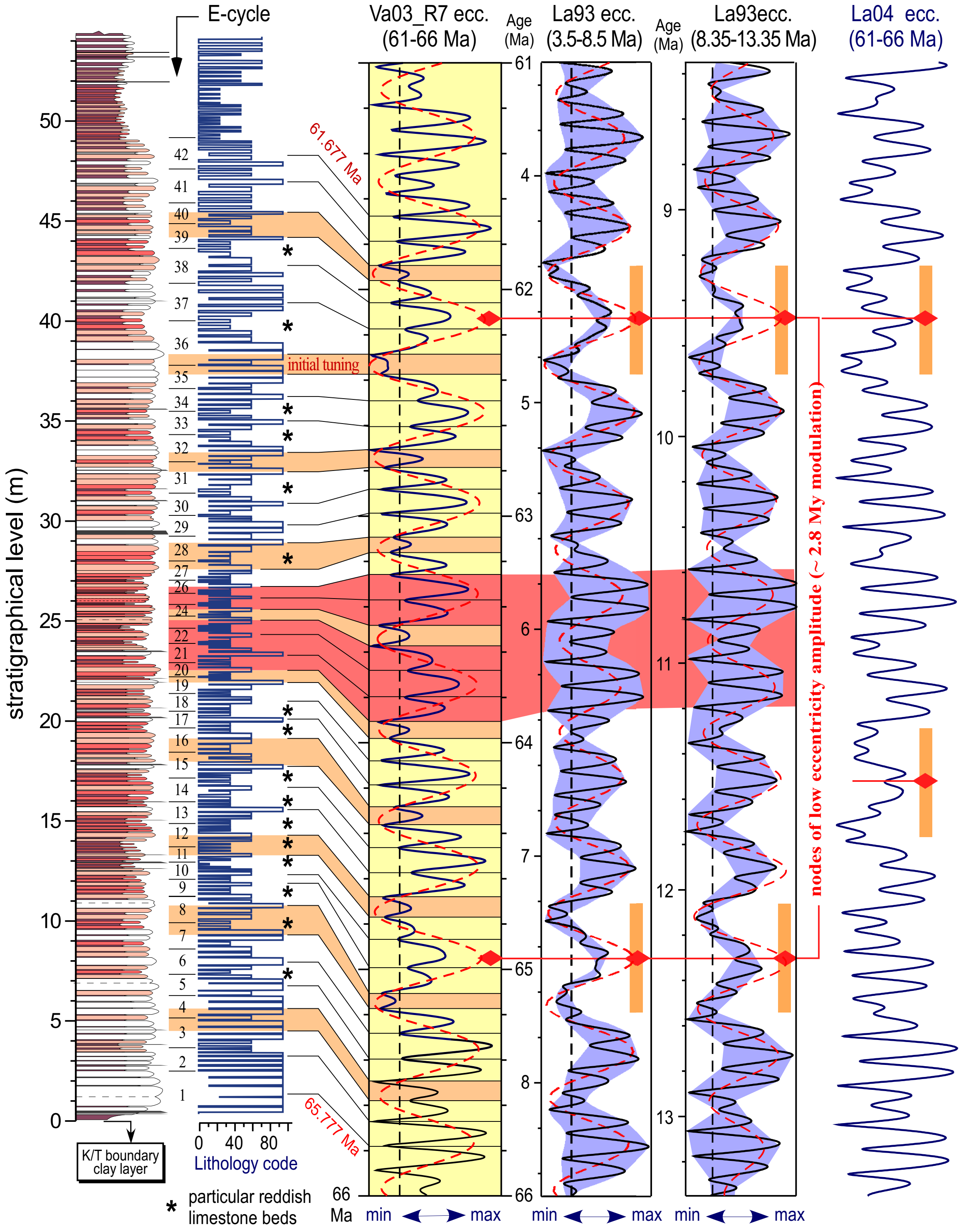


Fig. 20.- Tuning of the lithologic bundles (E-cycles) to eccentricity minima in the eccentricity time series of solution Va03_R7 [Varadi *et al.*, 2003]. Corresponding 404-kyr eccentricity minima are shaded. Two equivalent time slices of the eccentricity time series of solution La93 (Laskar *et al.*, 1993) with an envelope around eccentricity minima and maxima are plotted for comparison. The two La93 solution time slices have been chosen to anchor on two nodes of low eccentricity amplitude in the Va03_R7 solution. Initial tuning matches the two carbonate-rich bundles 35 and 36 to consecutive eccentricity minima that conform a minimum in the 404-kyr eccentricity cycle related, in turn, to a node of low eccentricity amplitude of the ~2.4 Myr eccentricity modulation. The new Laskar La04 orbital solution is also shown.

Table 1
Cycle-tuned estimates for the duration of Early Paleocene chrons. Cycles refer to precession otherwise is specified.

Chron	CK95	ODP Sites 1001A, 1050C Rohl et al (2001)		Bjala (Bulgaria) Preisenger et al., (2002)		Zumaia Dinarès-Turell et al. 2003	
	Duration (kyr)	Cycles	Duration[a] (kyr)	Cycles	Durat.[a] (kyr)	Cycles	Duration[a] (kyr)
C29r_Pal.	245	~14	300	11	231	12	252 (228)
C29n	769	~32	670	31	651	38-39	796-819 (879)
C28r	342		-	9	189	12-13	252-273 (210)
C28n	1135		-	54	1134	49	1029 (1089)
C28	1477		1280[b]	63	1323	61-62	1287-1302 (1228)
C27r	1223	23-25 obliq.	929-1010	44	924	50	1050 (964)
C27n	356	11-13 obliq.	444-525	10	210	11	231 (273)
C27	1579	36 obliq.[c]	~1450	54	1134	61	1280 (1248)

a) Applying a modern mean precession and obliquity periods of 21 kyr and 40.4 kyr respectively). Duration in brackets is from the eccentricity-based tuning to the Va03_R7 astronomical target.

b) Estimated assuming an intermediate spreading rate (16.7 km Ma-1) for C28 (interpolated from the South Atlantic magnetic anomaly profile.

c) Cycle counting in Hole 1001A indicates 47 precession cycles + 11 obliquity (obliq.) cycles for C27.

successive eccentricity minima in the Va03_R7 solution (Fig. 20). For comparison purposes two time-slices of the La93 solution containing an equivalent ~2.8 Myr modulation cycle are plotted in Fig. 20 in addition to the most recent La04 solution.

In order to be more useful we compute elapsed time (as opposite to absolute ages) and estimate duration of chrons and relative ages for bioevents by both cycle counting precession cycles and by the eccentricity-based tuning to the Va03_R7 orbital solution. A cycle-tuned duration for the different Early Paleocene chrons from C29r to C27n at Zumaia is presented in Table 1

3.3. Planktic foraminifera biostratigraphy of the Danian Limestone (X. Orue-Etxebarria, E. Apellaniz)

All studied samples contained planktic foraminifera although the content was variable with the poorest at the K/P boundary clay (200-300 specimens/g and smaller than 100 um). Higher in the section, the richness, diversity and size of planktic foraminifera increase gradually, allowing the separation of more than 500 specimens per sample.

The distribution of species at Zumaia (Fig. 21) largely coincides with that pro-posed in the standard biozonation although some significant differences have been detected. Firstly, the highest occurrence (HO) of *Parvularugoglobigerina eugubina* was recognised 5 cm above the lowest occurrence (LO) of *Parasubbotina pseudobulloides*, which in turn coincides with the LO of *Subbotina triloculinoides* and other significant species such as *Eoglobigerina edita* or *Globoconusa daubjergensis*. Secondly, the LO of *Globanomalina compressa* is clearly older (4.52 m stratigraphically below) than the LO of *Praemurica inconstans-trinidadensis*. The same biostratigraphic circumstances have also been observed in higher-resolution studies of other Lower Paleocene sections from the western Pyrenees such as Sopelana and Bidart (Apellaniz *et al.,* 1997; Apellaniz, 1998), which raises questions on the validity of the criteria used by Berggren *et al.* (1995) to establish the Pa/Pl and Plb/Plc biozonal boundaries. Consequently, we kept the criteria used in an earlier work by Berggren and Miller (1988) that use the first appearance datum (FAD) of *Ps. pseudobulloides* and the FAD of *Pr. inconstans-trinidadensis* to define the Pa/Pl and Plb/Plc boundaries respectively. Likewise, the coincidence of the LOs of *Ps. pseudobulloides* and *S. triloculinoides*, which are older than the HO of P. eugubina, makes it impossible to differentiate the P1a and Plb subzones as proposed by Berggren *et al.* (1995) and Berggren and Miller (1988).

The P1/P2 and P2/P3 biozones can be identified unambiguously at Zumaia because the species that define them (*Praemurica uncinata* and *Morozovella angulata* respectively) are abundant since their first occurrence. Finally, it is worth noting that due to the loose sample spacing in the upper part of the studied section, the LO of *Igorina albeari*, a datum that defines the P3a/P3b subzone according to Berggren *et al.* (1995), cannot be precisely determined.

*3.4. **Calcareous nannofossil biostratigraphy of the Danian Limenstone*** *(G. Bernaola)*

Preservation and abundance of calcareous nannofossils vary along the section in a similar way to the planktic foraminifera. Calcareous nanno-fossil abundance is low (0.77 specimens/field of view, spp./f.v.) in the marly interval above the K/P boundary and progressively increases along the first 12 m of the Danian limestones where it reaches similar values to those observed in the uppermost Cretaceous (~5 spp./f.v.). The nannofossil abundance keeps constant, or increases slightly, up to the marl-limestone to reddish marl transition at 49 m, where it increases sharply with values of 22 spp./f.v. at the base of the reddish marls (Fig. 21).

In accordance with the vertical distribution of the calcareous nannofossil species, we recognised at Zumaia the first five biozones of the biostrati-graphic scale of Martini (1971) (Fig. 2l). It is important to point out that to establish the NP1/ NP2 and NP2/NP3 biozone boundaries we considered the modifications introduced by Perch-Nielsen (1981, 1985) based on the taxonomic clarifications of Romein (1979). The NP1/NP2 is consequently defined by the LO of *Cruciplacolithus primus* (large) (and not the LO of *Cr. tenuis*) whereas the NP2/NP3 boundary is defined by the LO of *Cr. edwarsii* (and not the LO of *Ch. danicus*). The precise location of all the zonal boundaries is shown in Fig. 21. The NP3/NP4 and NP4/NP5 boundaries are defined by the LO of *Ellipsolithus macellus* and Fasciculithus tympaniformis respectively (Martinii, 1971). Although *E. macellus* appears discontinuously during the first 16 m after its first appearance (Fig. 2l) we think that its LO at 23.95 m is reliable since it is very close to the LO of *Neochiastozygus saepes* (occurring at 25.75 m at Zumaia), used at high latitudes to approximate the NP3/NP4 boundary (Perch Nielsen, 1979). The first occurrence of *F. tympaniformis* at 50.5 m is similar to that reported by von Salis in Schmitz *et al.* (1997). We also note that the LO of *Neochiastozygus perfectus*, which in North Sea sections first appears just below the unconformity separating Danian and Selandian deposits occurs 150 cm below the first record of *F. tympaniformis* (Thomsen, 1994).

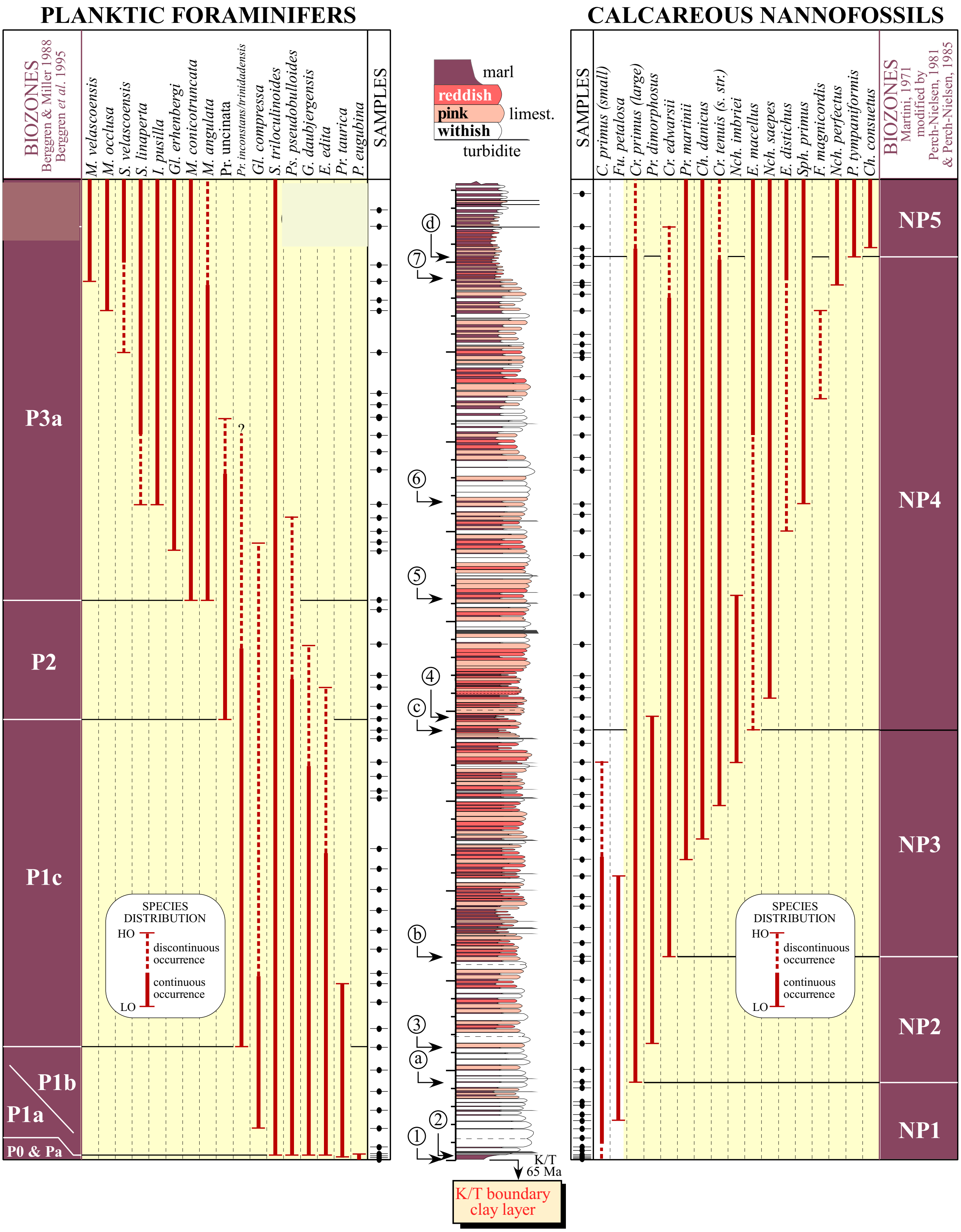


Fig. 21.- Selected calcareos nannofossil and planktonic foraminifera species distribution across the Danian Limestone of the Zumaia section. 1-7 and a-d Planktonic foraminifera and calcareous nannofossil biovents as shown in fig. 17.

4.- THE MID-PALEOCENE INTERVAL AT ZUMAIA

4.1. An overview *(J.I. Baceta, G. Bernaola)*

The mid-Paleocene represents one of the most interesting intervals of the Zumaia section, as it includes abrupt lithologic changes, a short-term episode of remarkable paleoceanographic change, and one of the most complete records of the main biostratigraphic events and magnetic reversals that took place during the late Danian-middle Thanetian time span. Some of these attributes were already stressed in several previous works on the section (e.g. Pujalte *et al.*, 1993, 1995 and 1998; Arenillas, 1996, Baceta, 1996, Schmitz *et al.*, 1998; Dinares-Turell *et al.*, 2003, Bernaola, 2002, Baceta *et al.*, 2004), although they have been overshadowed by the more striking events associated to the Cretaceous/Paleogene and Paleocene/Eocene boundaries. A raising interest on some of the biotic, oceanographic and even astronomic signals across the interval and the need of a formal decision about the internal subdivision of the Paleocene epoch, have prompted a detailed re-inspection of the mid-Paleocene of the Zumaia section. In fact, it has been proposed as a firm candidate for the definition of the GSSP's of the Danian-Selandian and Selandian-Thanetian boundaries (see www.stratigraphy.org/gssp.htm) and, under this perspective, the field trip will deal on the main results obtained in ongoing studies and the initiation of the discussion about the most suitable criteria.

For a better understanding of the time interval involved, the mid-Paleocene is considered in this and following sections as the stratigraphic interval comprising from the reversal defining the top of the magnetochron 27n (upper Danian) to the NP6-NP7 zone boundary (Fig. 22), just a few metre upwards from the reversal at the base of magnetochron 26n (lower? Thanetian). This interval is nicely exposed in the cliffs along the southern half of the Itzurun beach (Fig. 23), which will be firstly observed from the distance from the cliff top and later analysed in detail at the cliff base. The San Telmo chapel represents the best reference point for the two stops; it is a small construction overlooking the beach, than is easy to recognise by being resting just on the cliff edge.

4.2. Lithostratigraphy *(J.I. Baceta, G. Bernaola, J. Arostegi)*

The mid-Paleocene succession at the Itzurun beach is 48m thick and occurs dipping 40-60º to the NNE. The outcrops have a similar orientation and thus offer a continuous dip section, only interrupted by the presence of three fault systems (F1 to F3 on Fig. 23). These are normal faults with associated normal and reverse faults and joints, that in the case of F2 and F3 have duplicate significant intervals of the succession. A careful correlation of distinctive beds in the hangingwalls and footwalls of the fault systems has allowed a detailed reconstruction of the stratigraphy of the mid-Paleocene succession and its more representative intervals (Fig. 22).

Stratigraphically, the mid-Paleocene comprises the upper part of the Danian Limestone (Apellaniz *et al.*, 1983) and the lower-middle parts of the Itzurun Fm (Baceta *et al.*, 2004). These two intervals are respectively 16 m and 32 m thick. The two units are mainly made up of regular alternations of (hemi)pelagic indurated limestones, marlstones and marls. In the Danian Limestone the indurated limestones are dominant, whereas in the Itzurun Fm the proportions of the three lithologies vary considerably. Additionally, they also include minor but significant amounts of thin-bedded turbidites, of siliciclastic, calcareous and mixed nature. The transition between the two formations is marked by an abrupt lithologic change, which is interpreted as the expression of the prominent episode of sea level fall that characterised the end of the Danian across the whole Pyrenean Domain (Baceta *et al.*, 2004, Baceta *et al.* 2005).

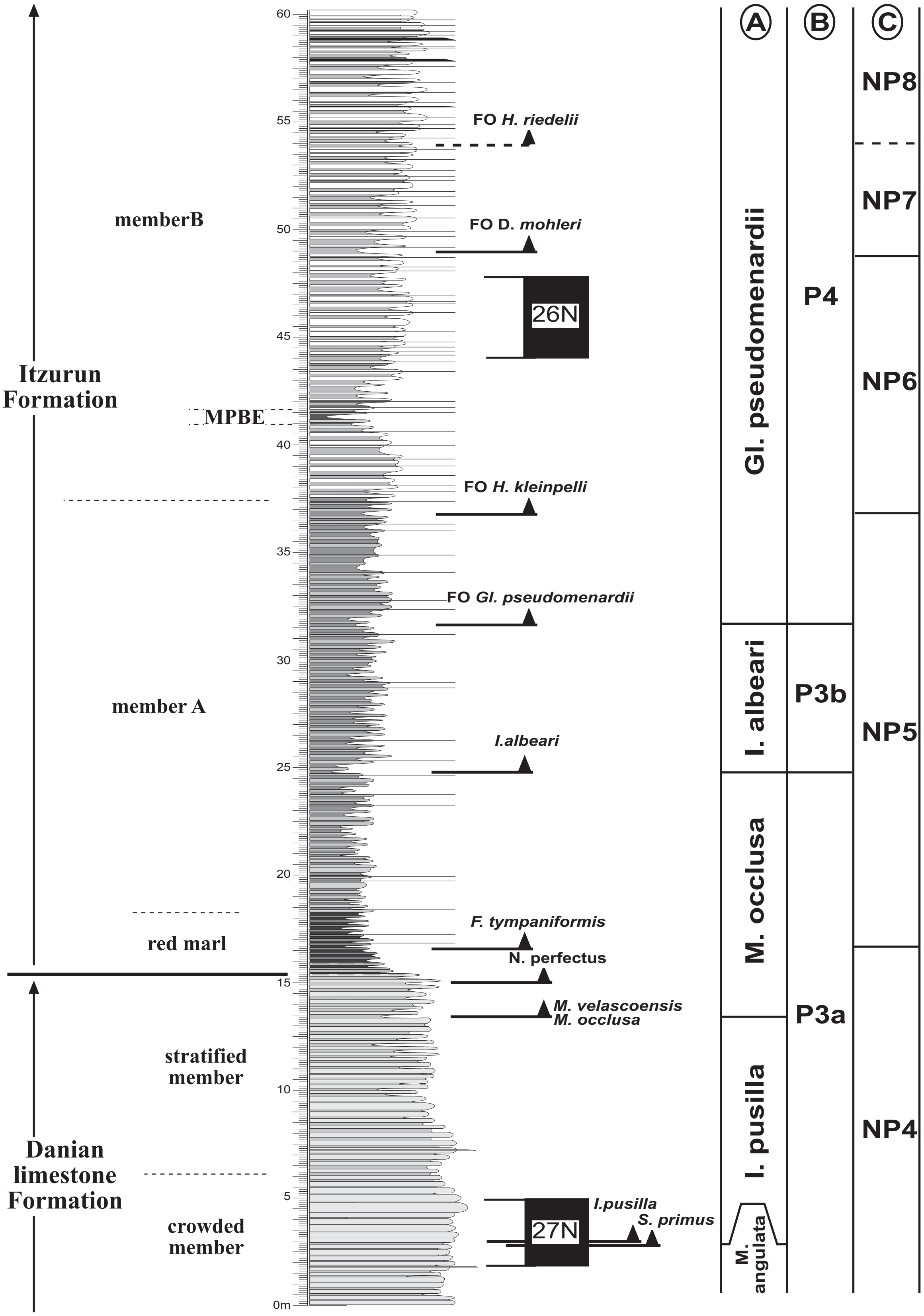


Fig. 22.- Integrated lithostratigraphy and biostratigraphy of the Mid-Paleocene of the Zumaia section. Biozonations: (a) Orue-Etxebarria *et al.* (this volume), (b) Berggren *et al.* (1995), (c) Bernaola (2002) according to Martini (1971).

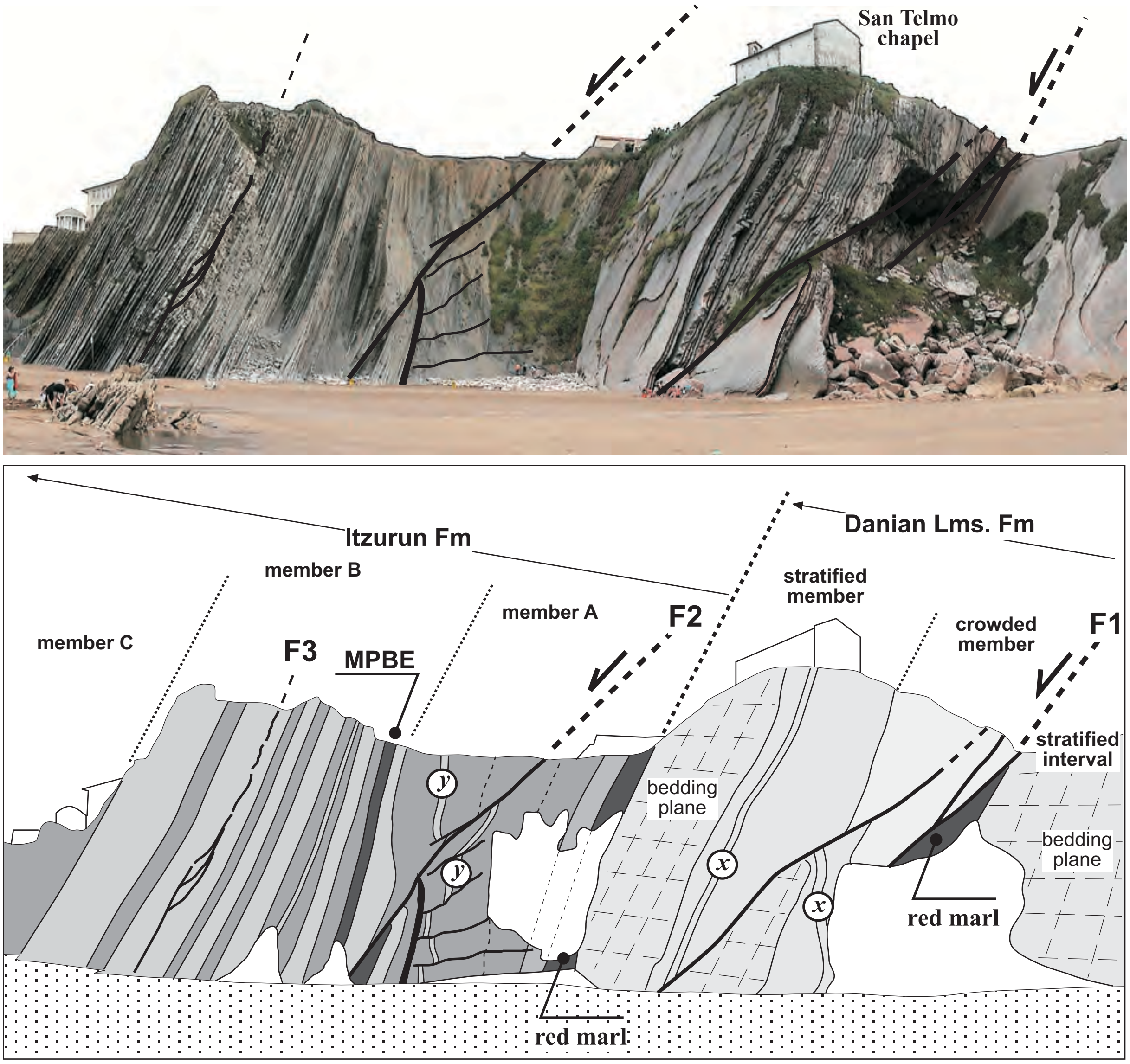


Fig. 23.- Panoramic view and synthetic sketch of the Mid-Paleocene outcrops in the Itzurun beach, showing the distribution of the main lithological units/members and the three fault systems that disrupt the succession.

For descriptive purposes, the Danian Limestone and Itzurun Fms can be subdivided in a series of informal members that can be easily identified in the outcrops. In the upper part of the Danian Limestone, largely of pink-reddish colours, two members named the "crowded" and the "stratified" intervals can be distinguished (Figs. 22 and 23). The crowded member is 7 m thick and consists of limestones amalgamated or with very thin marly interbeds. The stratified member, 9 m thick, takes its name from the good definition of the bedding and the clear contrast between limestones and intervening marlstones, which occur forming some of the best examples of ca.100 ky eccentricity cycles identified in the section (see Dinares-Turell *et al.*, 2003 and descriptions below). In the upper part, 4 m below the top of the unit, are located the "Jan Smit beds", a distinct couple of beds named in honour to Prof. J. Smit, who was the first noting than these and other characteristic beds of the upper Danian Limestone can be identified and correlated between Zumaia and the also well-known section of Hendaia I. Later works have demonstrated that the Danian Limestone can be correlated bed-by-bed across the whole basin floor domain, including sections such as Sopelana, Trabakua, Monte Urko, Zumaia, Artadi and Hendaia I and II (Baceta, 1996, Pujalte *et al.*, 1998).

The sedimentologic features of the Danian Limestone are rather constant across the section. The limestones range in composition between 70-95% in $CaCO_3$ and the marlstones between 40-65%, with clay minerals as subordinate component and traces of quartz, dolomite, plagioclase and Fe oxides and sulphides. The study of 18 samples from marly interbeds in the upper 12 m of the unit has revealed that the clay association includes illite (40-78%), smectite (7-15%) and chlorite/smectite mixed-layer minerals (33-50%) and that kaolinite only occurs as a trace component in 3 samples. These results differ significantly from those of Gawenda *et al.* (1999), which stressed that kaolinite occurs in proportion between 20-30%. In thin section, the limestone occur as burrowed micritic mudstone-wackstones with tests of planktic foraminifers and rare calcispheres. Trace fossils are very common in both marlstones and limestones, mainly corresponding to *Planolites*, *"Thalassinoides"* and occasional *Zoophycus*. The upper part of the unit also intercalates two levels 1-3 cm thick of calcareous turbidites, largely composed of tests of planktic foraminifers. Other turbidite levels within the Danian Limestone also include significant amounts of fragments of shallow benthic biota, such as red algae, benthic foraminifers and corals, indicating that although being located in a distal position the Zumaia section received resedimented material from the shallow platform areas.

As mentioned above, the boundary between the Danian Limestone and the Itzurun Fm is marked by a distinct lithological change, from limestone- to marl-dominated, that creates a distinct retreat of the cliff face. This change is associated to a progressive but rapid 30-40% decrease in $CaCO_3$ in both the hard (limely) and softer (marly) beds. However, it is important to note that this lithologic boundary does not imply any important change in the clay mineral assemblages. Based on calcareous nannofossil age estimations and detailed mapping, this important lithologic boundary can be correlated with the prominent discontinuity that segments the Paleocene succession across the slope and shallow platform domains, known as the mid-Paleocene unconformity (Baceta *et al.*, 2001 and 2005). This discontinuity developed as a consequence of a prominent sea level fall, both in time and magnitude, leading to the subaerial exposure of the Paleocene carbonate platforms during an interval estimated in about 2.5 My (Baceta *et al.* 2005). According to this correlation, the rapid decline in carbonate detected at the top of the Danian Limestone in Zumaia and other coeval section of the basin floor is interpreted as a switch-off in the support of fine-grained carbonate sediment (micrite) from the carbonate platforms to the open sea (Baceta *et al.*, 2004). The biostratigraphic analysis and detailed correlation between Zumaia and the sections of Artadi and Trabakua indicate that there is no significant stratigraphic gap associated to the Danian Limestone-Itzurun Fm boundary and, consequently, the marked lithologic change must be viewed as a correlative distal conformity of the mid-Paleocene unconformity.

The 32 m thick interval of the Itzurun Fm considered here shows higher vertical variations in both the relative proportion of (hemi)pelagic sediments and the amount of turbidite intercalations. It comprises the lower two informal members in which can be subdivided the formation (members A and B), of 24 m and 52 m in thickness, respectively (Fig. 22). The member A is largely dominated by marly lithologies, whereas the member B includes significant proportions of indurated limestones. The boundary between the two members has been established arbitrarily at the point in which limestones beds reach and maintain values in $CaCO_3$ higher than 60%. In contrast to the Danian Limestone, dark to light grey colours dominate across the Itzurun Fm. However, the base of the formation is defined by a 5.5 m thick interval of marls and marlstones of a characteristic red colour (the red marl, Figs. 22 and 23), which records a decrease in the thickness of the sedimentary couples of about 25-30%. This is interpreted in terms of a significant drop in the sedimentation rate, linked to the prominent mid-Paleocene drop of sea level identified across the Pyrenean Basin.

On the great scale, the members A and B of the Itzurun Fm record a progressive increase in $CaCO_3$ after the abrupt decrease that labels its lower boundary, culminating in the upper part of member B with values similar to those in the Danian Limestone. The composition of the limestones in the Itzurun Fm is rather similar to those in the underlying Danian Limestone (micritic mudstone-wackstones with planktic foraminifers), but occasionally they also contain minor amounts of silt-sized quartz and glauconite, this later as sub-mm grains and infilling the foraminiferal tests. 18 and 16 samples from the members A and B, respectively, have been analysed to determine the clay mineral associations in the Itzurun Fm. They have reavealed that the unit records a progressive change from illite-rich to illite/smectite-rich clays. Trace fossils are also common, but in this case the association is dominated by *Zoophycus*. Moreover, the Itzurun Fm differs from the Danian counterpart in the amount and type of turbidite intercalations. In the marly member A they mostly are of siliciclastic nature, whereas in the member B, which shows a higher number, they are usually of siliciclastic or mixed carbonatic-siliciclastic nature. The carbonate grains mainly correspond to abraded tests of planktic foraminifers. The presence of fragments of red algae and large benthic foraminifers (nummulitids and orthophragminids) in several turbidites of the middle-upper part of the member B evidences that by this time the platform areas were again submerged and were able to produce and export resedimented material to the basin.

A few meter above the limit between the members A and B of the Itzurun Fm, a prominent re-entrant can be recognised in the cliffted outcrops. Its corresponds to a 1 m thick interval recording a drastic decrease in $CaCO_3$, relatively high values in magnetic susceptivility that is interpreted as the expression of the mid-Paleocene biotic event (MBPE, see Figs. 22 and 23 and description below).

4.2. Magneto- and cyclostratigraphy of the Mid-Paleocene interval (J. Dinarès-Turell, J.I. Baceta, V. Pujalte)

In order to locate chron C26n along the mid-Paleocene succession a total of 101 samples from different stratigraphic levels were sampled using a gasoline-powered drill or as oriented hand-samples. This study extends upwards the integrated magnetostratigraphy presented by Dinarès-Turell *et al.* (2003) and represents the completion of the entire Paleocene Epoch interval, for which now all magnetic reversals are documented at a bed level, jointly with the uppermost Paleocene magnetostratigraphy from Dinarès-Turell *et al.* (2002) (see also chapters 3.2 and 5.3 in this guide). In order to increase the reliability of chron C26n position and its cycle duration along the magnetically weak (see below) mid-Paleocene interval, an additional section (Ibaeta) was also studied (23 sample levels). Natural remanent magnetization (NRM) and remanence through all demagnetization stages were measured using a 2G-Enterprises high-resolution (45 mm diameter) pass-through cryogenic magnetometer equipped with DC-squids and operating in a shielded room at the Istituto Nazionale di Geofisica e Vulcanologia in Rome, Italy.

In general, the intensity of the NRM was relatively low, around 4 x 10-4 A/m although it was lower for some of the carbonate-rich beds. Two magnetisation components can be recognized in most samples upon demagnetisation, in addition to a viscous magnetisation removed below 5 mT or 100º-150°C. Magnetic behaviour upon demagnetization was similar to the one documented for the uppermost Paleocene samples (Dinarès-Turell *et al.*, 2002). Component H conforms the ChRM and has either normal or reverse polarity in bedding-corrected coordinates (Fig. 24). Before bedding-correction, component H has a steep inclination not compatible with a Paleocene to recent geomagnetic field direction, whereas the mean ChRM inclination after bedding correction is near the expected Paleocene inclination for the site paleolatitude. The VGP latitude together with the lithologic log for both the Zumaia and Ibaeta sections is shown in Fig. 24.

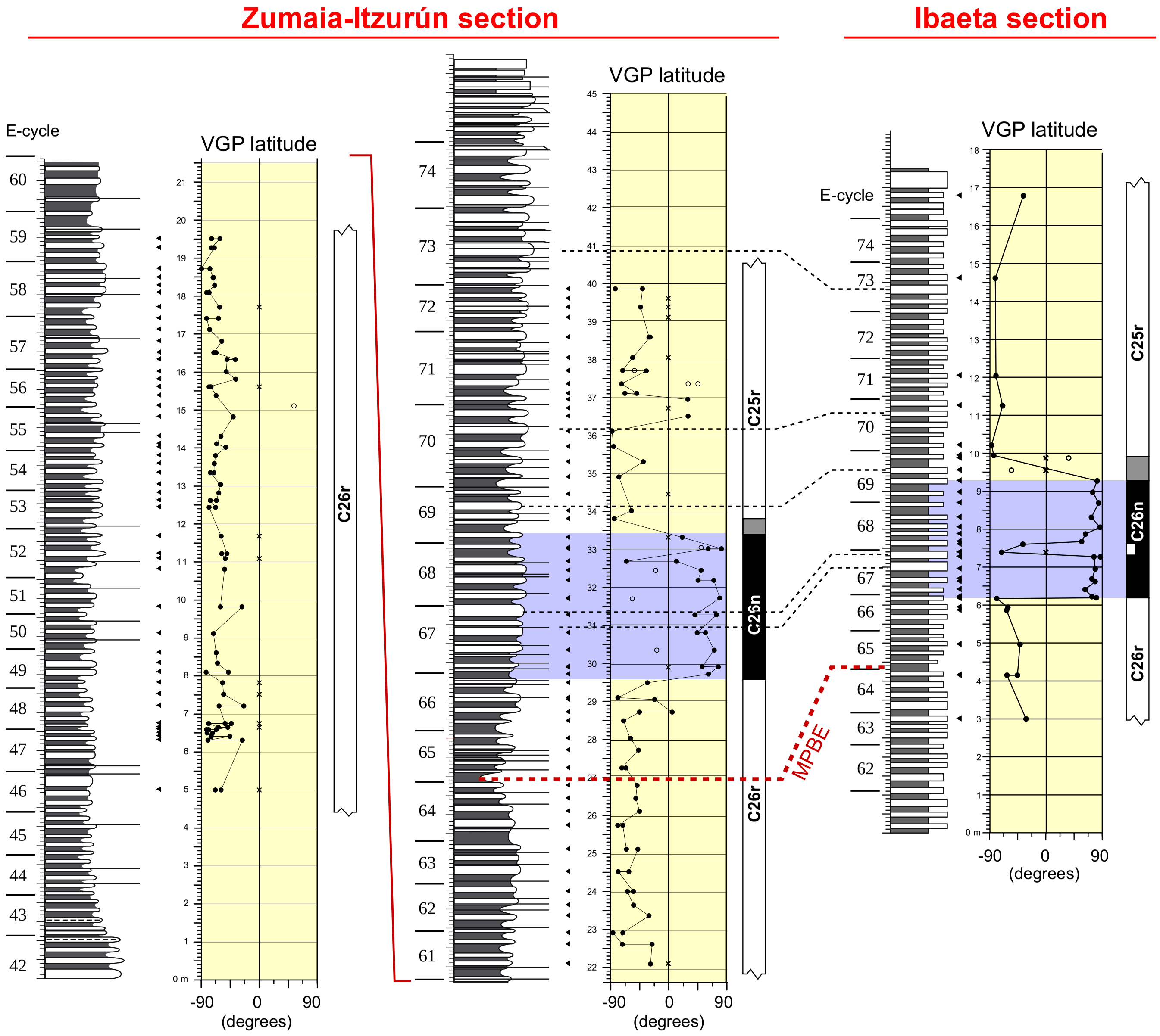


Fig. 24.- Virtual geomagnetic pole (VGP) latitude and lithologic logs for the Zumaia and Ibaeta sections. The position of chron C26n and correlation between both sections is shown. MPBE denotes the Mid Paleocene biotic event (see chapter 4.8 in this guide). Open circles denote unreliable data and crosses mark the position of samples that have provided no data. The numbering of the eccentricity (110 kyr) related E-cycles follow numbering for underlying strata that starts above the K/P boundary as reported by Dinarès-Turell *et al.* (2003).

The identification of the (~110 kyr) eccentricity related bundles (E-cycles) for the mid-Paleocene interval is shown in Fig. 24. Numbering of the E-cycles follows that of the underlying strata as reported in Dinarès-Turell *et al.* (2003) and in Fig. 20 in this guide. A calcium carbonate and low-field magnetic susceptibility proxy records along the mid-Paleocene interval is currently under evaluation in order to substantiate the cyclostratigraphic stacking pattern and long term quasi-periodic cycles with available orbital target solutions before attempting any tuning. However, a cycle duration of chron C26r and C26n and relative positioning of the Mid-Paleocene Biotic Event (MPBE) and biostratigraphic events (see below) can be established. Cycle counting for chron C26r tentatively provides about 31 eccentricyty ~110 kyr E-cycles or about 137 precession (~21 kyr) cycles. Chron C26n contains 11-12 precession cycles and the MPBE is located ~8 precession cycles below the base of chron C26n in correspondence to an eccentricity maxima if the phasing between the lithologic cycles and the orbital parameters hypothesized in Dinarès-Turell *et al.* (2003) for the lower Paleocene is correct.

4.3. Calcareous nannofossils assemblages across the Mid-Paleocene *(G. Bernaola, Y. Nuño-Arana)*

The calcareous nannofossil assemblages recorded in the mid-Paleocene of Zumaia section are well preserved, quantitatively rich and very diverse (Bernaola, 2002). Throughout the last 15 m of the Danian Limestone the assemblage is dominated by *C. pelagicus*, *Pr. martini* and *Br. bigelowi* that together with *Sph. primus*, *Ericsonia* spp., *Pr. bisulcus* and *Tw. pertusus* are the mayor components of the assemblage. In the uppermost Danian the calcareous nannofossil diversity is relatively high (S-H 2,1 and 26-31 species per sample), especially if we compare it with that registered in the lower Danian (Bernaola, 2002). Althought the calcareous nannofossil assemblage do not show significant changes during the upper Danian, few meters below the lithological change from the Danian Limestone to the overlying red marls of the Itzurun Fm a relative abundance decrease of *Br. bigelowi* and *Sph. primus* and a relative abundance increase of *Tw. pertusus is* registered. In the upper 15 m of the Danian Limestone we have recorded the FOs of *Sphenolithus primus*, Ch. bidens, and the first specimens of the genus *Fasciculithus* (Fig. 25).

Nch. perfectus first occur in the last marly interval of the Danian Limestone. This finding could be important in the location of the Danian/Selandian boundary at Zumaia section as the FO of this taxon in Zumaia concurs with it first appearance in the uppermost part of Danian limestones in Denmark (Thomsen (1994). However, the most important calcareous nannofosil assemblage turnover across the mid-Paleocene at Zumaia occurs in connection with the lithological change from the Danian Limestone to the overlying Itzurun Fm. At this lithological change, we observe a significant increase in the total abundance of calcareous nannofossils and, also, an abrupt decrease in the relative and total abundance of *Br. bigelowi,* that changes from being abundant in the uppermost Danian Limestone (from 8 to 22%) to be very rare in the Itzurun Fm (Fig. 25). In connection with these assemblage changes, we also note the diversification of the genera *Fasciculithus* (FOs of *F. bitectus*, *F. janni* and *F. ulii*) (Fig. 25).

In the red marl, the calcareous nannofossil assemblage is dominated by the same species that prevail in the Danian Limestone, with the exception of *Br bigelowi*. In this interval, besides the FOs of the abovementioned fasciculiths, we also observe the FOs of *Ch. consuetus, F. billi* and *F. tympaniformis*. The presence of *F. bitectus F. janni* and *F. ulli* is restricted to the red marl interval.

Close to the boundary between the red marl and the overlying grey marly beds, we observe a slight total abundace decrease in calcareous nannofossil but no other significant change, with the exception of the FOs of *F. pileatus* and *Tw. tovae.*

The calcareous nannofossil assemblage across the grey marlstones of the member A of the Itzurun Fm is similar to that found in the underlying red marl and is mainly dominated by *C. pelagi-*

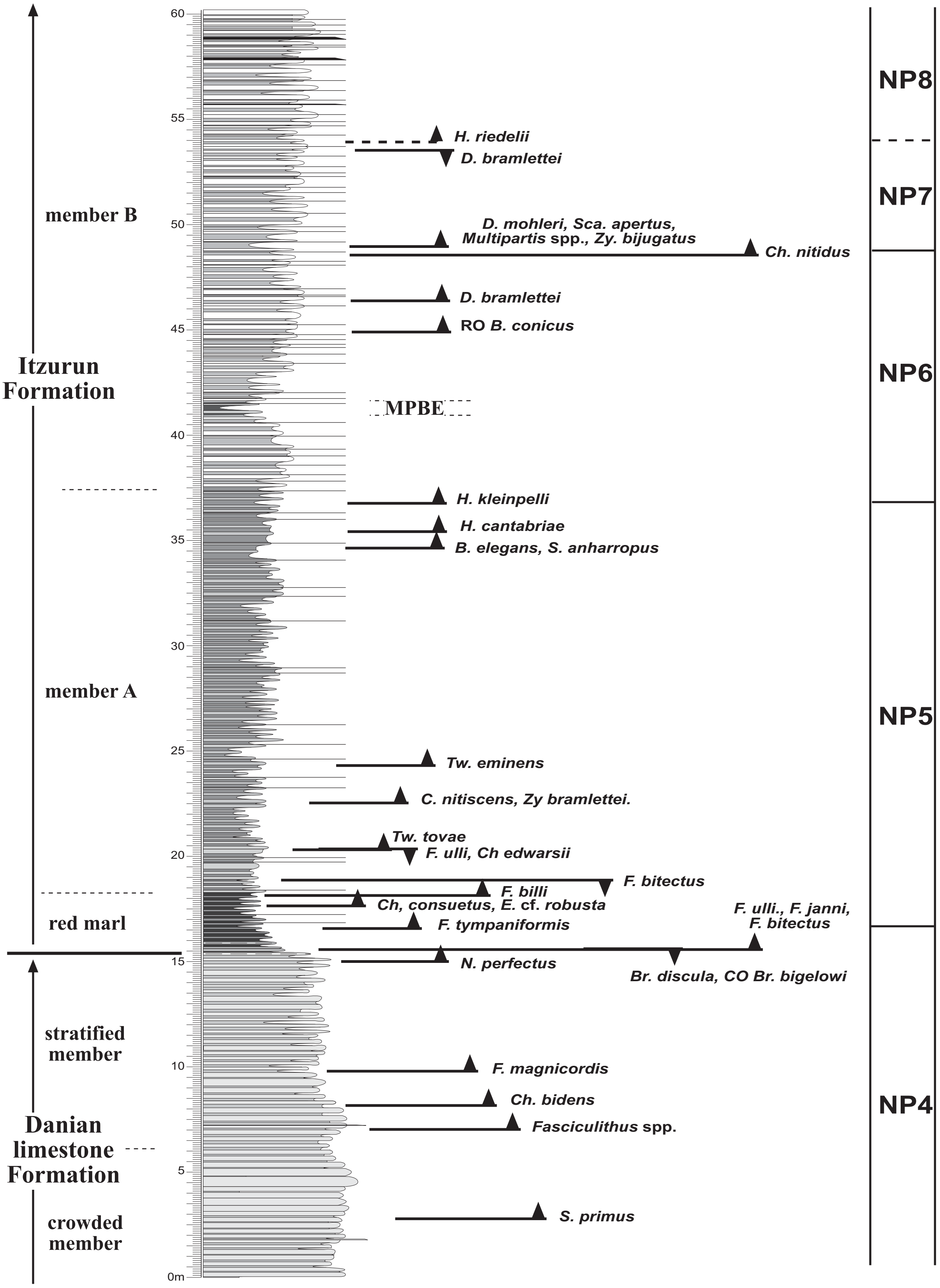


Fig. 25.- Main calcareous nannofossil bioevent across the mid-Paleocene interval in the Zumaia section. CO = common occurrence, RO = rare occurrence.

cus, *Pr. martini*, *Pr. bisulcus* and *Tw. pertusus.* The most important change in calcareus nannofossil assemblages across the A and B members of the Itzurun Fm is the occurrence and diversification of the genus *Heliolithus* and the first occurrence of the specimens of the genus *Discoaster*, one of the most important calcareous nannofossil group throughout the whole Paleogene. Across the upper A and the B members of the Itzurun Fm we also record the FOs of *C. nitiscens*, *Z. branlettei*, *Tw. eminens*, *B. elegans*, *H. cantabriae*, *S. anharropus*, *H. kleinpelli*, *B. conicus* and *D. bramlettei,* among others (Fig. 25).

4.4. New planktic foraminifera biozonation from the Mid-Paleocene *(X. Orue-Etxebarria, F. Caballero, E. Apellaniz)*

The high planktic foraminifera total abundance and species richness found in all the samples from the mid-Paleocene as well as the thickness of the Paleocene in the Zumaia section, have allowed the establishment of a new planktic foraminiferal biozonation. We have differentiated 5 biozones based on the first occurrence biohorizons (FOs) of selected species in the studied succession (Fig. 22).

Morozovella angulata **Interval Zone** (= lower part of P3a of Berggren *et al.*, 1995 and of Berggren & Pearson, 2005).

Definition: Biostratigraphic interval between the LO of *Morozovella angulata* and the LO of *Igorina pusilla*.

Remarks: In the studied section only the final part of this biozone is represented, and includes the lower 3 m of the succession. The most characteristic species of the assemblage are *M. angulata*, *M. conicotruncata*, *Subbotina triloculinoides*, *Parasubbotina varianta* and *Globanomalina ehrenbergi* among others.

Igorina pusilla **Interval Zone** (emended herein; =middle part of P3a of Berggren *et al.*, 1995 and of Berggren & Pearson, 2005).

Definition: Biostratigraphic interval between the LO of *Igorina pusilla* and the LO of *Morozovella occlusa*.

Remarks: This biozone is 10 m thick and has a planktic foraminifera assemblage similar to the previous biozone, also including *S. linaperta*, *S. velascoensis*, different igorinids and the first specimens of the genus *Acarinina.*

Morozovella occlusa **Interval Zone** (herein defined; =upper part of P3a of Berggren *et al.*, 1995 and of Berggren & Pearson, 2005).

Definition: Biostratigraphic interval between the LO of *Morozovella occlusa* (= *M. crosswicksensis*) and the LO of *Igorina albeari*.

Remarks: We use the FO of *M. occlusa* to mark the base of this 11 m thick biozone because this taxon is the most abundant of the *velascoensis* group in all the samples of the Middle and Upper Paleocene. *M. velascoensis* first occurs together with *M. occlusa*.

Igorina albeari **Interval Zone** (= P3b of Berggren *et al.*, 1995 and P3b and P4a of Berggren & Pearson, 2005).

Definition: Biostratigraphic interval between the LO of *Igorina albeari* and the LO of *Globanomalina pseudomenardii*.

Remarks: The first specimen of *I. albeari*, which mark the base of this biozone, appear 10 m above the lithological change from the Danian Limestone to the overlying red marls of the Itzurun Fm. *I. albeari* includes those forms with a biconvex trochospiral growth, more than 7 chambers in the last whorl with a peripherial margen with muricocarina in the last chambers and a high spiral side in the central part. This biozone is 7 m thick.

***Globanomalina pseudomenardii* Interval Zone** (= lower part of P4a of Berggren *et al.*, 1995 and of P4b of Berggren & Pearson, 2005).

Definition: Biostratigraphic interval between the LO of *Globanomalina pseudomenardii* and the LO of *Acarinina soldadoensis*.

Remaks: The upper part of the studied section belongs to the *Gl. pseudomenardii* biozone, whose base has been defined by the first occurrence of this taxon. *Gl. pseudomenardii* is very rare in the first samples of the biozone and the first specimens are relatively small and have an indistinct muricocarina.

In general, the planktic foraminifera total abundance and species richness, particularly those of morozovellids and igorinids, increase progressively from the upper part of the Danian Limestone to the top of the studied succession, in agreement with the ideas of Norris (1997) and Olsson *et al.* (1999). At the beginning of the *Gl. pseudomenardii* biozone there is a new increase in the planktic foraminifer diversity, especially in acarininids and globanomalinids.

On the other hand, the ratio of planktic foraminifera to the total foraminifera content increases from 0.8 in the lower part of the succession to 0.9 in the *Gl. pseudomenardii* biozone. Regarding to the Danian Limestone and the overlying red marl transition we do not find any important change in the number of first and last appearances of planktic foraminifer species.

4.5. Benthic foraminifera from the Danian-Selandian transition (S. Ortiz)

Benthic foraminifera from the Danian-Selandian transition in the >100 mm size fraction are better preserved in the Selandian marls (*Morozovella crosswicksensis* Zone of Arenillas 1996), where the benthic foraminiferal assemblages are also more diverse and heterogeneous. Deep-bathyal taxa, such as *Bulimina trinitatensis*, *Cibicidoides hyphalus*, *Cibicidoides velascoensis*, *Gyroidinoides globosus*, *Stensioeina beccariiformis*, *Pullenia coryelli*, *Nuttallides truempyi*, *Clavulinoides amorpha*, *Clavulinoides trilatera*, and *Gaudryina pyramidata*, are abundant, most of them typical from the Velasco-type fauna (van Morkhoven *et al.*, 1986) (Fig. 26). Taxa typical from sublitoral settings are also common (e.g., *Angulogavelinella avnimelechi*, *Osangularia plummerae*). Flysch-type agglutinated foraminifera, such as *Hyperammina dilatata* and *Rhabdammina* spp., and calcareous-cemented foraminifera (e.g., *Clavulinoides* spp.), are both common. Agglutinated assemblages are typical of the “low to mid-latitude slope” DWAF (Deep Water Agglutinated Foraminifera) of Kuhnt *et al.* (1989). These assemblages suggest deposition at the middle-lower part of the slope.

Calcareous taxa are common, especially at the Selandian (up to 80%), which mainly consist of planoconvex trochospiral morphotypes (e.g., *Nuttallides truempyi*, *Angulogavelinella avnimelechi*), biconvex trochospiral (e.g., *Stensioeina beccariiformis*, *Osangularia plummerae*), and cylindrical tapered morphotypes (e.g., *Bulimina* spp.). Agglutinated taxa are more abundant at the Danian (up to 95%), which mainly consist of cylindrical tapered morphotypes (e.g., *Clavulinoides trilatera*, *Clavulinoides amorpha*, *Marssonella* spp., *Karrerulina horrida*) and rounded trochospiral/streptospiral morphotypes (e.g, *Cribrostomoides trinitatensis*, *Recurvoides* spp.).

Epifaunal and infaunal morphogroups show constant values close to 50% throughout the section, suggesting intermediate trophic levels and well-oxygenated waters at the seafloor. A transition from oxic to suboxic conditions at the seafloor has been recorded in several shallow-sea settings worldwide. Further studies are needed to corroborate whether these paleoenvironmental changes are global at different paleodepth settings.

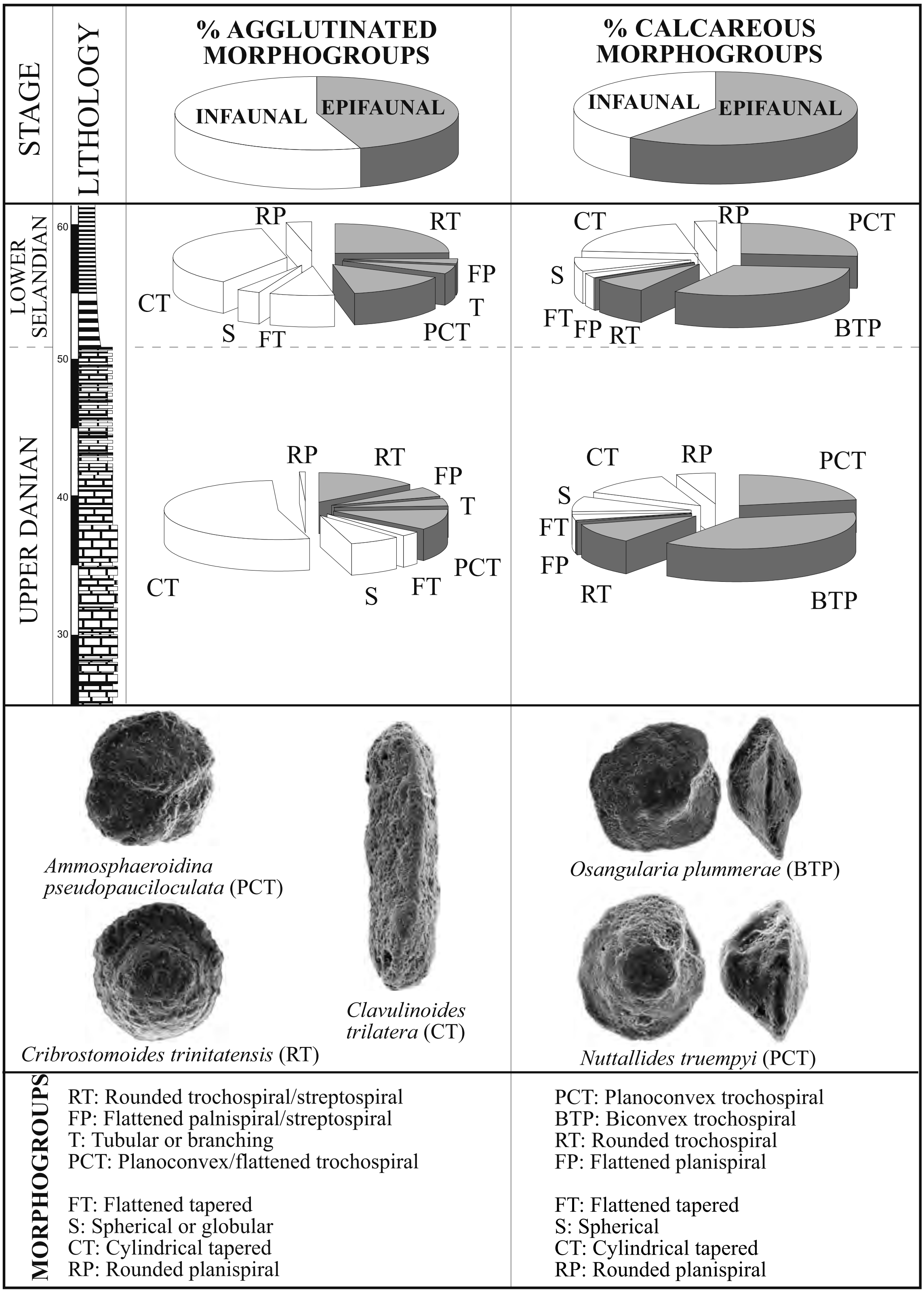


Fig. 26.- Benthic foraminiferal assemblagles across the Danian/Selandian transition in the Zumaia section

4.6. Benthic foraminifera from the Selandian/Thanetian transition at Zumaia (L. Alegret, M. Martín-Rubio)

Samples from the Selandian-Thanetian transition at Zumaia contain well-preserved benthic foraminifera in the >63mm size fraction. Benthic assemblages are highly diverse and heterogeneous through the studied interval. They contain abundant representatives of the bathyal to abyssal Velasco-type fauna, as well as species that are common at bathyal settings. Nevertheless, the presence of species that are common a shallower depths (e.g., *Angulogavelinella avnimelechi*) suggests deposition in the middle part of the slope (Fig. 27).

Assemblages contain 50-60% of calcareous taxa, which mainly consist of elongated tapered morphotypes (e.g., laevidentalinids, *Pleurostomella* spp., *Pyramidina rudita*), planoconvex trochospiral (e.g., *Angulogavelinella avnimelechi, Cibicidoides hyphalus, C. howelli, C. proprius, Gyroidinoides beisseli, Nuttallides truempyi, Nuttallinella florealis*), and biconvex trochospiral morphotypes (e.g., *Anomalinoides* spp., *Cibicidoides pseudoperlucidus, Osangularia* spp., or *Stensioeina beccariiformis*).

Agglutinated assemblages are typical of the “low to mid-latitude slope” DWAF (Deep Water Agglutinated Foraminifera) biofacies of Kuhnt *et al.* (1989), and they are mainly dominated by flattened trochospiral (trochamminids), streptospiral (*Recurvoides* spp.) and rounded planispiral forms (*Haplophragmoides walteri, Haplophragmoides* spp.). Cylindrical tapered (*Clavulinoides, Dorothia, Marssonella*), elongate tapered (*Karrerulina, Reophax, Subreophax*), tubular or branching (*Bathysiphon, Hyperammina, Rhizammina*) and flattened planispiral and streptospiral forms (*Ammodiscus, Glomospira, Repmanina*) are also common to abundant. Agglutinated assemblages mainly consist of morphogroups with an inferred active deposit feeding habit, with a small percentage of suspension feeders (Kaminski and Gradstein, 2005).

Both agglutinated and calcareous assemblages consist of mixed infaunal and epifaunal morphogroups, suggesting that mesotrophic conditions occurred at the seafloor during most of the studied interval. Nevertheless, changes in the relative abundance of various benthic foraminiferal species and morphogroups in a short interval in coincidence with the Selandian/Thanetian transition, where a dramatic decrease in the planktic/benthic ratio is observed, suggest a major disruption in the paleoenvironmental conditions at the seafloor at that time. The question remains whether changes in composition of benthic foraminiferal assemblages (and the related paleoenvironmental inferences) can be linked to major changes occurred not only in planktic groups in this area, but also to the environmental changes inferred from deep-sea cores worldwide.

4.7. The Mid-Paleocene Biotic Event (MPBE) (G. Bernaola, J.I. Baceta, X. Orue-Etxebarria, L. Alegret, M. Martín-Rubio, J. Arostegi, J. Dinarès-Turell)

Recent studies of deep-sea sections at Central Pacific (Shatsky Rise) and South Atlantic (Walvis Ridge) oceans have confirmed the existence of several dissolution levels interspersed in the carbonate sediments of the late Paleocene-early Eocene (Eocene Layer of Mysterious Origin, ELMO; X-event layer; Mid-Paleocene Biotic Event layer). Similarly to the Paleocene-Eocene Thermal Maximum (PETM) interval, these levels are characterized by the presence of a clay-rich layer, associated with an abrupt drop in carbonate content, and a pronounced peak in magnetic susceptibility (MS) that reflects an increase in clay content. All these dissolution intervals have been interpreted as the sedimentary response to abrupt climatic changes; in fact, the few micropaleontologic and geochemical analyses carried out hitherto suggest that they represent short lived events of considerable evolutionary significance probably related to hyperthermal events.

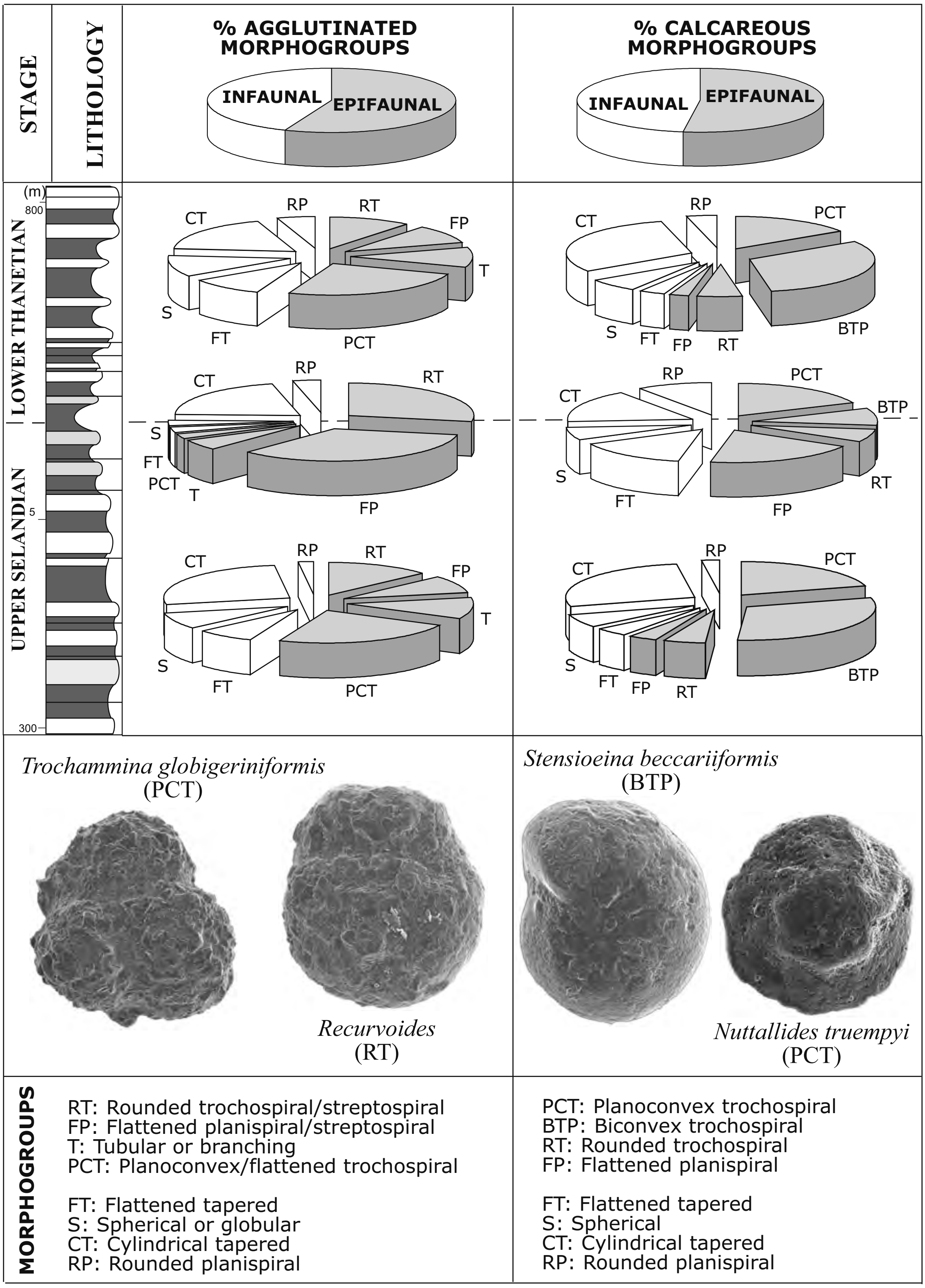


Fig. 27.- Benthic foraminiferal assemblagles across the Selandian/Thanetian transition in the Zumaia section

Developing studies in the mid-Paleocene succession of Zumaia (Western Pyrenees) have revealed the presence of a 1 m thick clay-rich interval characterized by a drop in carbonate content, a peak in MS and important calcareous nannofossil and foraminifer assemblage changes. This interval is located 4 m above the FO of *Heliolithus kleinpelli*, the marker of the base of Zone NP6, 2 meters below the C25r/C26n reversal and within thelower part of the Zone P4 of planktic foraminifers. According to this data, it might be equivalent to the Mid-Paleocene Biotic Event found at Shatsky Rise and Walvis Ridge.

The micropaleontologic analyses show an important drop in the planktic/benthic ratio in connection with the clay-rich interval, from 90% to less than 10%. During the event, planktic foraminifers form a low diversity assemblage dominated by subbotinids. Benthic foraminifer assemblages are characterized by a maximum in the percentage of agglutinated taxa, an increase in the percentage of calcareous taxa *Nuttalides truempyi* and *Stensioeina beccariformis* and a decrease of agglutinated suspension feeders and calcareous infaunal groups. Finally, calcareous nannofossils show important assemblage changes in connection with the clay-rich interval, mainly defined by a sharp increase of *C. pelagicus, Ericsonia* spp. and *Heliolithus* spp. and a decrease of *Tw. pertusus* and *Prinsius* spp.

According to preliminary cyclostratigraphic data, this event (MPBE) is located ~8 precession cycles below the base of chron C26n in correspondence to an eccentricity maxima.

4.8. Looking for a suitable criterion to define the Danian/Selandian & Selandian/Thanetian boundaries (G. Bernaola, X. Orue-Etxebarria, V. Pujalte, J.I. Baceta)

During the nineteenth and early twentieth centuries many different Paleogene stages were defined. The most well-known Paleocene historical stages, e.g., the Danian, Montian, Heersian, Sparnacian, Thanetian, Landenian, and Selandian, have their type areas in northwestern Europe where they were mostly defined at unconformities corresponding to sequence boundaries. At a meeting in Washington, D.C., 1989 it was voted and agreed on by the International Subcommission on Paleogene Stratigraphy (ISPS) that the Paleocene shall be divided into three stages: the Danian, Selandian, and Thanetian (Jenkins & Luterbacher 1992). In fact, twentieth century research has shown that the Danian, Selandian, and Thanetian together represent most of Paleocene time.

Later, in the symposium on "Stratigraphy of the Paleocene" held in Göteborg (Sweden) on 1993, the principies for dividing the Paleocene into stages were discussed and it was formally decided a three-fold división of the Epoch.

When selecting GSSP's it is important to consider the potential utility of the new global stage concepts, however, historical tradition should be respected. The respective time-rock extents of the new Selandian and Thanetian GSP-delimited stages should, ideally, reflect global change, but also roughly coincide with the stratigraphic extents of the respective historical, regional stages (Knox, 1994).

In the type region, Denmark, the base of the Selandian is marked by an unconformity between the limestones of the Danskekalk Fm and the greensands and marls of the Lellinge Fm (Thomsen and Heilmann-Clausen, 1985). The limestone/chalk deposition had prevailed throughtout the entire Maastrichtian and Danian. After the Da/Se transition, grey marls and clays dominated sedimentation for several million years. In terms of calcareous nannofossil zonations, the basal Selandian in Denmark falls in the upper part of Zone NP4 or to the lower part of Zone NP5 (Thomsen, 1994).

Prestwich (1850-54) introduced the term "Thanet Sands" to describe the succession of fine, locally glauconitic and argillaceous sandstones and sandy clays wich overlie the Upper Cretaceous chalk in the Isle of Thanet and adjcent districts in southern England. In its original type area, the unconformity related base of the Thanetian has been dated with the upper part of the nannofossil zone NP6 (Aubry, 1994; Knox, 1994).

All the events traditionally used to place the Danian/Selandian (i.e., the top of chron C27N, FO of *M. angulata* and *I. albear*i planktic foraminifera, The FO of *F. tympaniformis*) and Selandian/Thanetian (i.e., the base of chron C26N) boundaries can be identified in the Zumaia section (Figs. 28 and 29). The study of the mid-Paleocene of the Zumaia section has revealed that all these events occur forming a distinct sequence. Before selecting a section to place the Danian/Selandian and Selandian/Thanetian boundary stratotypes, the criterion to identify these boundaries should be precisely defined. A first step should be the description of the succession of the events that characterizes both boundary intervals. In this work the succession of events that characterised the Danian/Selandian boundary has been informally considered as extending from the base of Chron C27n to the FO of *I. albeari* (Fig. 28). On the other hand, the succession of events that characterised the Selandian/Thanetian boundary has been informally considered as extending from the FO of *Gl. pseudomenardii* to the FO of *D. mohleri* (Fig. 29).

Evaluating the pros and cons, and the usefulness of each of the events identified at Zumaia is beyond the scope of this fieltrip guidebook, as this decision is in the domain of the International Subcommission on Paleogene Stratigraphy (ISPS). Consequently, only the relative position of the different events suitable to be selected as "official" markers of the Da/Se and Se/Th boundaries, and their correlation with other zonal scales, are highlighted in figures 28 and 29.

For the Danian/Selandian boundary, the Zumaia section provides 9 different events as suitable criteria, from older to younger: a) FO of *I. pusilla* and/or *S. primus,* b) top of chron C27n, c) FO of *M. occlusa*, d) FO of *Nch. perfectus*, e) mid-Palecocene sea level fall with the associated paraconformity in deep-basin areas and the mid-Paleocene unconformity (MPU) in near shore areas, f) diversification of fasciculiths, g) FO of *F. tympaniformis*, h) δ^{13}C relative minimum and i) FO of *I. albeari* (Fig, 28).

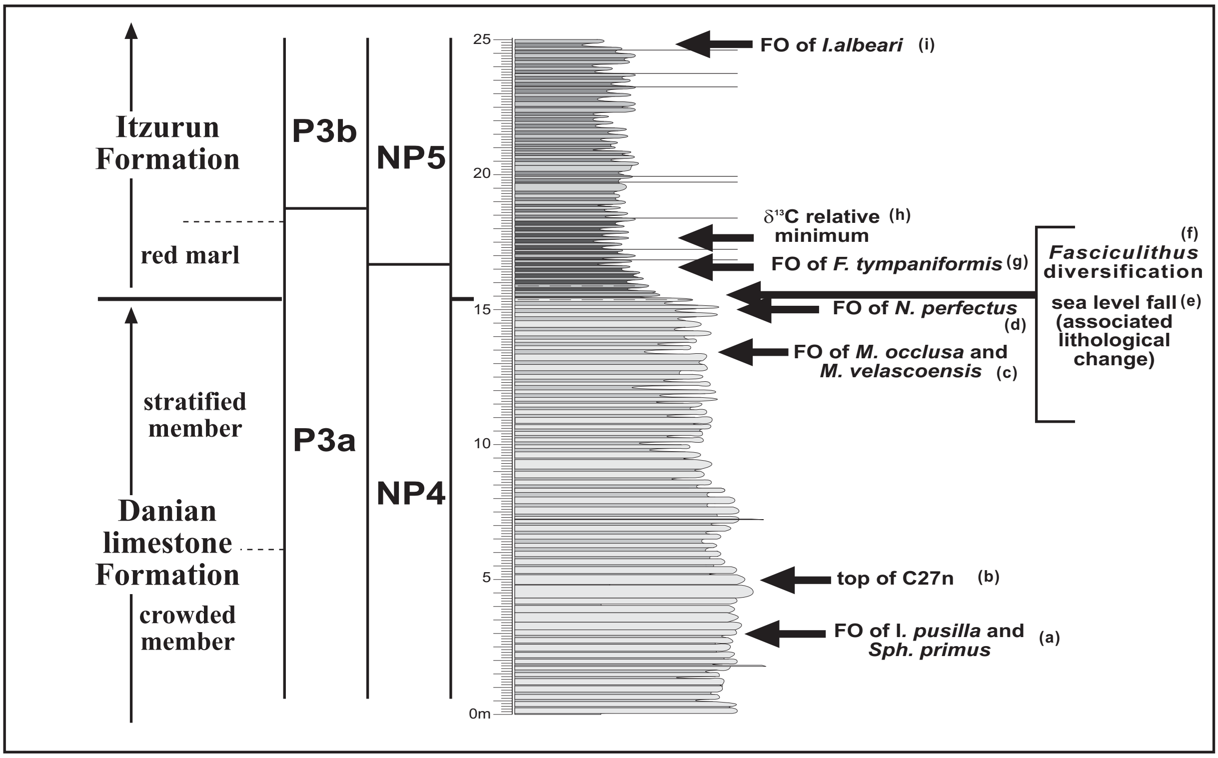


Fig. 28.- Suitable criteria to define the base of the Selandian

As is shown in figure 28, six of the nine events are located within an interval of 5 meters encompassing the uppermost Danian Limestone and the lowermost Itzurun Fm. At Zumaia the FO of *M. angulata*, usually used to define the base of the Selandian stage, is found 8 m below the top of chron C27n.

For the Selandian/Thanetian boundary, the Zumaia section provides 6 different events as suitable criteria, from older to younger: a) FO of *P. pseudomenardi*i, b) FO of *H. kleinpelli*, c) The MPBE, d) base of chron C26n, e) top of chron C26n and FO of *D. mohleri* (Fig. 29).

As is shown in figure 29, the six events are located within an interval of 17,5 meters encompassing the uppermost A and the lowermost B memberf of the Itzurun Fm.

The Zumaia section satisfies most of the infrastructural, biostratigraphic and geological requirements listed by the International Commission on Stratigraphy (ICS) (Remane *et al.*, 1996), as demonstrated above (see section 1.2 of this guide). In particular, the great sedimentary thickness and expanded charactyer of the section, have allowed the accurate location of successive biostratigraphic and magnetostratigraphic events and their chronological succession can be easily established.

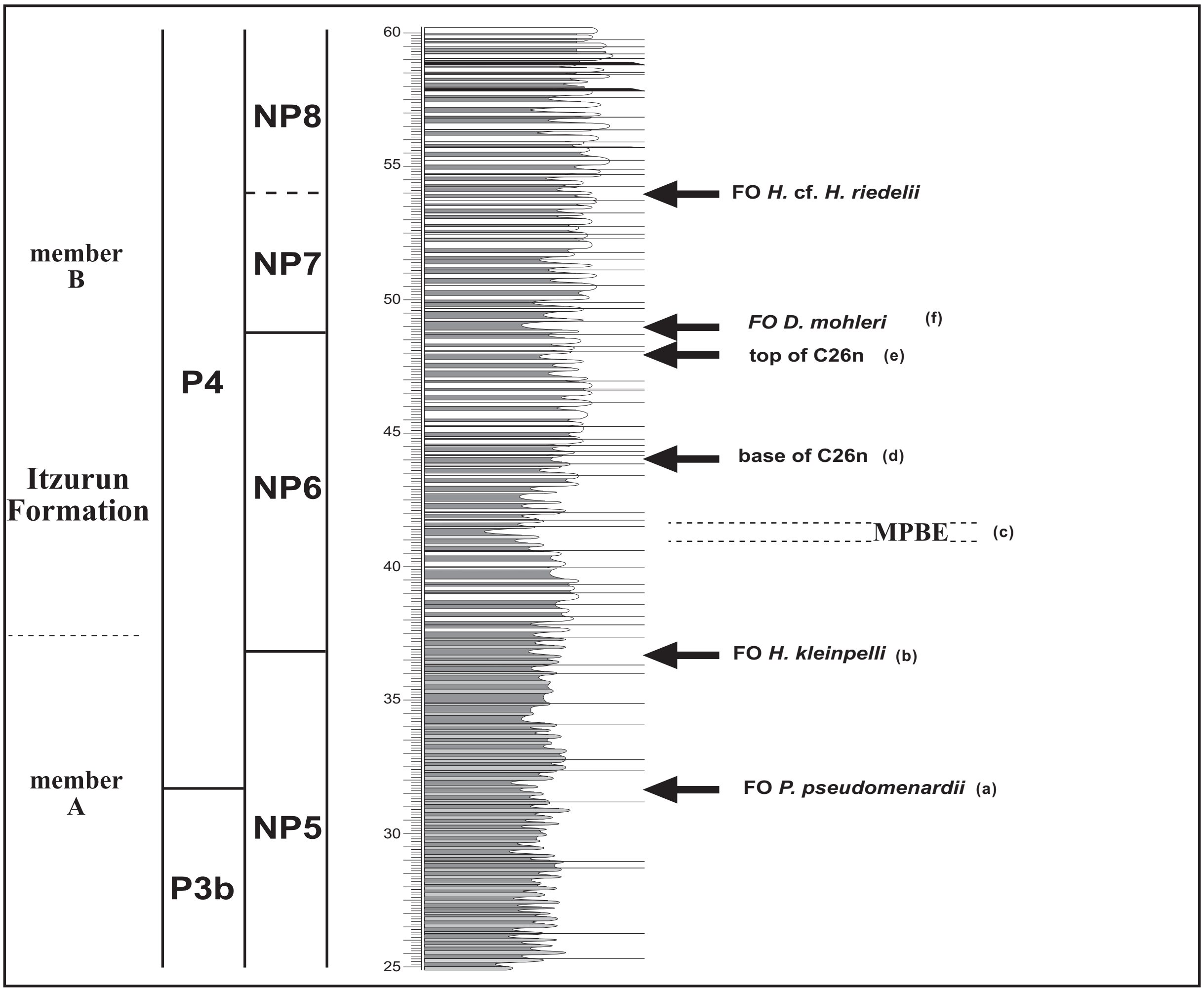


Fig. 29.- Suitable criteria to define the base of the Thanetian

5 THE PALEOCENE/EOCENE BOUNDARY INTERVAL

5.1. An overview (G. Bernaola, X. Orue-Etxebarria)

The P/E boundary transition, characterized by a prominent negative excursion of $\delta^{13}C$ values (CIE) in both marine and continental records (Stott *et al.*, 1990; Kennet & Stott, 1991; Koch *et al.*, 1992; 1995; Bralower *et al.*, 1995; Thomas & Shakleton, 1996; Beerling & Jolley, 1998), has focused the interest of the scientific community in the last two decades. This interval is known as the PETM (Paleocene/Eocene Thermal Maximum) and is one of the most abrupt and transient extreme climate global events documented in the ancient geologic record. According to recent works the PETM only lasted for ~220 k.y. (Röhl, 2000, Agnini *et al.*, 2005), and is characterized by extremely warm climatic conditions, drastic changes in the global carbon budget, modifications in oceanic circulation and chemistry (possible change in the thermocline and circulation in the oceans) and reorganization of ecosystems. In fact, this event had an important biotic impact on both land and marine organisms, affecting groups as diverse as land mammals (e.g. Wood *et al.*, 1941; Clyde & Gingerich, 1998), deep-water benthic foraminifers, (Thomas, 1990; 1998; Thomas & Shackleton, 1996; Speijer *et al.*, 1996), benthic communities of middle and outer neritic marine settings, (Speijer 1994; Speijer *et al.*, 1996; Speijer and Schmitz, 2000), and shallow marine larger foraminifera fauna (Orue-Etxebarria *et al.*, 2001; Pujalte *et al.*, 2003).

The PETM also affected the photic zone, where various groups of planktic marine organisms suffered considerable turnovers. Dinoflagellates showed a global dispersion and an abundance increase of the representatives of genus *Apectodinium* in coastal environments (Crouch, *et al.*, 2001), planktic foraminifera experienced a radiation of shallow-dwelling tropical taxa (Kelly *et al.*, 1996; 1998; 2005; Berggren & Ouda, 2003) and the calcareous nannoplankton communities underwent an important assemblage turnover that lasted for the duration of the event (Aubry, 2001; Bralower, 2002; Kahn & Aubry, 2004; Orue-Etxebarria *et al.*, 2004).

Studies conducted in the last decade have revealed the Pyrenean area as a key region to understand the P/E boundary events. Besides its strategic paleogeographical position, approximately 35ºN in

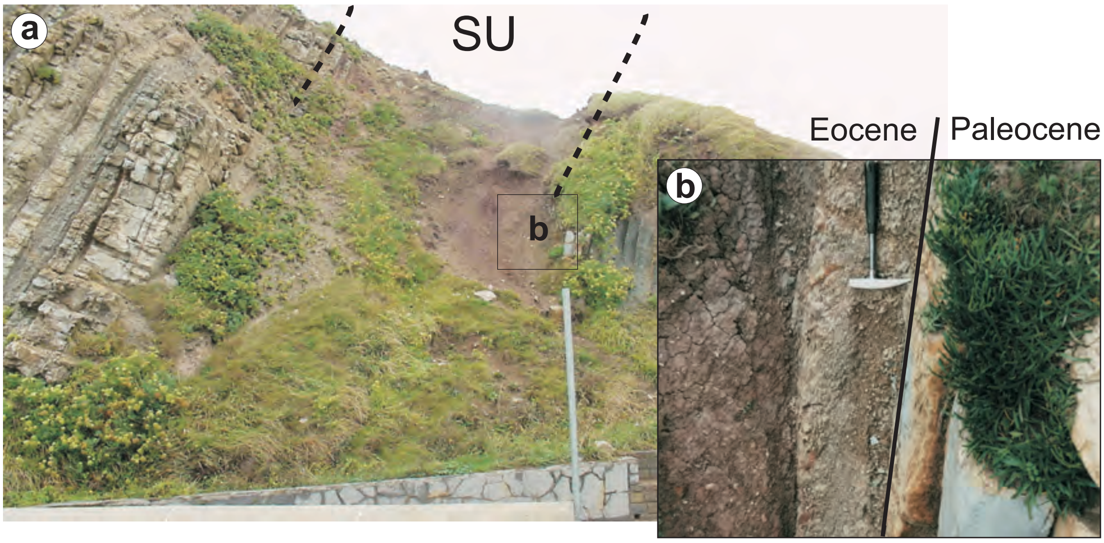


Fig. 30.- a) Outcrop photograph of the P/E boundary interval at Zumaia showing the Siliciclastic Unit (SU). (b) Detailed view of the P/E boundary.

the net evaporation zone and between the classic stratotypic sections of the NW Europe and the newly proposed GSSP in Egypt, the Pyrenean area has numerous outcropped successions, usually expanded and of good quality of exposure, that represent a wide range of settings, from continental to deep marine. A high-resolution correlation between continental, shelfal and bathyal marine successions by means of the CIE has proved that a major turnover of larger foraminifers occurred during the PETM. This fact demonstrates that the impact of the PETM upon the evolution of benthic fauna was grater than previously thought.

5.2. The Paleocene/Eocene boundary at Zumaia section (*J.I. Baceta, V. Pujalte, G. Bernaola*)

The Zumaia section is one of the most expanded and biostratigrafically complete P/E sections with deep water marine deposits so far reported.

The outcrop of the P/E boundary interval (P/E BI) is located partly at the Itzurun beach and partly along the NE side of the footpath from the village of Zumaia to the beach (latitude/longitude 42º18.00’N/2º15.30’W) (Fig. 30). In fact the footpath splits the outcrop in two segments, referred to below as the southern and northern ones (Fig. 31). The lower part of the P/E BI occurs in the northern segment.

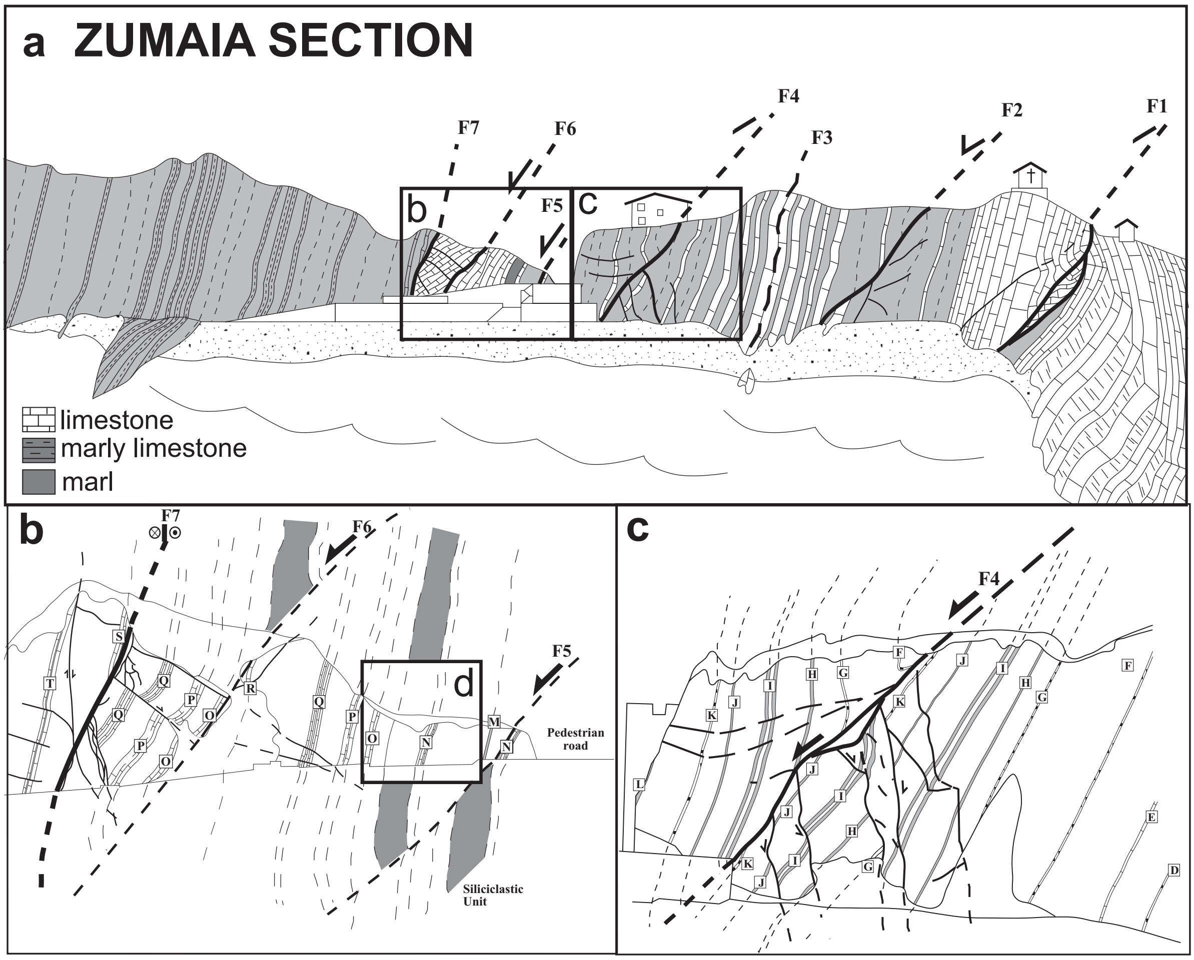


Fig. 31.- a) Field-sketch of the outcrop, in the cliffs of the Itzurun beach, Zumaia section. The limestones in the righ hand (SW) side of the outcrop are of Danian age, while the turbiditic succession in the left hand (NE) side correspond to the early Eocene. b) Field sketch of the northern segment of the P/E boundary interval. c) Field sketch of the southern segment of the P/E boundary interval.

The existence of several small faults or «tectonic disturbance zones» has been quoted in most published descriptions of the Zumaia section (e.g., Hillebrandt, 1965; Gawenda *et al.*, 1999). However, only Baceta *et al.* (2000) made an specific mention about how these faults omit or repeat parts of the stratigraphic succession.

The main tectonic disturbance in this outcrop is the fault labeled F4, which is accompanied by a swarm of smaller, satellite faults. Clearly, the main effect of fault F4 has been the duplication of the interval comprised between reference beds F and K (Fig. 31c). Our data show that this duplication would not be detected by biostratigraphic methods, as the whole duplicated interval is situated within the same planktic foraminiferal and calcareous nannofossil biozones. Nevertheless, failure to recognize the tectonic duplication would result in an overestimated stratigraphic thickness. It is also relevant to mention that the lower half of this outcrop is not longer visible due to the construction of a concrete staircase descending to the beach. Therefore, the duplication illustrated in figure 31 is even more difficult to appreciate nowdays than it was 12 years ago.

The northern outcrop is affected by three additional faults (F5, F6 and F7 in 31b). The attitudes and the effects of faults F5 and F6 are similar to that of fault F4, having both produced duplications of parts of the succession (Fig. 31).

Fault F4 is different from faults F4, F5 and F6, and no match can be attained between the stratigraphic successions separated by fault F7 (Fig. 31b). Such a mismatch is clear proof that fault F7 omits an interval of the stratigraphic succession. The extent of this missing interval is thought to be comparatively small.

Thus, it can be affirmed that most faults segmenting the P/E BI of the Zumaia section just repeat parts of the succession, the only exception being the minor omission of fault F7. Properly dealt with, the disruption of these fauts can be removed, and the original stratigraphic succession can be successfully reconstructed. To facilitate the identification of the different intervals of the succession, we have cross-referenced the litholog and the outcrop section (letters D to T in figures 31 and 35).

Following the usage of some previous authors, we have placed the zero level of our litholog at the horizon in which Schmitz *et al.* (1997) found the onset of the negative Carbon Isotopic Excursion (CIE). The deposits below and above our zero level belong respectively to the Paleocene and to the Eocene (Fig. 32).

The bulk of the P/E boundary interval is represented by rhythmic alternations of hemipelagic limestones and marlstones, plus numerous intercalations of thin-bedded turbidites (mostly 1-4 cm thick, but some 10 cm thick, Baceta *et al.*, 2000). The uppermost part of the Paleocene is represented by a 0.8 m thick unit (Unit III) mostly composed by hemipelagic limestones that include a 4 cm thick turbiditic level. This unit, that in fresh cuts presents a greenish color due to the high glauconite content, is a widespread marker bed observed in several other sections of the Basque Basin and used in regional correlations (Baceta, 1996; Pujalte *et al.*, 1998; Smichtz *et al.*, 2000). The green limestone is capped by a 35-40 cm thick marlstone bed (Unit IV), the base of which contains the onset of the CIE and the top the begining of the Benthic Foraminiferal Extinction Event (BEE) (Schmitz *et al.*, 1997). The marls have a greenish brown appearance, quite different from the grey marls below (Units I and II).

Above the 40 cm thick marlstone of the lowermost Eocene (Unit IV), there is a 4 m thick interval (Unit V) composed almost exclusively of claystones and silty claystones called “Siliciclastic Unit” (SU). The transition from marl to clay is gradual over about 5-10 cm. The lower 1.3 meters of the SU are almost devoid of carbonate microfossils. In the upper 2.7 meters of the SU the carbonate content increases slightly but relatively low values persist (Schmitz *et al.*, 1997).

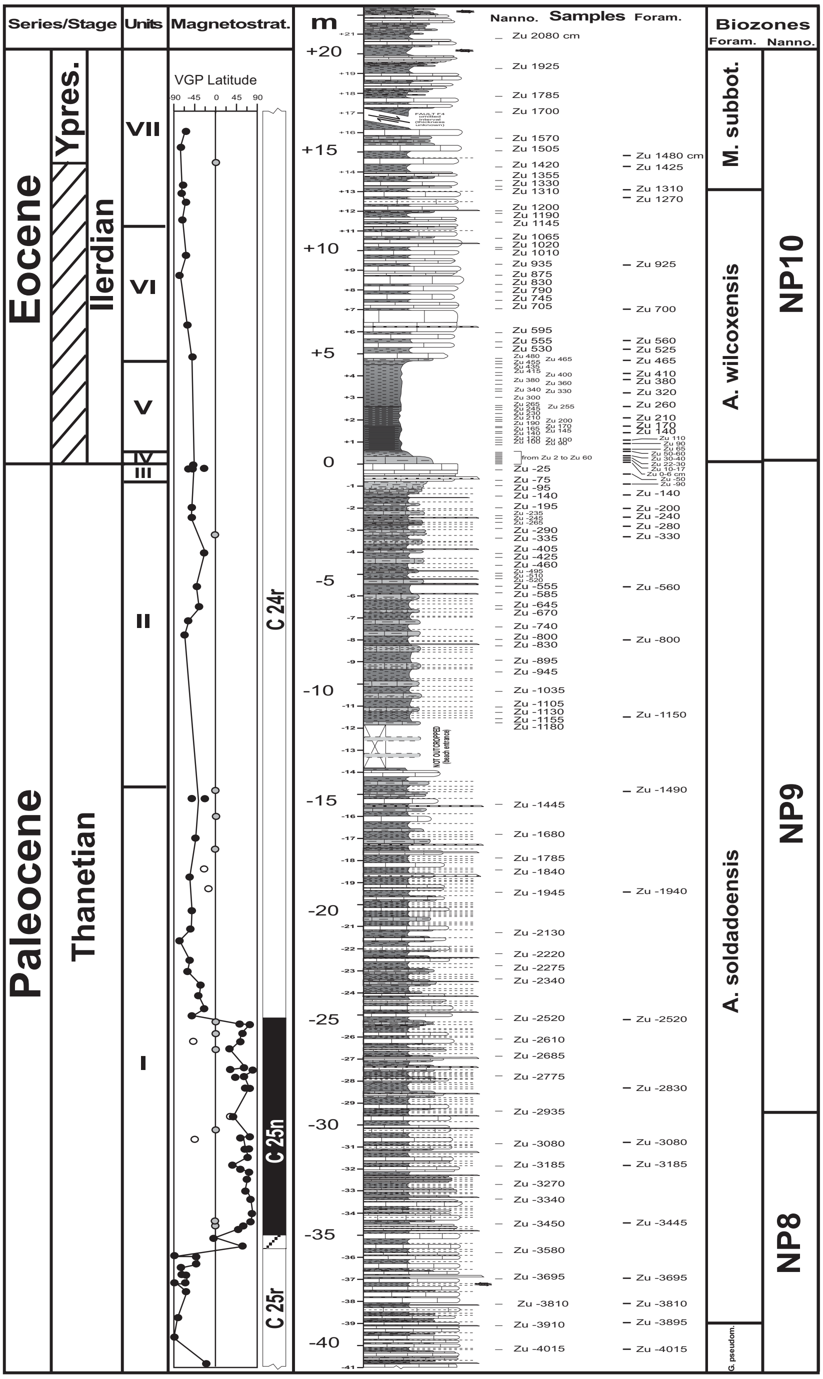


Fig. 32.- Lithostratigraphic column and magnetostratigraphy of the Zumaia Paleocene/ Eocene transition. The position of calcareous nannofossil and planktic foraminifer samples are shown separately.

Above the SU, and after a rapid and progressive transition, the succession is composed again by rhythmic alternations of hemipelagic limestones and marlstones plus some intercalations of thin-bedded turbidites (Unit VI), particularly from meter 10 onward (Unit VII).

The $CaCO_3$ $\delta^{13}C$ and $\delta^{18}O$ measurements across the P/E BI at Zumaia section were carried out by Schmitz *et al.* (1997) (Fig. 33).

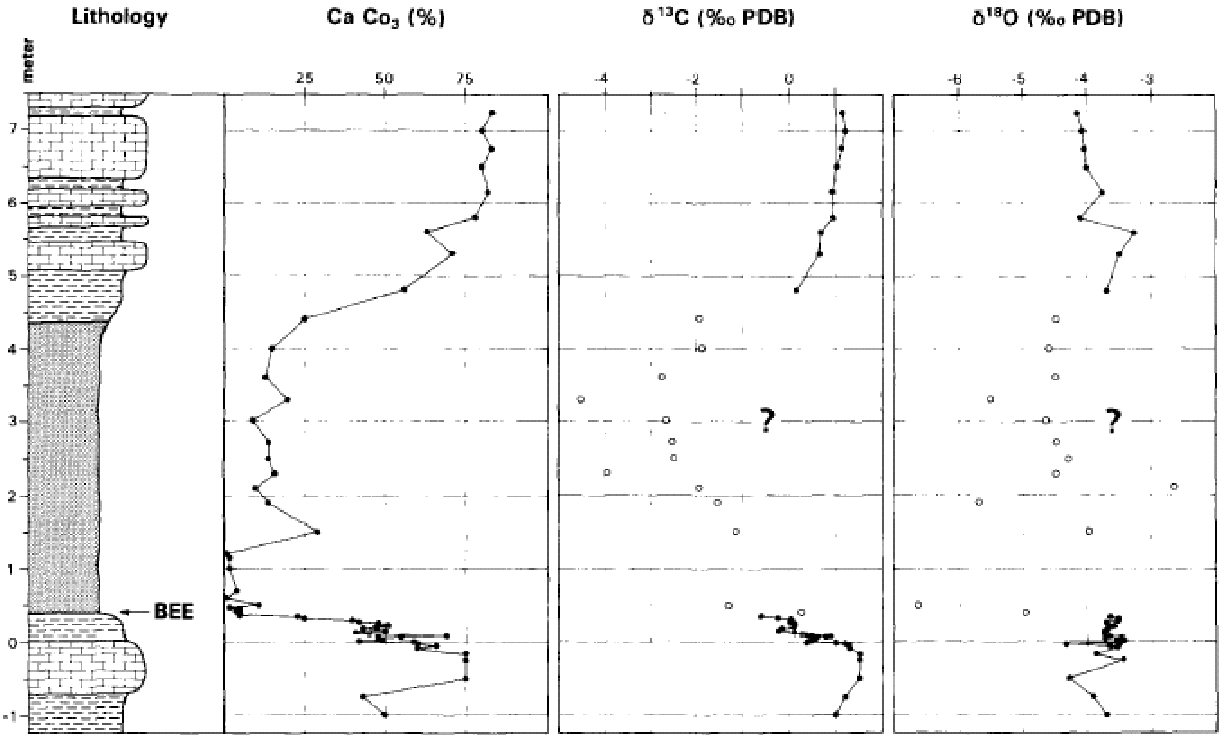


Fig. 33.- CaCO$_3$, δ^{13}C and δ^{18}O profiles across the P/E boundary interval at the Zumaia section (from Schmitz *et al.*, 1997)

According to these authors, calcite concentrations lie typically at 75-80% in the limestones and 40-50% in the marls (Units I, II and III). In the marls between +0.30 and 0.35 m (unit IV) calcite falls to 25% (Fig. 33). At +0.35-0.40 m calcite drops to 5%. In the lower 1 m of the clay interval calcite concentrations lie typically in the range 0-5%. In the upper 3 m there are many horizons with 10-20% calcite (Fig. 33). Examination of sieving residues (>63 µm) from the clay interval reveals common diagenetic calcite.

The δ^{13}C record shows a relatively stable pattern throughout the 11 m interval below the zero level (units I, II and III), with most values in the range 1-1.7‰. From the topmost few centimetres of the limestone and across the 30-40 cm of marls immediately below the benthic extinction event (unit IV), δ^{13}C shows a negative shift on the order of 1.4 -1.8‰ (Fig. 33). From values around 1.2‰ in the uppermost few centimetres of the limestone, δ^{13}C falls to values in the range -0.2‰ to -0.6‰, in the marl interval between +0.30 and 0.35 m. In the poorly calcareous parts of the overlying clay interval δ^{13}C values continue to fall (except for the +0.35 0.40 m sample), reaching extremely negative values of -3.9‰ to -4.5‰ at some levels in the interval 2-3.3 m above the zero level (Fig. 33). In the marls above the dissolution clay δ^{13}C returns gradually to more positive values and thereafter, throughout the overlying 8 m of marl limestone, fluctuates around 1‰, (Fig. 33).

The δ^{18}O values are relatively stable throughout the marl and limestone parts of the section, with most values in the range -3.5‰ to -4‰ (Fig. 33). In the clay interval, however, δ^{18}O shows widely scattered distribution, ranging between -2.6‰ and -6.6‰. The wide scatter indicates that diagenesis may have severely affected the original isotopic signals in this part of the section.

In a 1 cm thick layer between units III and IV Schmitz *et al.* (1997) found a small Ir anomaly. The higest Ir concentration measured in the layer was 157 ppt, compared to an average value of 51 ppt, for the rest of the section.

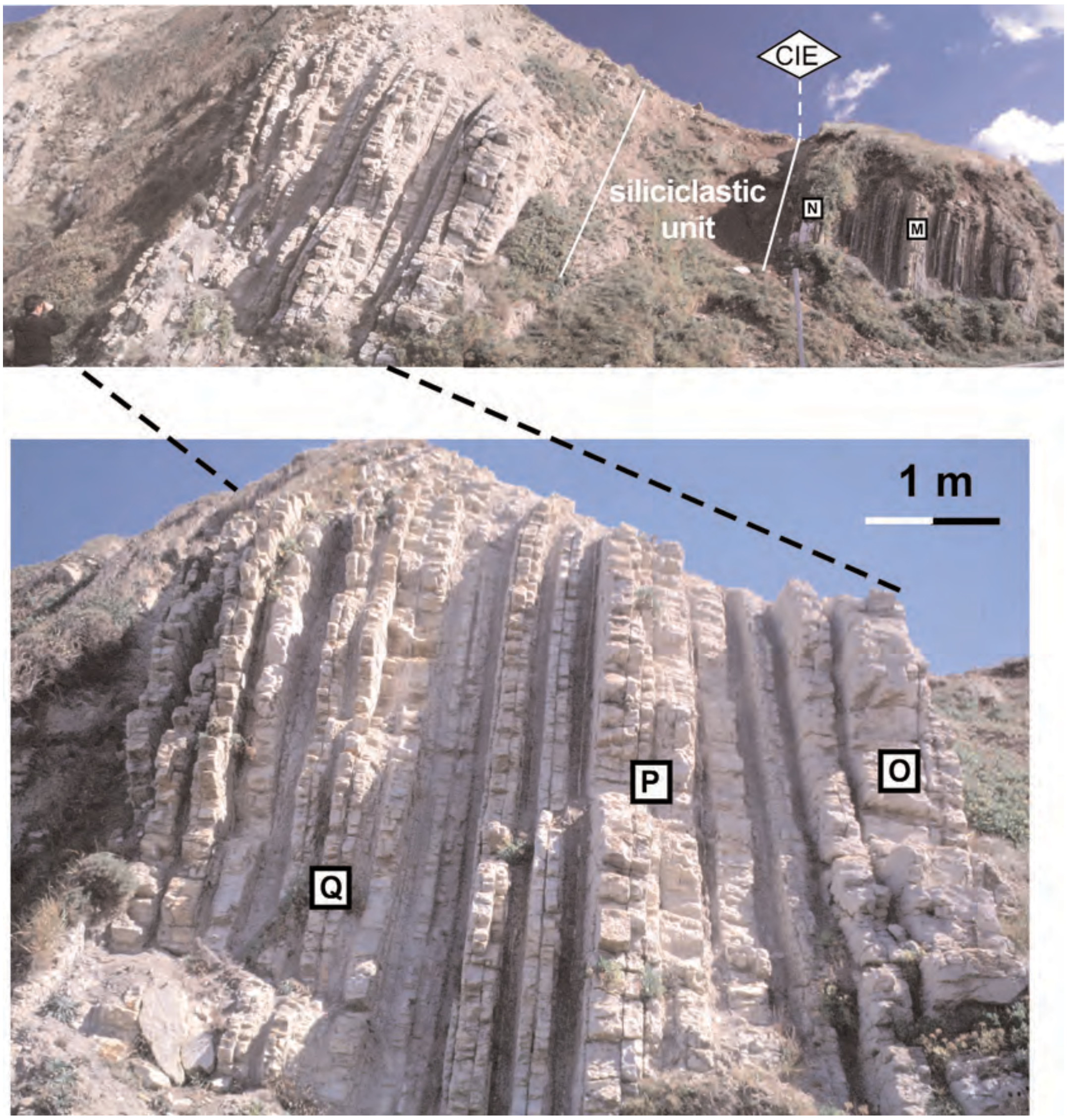


Fig. 34.- Outcrop photograph of the P/E boundary interval at Zumaia (a). The base of the Carbon Isotope Excursion (CIE) marking the P/E boundary is shown. The cyclic interval above the siliciclastic unit into the Eocene (b) shows clear bundling of 5-6 basic marl-limestone couplets indicating that correspond to precession (~21 kyr) and eccentricity (~110 kyr) cycles respectively. See lithologic column in Fig. 2 for exact location of the labelled reference beds.

5.3. Magnetostratigraphy of the P/E boundary interval (J. Dinarès-Turell, J.I. Baceta, V. Pujalte)

In order to define chron C25n along the late Paleocene and earliest Eocene time interval sixty-seven stratigraphic levels were sampled along 57 m of stratigraphic succession. Samples were mostly taken from the indurated limestone beds. Natural remanent magnetisation (NRM) and remanence through all demagnetisation stages were measured using a 2G-Enterprises high-resolution (45 mm diameter) pass-through cryogenic magnetometer equipped with DC-squids and operating in a shielded room at the Istituto Nazionale di Geofisica e Vulcanologia in Rome, Italy.

The intensity of the NRM was moderate to weak, generally between 1x10^{-4} A/m and 5x10^{-5} A/m. Two magnetisation components can be recognized in most samples upon demagnetisation, in addition to a viscous magnetisation removed below 5 mT or 100º-150°C. A low coercivity and low temperature component (L) is unblocked between 5 mT and 20-30mT (or 150°C and 300°C), and a high coercivity and tem-

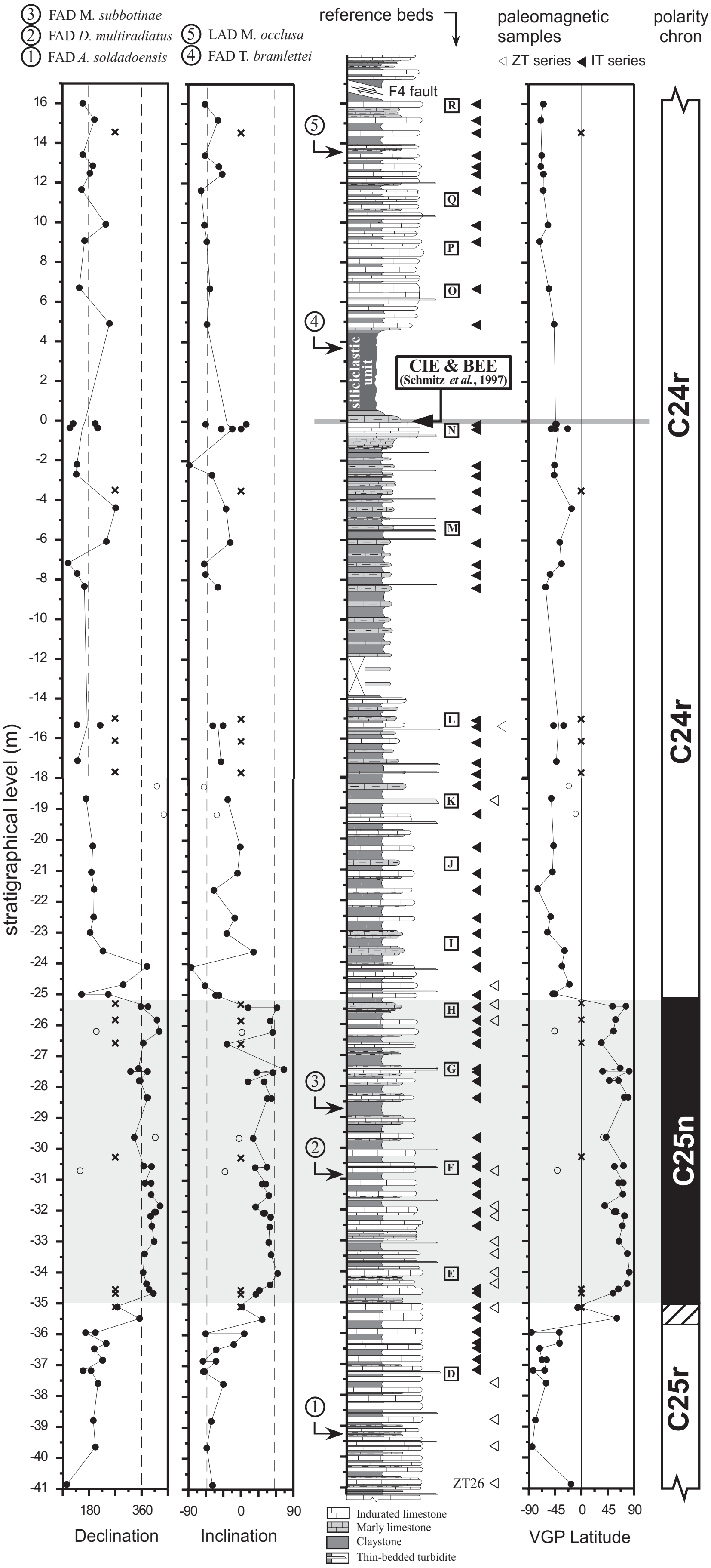


Fig. 35. Stratigraphic variations of declination, inclination of the ChRM vectors and virtual geomagnetic pole (VGP) latitude plotted on a lithologic log where key reference beds are given a letter. For a full detailed litholog covering the interval -53 m to +27 m see Baceta et al., (2000). The position of the ZT paleomagnetic series and the new IT series are indicated. Open circles denote unreliable data and crosses mark the position of samples that have provided no data. Uncertainty in the placement of the base of normal polarity chron C25n is shown by a hatched band.

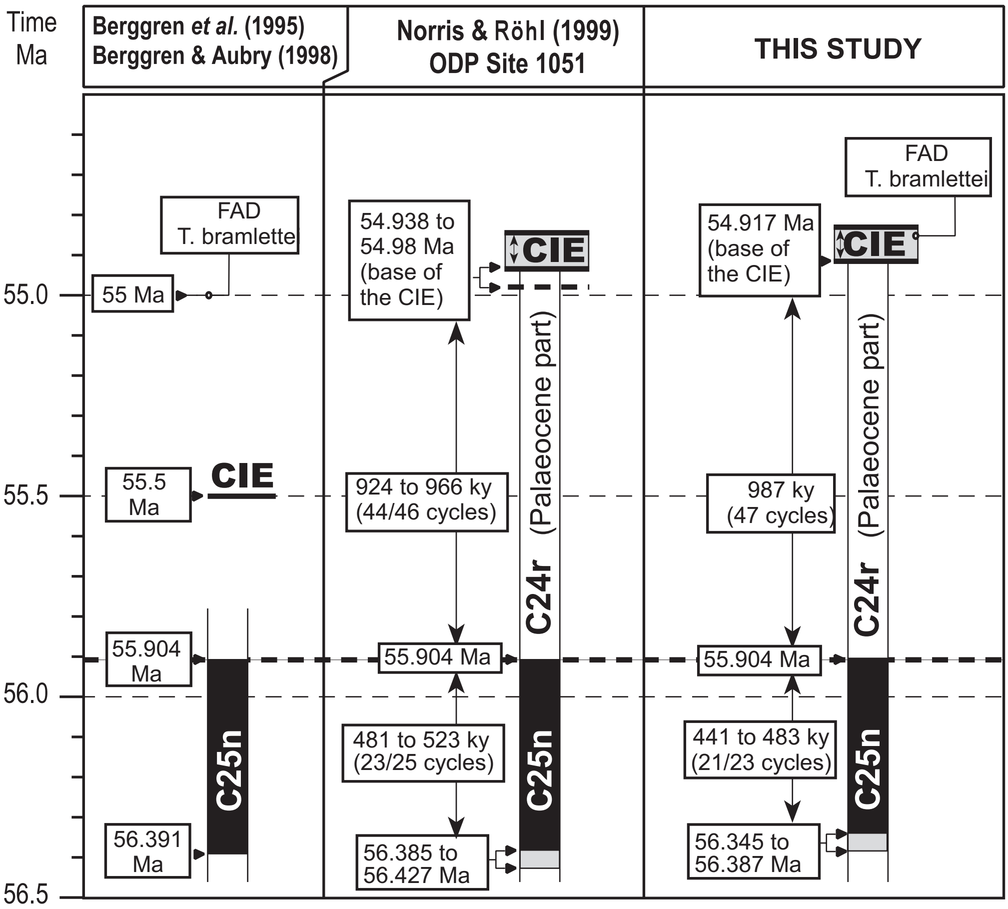


Fig. 36.- Different age estimates of relevant P/E boundary interval events. The top of chron C25n is used as fixed datum in the three cases.

perature component (H) is unblocked above 20-30 mT (or above 300°C) until fields around 100 mT (or about 550°C). The data is therefore compatible with magnetite as the main carrier of the remanence. Component L has a northerly declination and moderate to steep downward inclination (~60°) before rotation of bedding to the horizontal, and thus corresponds to the geocentric axial dipole direction for the site latitude. Consequently, component L is interpreted as a recent overprint. Component H trends toward the origin of the demagnetisation diagram and presents both normal and reverse polarities. No fold test is available due to similar bedding dip throughout the section (dip angle/dip direction, 64°/032°). Nevertheless, the following observations provide supporting evidence that the calculated ChRM directions are primary in age: (1) the presence of both normal and reverse ChRM directions, (2) the presence of a low coercivity/temperature present day field overprint in addition to the ChRM magnetization, (3) the mean inclination after bedding correction is near the expected Palaeocene inclination for the site location, (4) mean ChRM directions before bedding correction are not compatible with Palaeocene to recent geomagnetic field direction.

The declination, inclination and VGP latitude together with relevant biostratigraphic and lithologic information for the studied interval is shown in Fig. 35. Chron C25n has been delimited between -25.2 m to about -35.0 m although the C25n lower boundary reversal could be located to a slightly lower position (-37.7 m) if a NRM delay acquisition process is advocated (see Dinarès-Turell *et al.*, 2002 for further details).

5.4. Cyclostratigraphy of the P/E boundary interval *(J. Dinarès-Turell, J.I. Baceta, V. Pujalte)*

Cyclostratigraphy has proved to be a valuable metronome to parse time in the geologic record and has recently been applied to the P/E transition interval from Ocean Drilling Program (ODP) Sites 1051 and 690 (Norris and Röhl, 1999, Röhl *et al.*, 2000, Cramer, 2001) and in the Forada section (Agnini *et al.* 2005). The polarity record of ODP Hole 690B is hampered by a strong drilling-induced overprint (Ali *et al.,* 2000). However, apparent good magnetostratigraphic results from the upper part of chron C25r up to the lowermost Eocene chron C24r for ODP Site 1051 has enabled to propose a climatic orbital origin for the cycles observed in colour reflectance, magnetic susceptibility and Fe intensity counts (Norris and Röhl, 1999). They estimate a duration of 481 ky to 523 ky for polarity chron C25n, which is consistent with the 487 ky duration determined by assuming constant seafloor-spreading rates between radiometric tie points (Berggren *et al.*, 1995). Moreover, 924 ky to 966 ky is also estimated for the duration of the Palaeocene part of chron C24r (i.e., up to the onset of the $\delta^{13}C$ anomaly or CIE) in ODP Site 1051 (Norris and Röhl, 1999).

Leaving aside the thin-bedded turbidites, the succession studied in Zumaia is composed of cycles or lithology couplets formed by an indurated limestone part and a marly interval (Figs. 34 and 35). Chron C25n includes about 21-23 of these cycles and, hence, taking the duration constraints for this chron derived by Norris and Röhl, (1999), an average duration of 21 to 25 ky can be estimated for each cycle. This is consistent with the ~21 ky mean duration of the orbital precession cycle (Herbert *et al.,* 1995) given the uncertainties in identifying the cycles and determining the paleomagnetic boundaries both in ODP Site 1051 and at Zumaia. Similarly, the number of couplets for the Palaeocene part of chron C24r in the Zumaia section contains ~47 (assuming that 2 cycles actually occur in the not outcropping interval -12 to -14 m, Fig. 35). This implies a mean duration of about 20-21 ky for the couplets, again suggesting that they likely represent the expression of orbital precession. At present, we cannot directly ascertain whether the lithologic couplets in the lowermost Eocene section are also expressions of orbital forcing, but grouping of 4-5 individual couplets in larger bundles (Figs. 34 and 35) does suggests they may correspond to the precession and the ~100 ky eccentricity Milankovitch cycles respectively.

As pointed out by Norris and Röhl (1999), the temporal calibration of the Palaeocene part of Chron C24r at ODP Site 1051 is in conflict with previous estimates, because their cyclostratigraphic analysis brings the base of the CIE to an age of 54.93-54.98 Ma (taking 55.9 Ma for the top of C25n, Cande and Kent, 1995). A similar conclusion can be reached in the present study, since the duration of the Palaeocene part of C24r at Zumaia, as devised by cycle counting, places the P/E boundary ~0.987 My younger (i.e. at ~54.913 Ma) than the top of C25n (Fig. 36). These figures would imply that the base of the CIE (i.e., the P/E boundary) might be between 570 and 520 ky younger than its current estimation of 55.5 Ma (Berggren and Aubry, 1998). It is important to remark that the FAD of *Tribrachiatus bramlettei* (marker of the NP9/NP10 boundary) occurs in Zumaia and elsewhere almost immediately above the onset of the CIE (Monechi *et al.*, 2000a, b; Fig. 36). Therefore, if the absolute age of 55 Ma assigned by Berggren and Aubry to this event is correct, the base of the CIE should only be slightly older than this figure.

However, the absolute calibration of the P/E boundary is conditionally constrained on an $^{40}Ar/^{39}Ar$ age of 55 Ma for the –17 ash layer within the Mo clay sequence, Denmark (Cande and Kent, 1992). The –17 ash is also present in DSDP Hole 550, and recent reappraisal of the original paleomagnetic data from this Hole (Ali and Hailwood, 1998) indicates the position of the base of Chron C24n should be older (~0.5 My) and the original estimate was in error. If this could be substantiated by new data elsewhere, the P/E boundary and other biostratigraphic and magnetostratigraphic boundaries should be shifted to older ages accordingly. Clearly, a new anchor point with a well-constrained absolute age is needed and sections like Zumaia will be invaluable to reassess the chronology of the P/E boundary interval.

The section provides a complete and unambiguous record of chron C25n that suggests an orbital origin for the lithologic cycles observed throughout the P/E boundary interval. This is a major achievement in any marine succession embracing the P/E transition and provides the necessary constraints to further valuate the chemobiostratigraphical events spanning this interval.

5.5. Calcareous nannofossil turnover across the P/E boundary interval (*G. Bernaola, E. Angori, S. Monechi*)

Except for the first 1,3 m of the Siliciclastic Unit which are almost barren of calcareous nannofossils, all the studied samples of the Zumaia section are generally rich and well diversified. Preservation varies from moderate to poor being moderate in the Upper Paleocene and poor to moderate in the Eocene part.

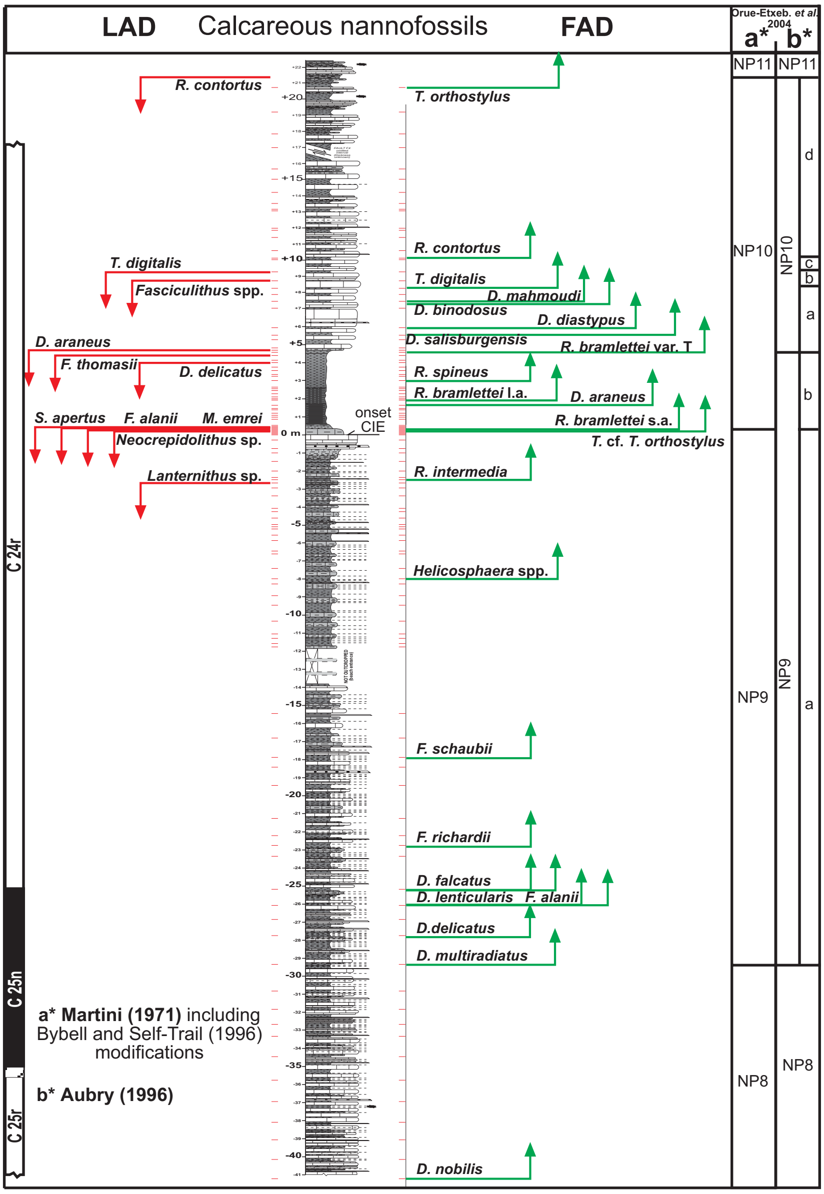


Fig. 37.- First (FO) and last (LO) occurrences of the most important calcareous nannofossils and their relative position within chrons C25r, C25n and C24r.

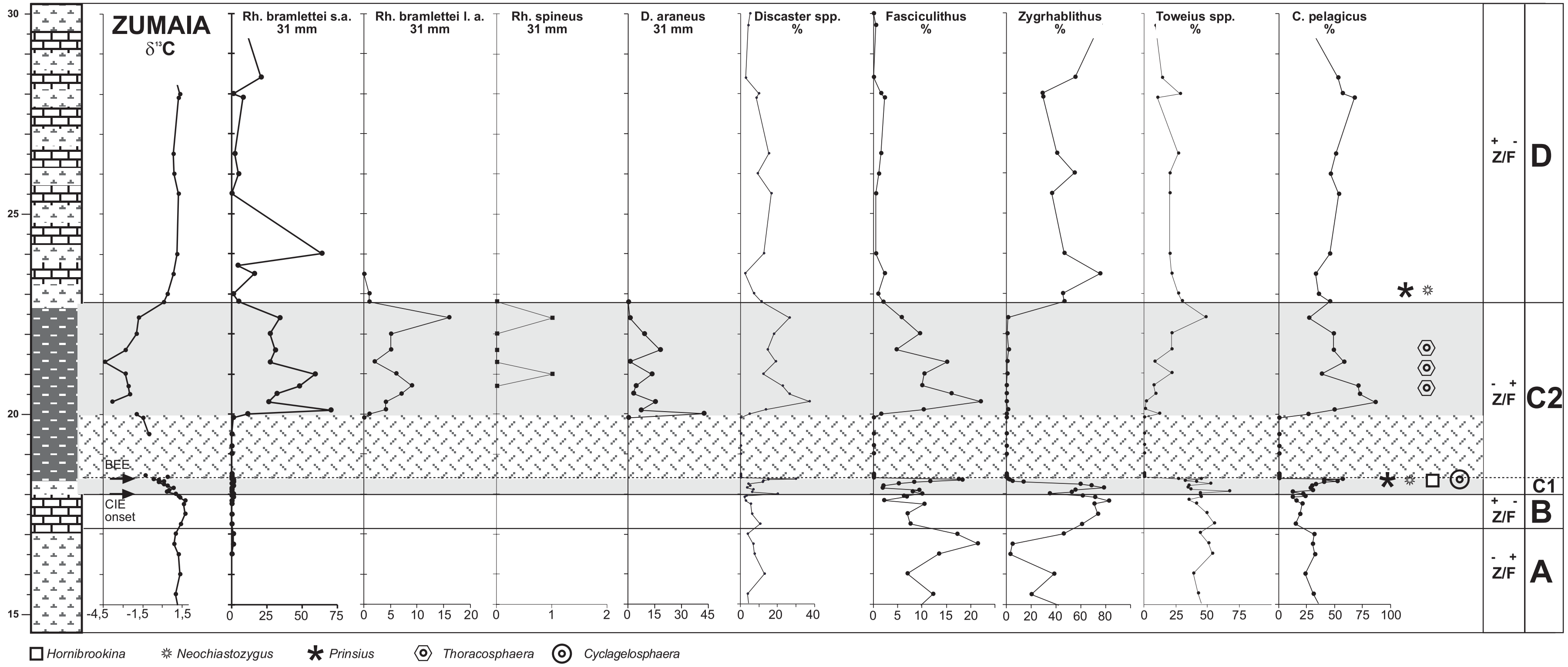


Fig. 38.- Relative and/or absolute abundance of selected calcareous nannofossil taxa across the Paleocene/Eocene transition in the Zumaia section. The relative abundance of *Zygrablithus*, *Discoaster* spp. and *Fasciculithus* spp. has been calculated from the non-(*Toweius pertusus* + *C. pelagicus*) assemblage.

We have observed significant changes in the occurrence and disappearance of several calcareous nannofossils taxa throughout the whole studied interval, however, the major turnover in the calcareous nannofossils has been recognized within the interval from about 5 m below to about 10 m above the P/E boundary, in which several genera become extinct or diversify (Fig. 37)

In the 40 cm thick marl bed capping the green limestone (Unit IV), just above the onset of the CIE and below the BEE, several first and last occurrences are registered, e.g. the LO of the last representatives of the genus *Neocrepidolitus*, a survivor of the mass extinction at the K/P boundary, as well as the LOs of *Munarrinus emrei*, *Fasciculithus alanii* and *Scapholithus apertus*. The disappearance of the latter two species has been observed shortly below the BEE in several other sections in the Tethys (Bybell & Self-Trail, 1995; Angori & Monechi, 1996; Monechi *et al.*, 2000a). In shallow marine deposits of the Basque Basin, the LO of *Sca. apertu*s has been used to approximate to the P/E boundary whenever the conventional markers are very rare or absent (Pujalte *et al.*, 2003). The FOs of the representatives of the *Rhomboaster-Tribrachiatus* lineage occur in the same thin marly interval.

> During the last 10 years, the morphological interpretation of the *Rhomboaster-Tribrachiatus* lineage has been cotroversial (Bybel & Self-Trail, 1995; Aubry *et al.*, 2000, von Salis *et al.*, 2000). This taxonomic problem is important, because Martini (1971) chose *Rhomboaster bramlettei* to mark the base of Zone NP10, an event usually used to approximate to the P/E boundary. In the present study and following the taxonomic remarks in Angori & Monechi (1996), we have differentiated 8 species of this lineage: 3 morphological variants of *Rh. bramlettei* (*Rh. bramlettei* "short arms", *Rh. bramlettei* "long arms", *Rh. bramlettei* var. T), *Rhomboaster spineus*, *Rhomboaster contortus*, *Tribrachiatus digitalis*; *Tribrachiatus* cf. *orthostylus* and *Tribrachiatus orthostylus*. We adopt the definition of Zone NP10 by Bybell & Self-Trail (1995, see also Angori & Monechi, in Molina *et al.*, 2000), according to which the lowermost occurrence of *Rhomboaster* (i.e. Angori & Monechi's *Rh. bramlettei* s.a.) defines the base of the Zone NP10.

The FO of *Rh. bramlettei* s.a. has been pinpointed 30 cm above the onset of the CIE and 10 cm below the BEE. This is somewhat lower than observed in other sections of the Basque Basin (Trabakua) and the Tethys area (Caravaca, Alamedilla) where the FO of this taxon occurs above the BEE. It has to be noted that only very rare specimens of *Rh. bramlettei* s.a. are found below the BEE at Zumaia which can only be detected by an exhaustive study over several tracks. Together with *Rh. bramlettei* s.a. a rare triradiate form has been identified. Those forms, called *T.* cf. *orthostylus* by Monechi *et al.* (2000a), have been observed in a similar stratigraphic position (some meters below the occurrence of *T. orthostylus* s.str.) in several other sections, e.g. Caravaca and Alamedilla (Angori & Monechi, 1996; Monechi *et al.*, 2000a).

Just above the 1.3 m non-calcareous interval of the lower part of the SU, we find the FO of *Discoaster araneus*, *Rh. bramlettei* l.a. and *Rh. spineus* in meters 1.65, 1.7 and 3 m respectively. Together with *Discoaster anartios*, which is absent in Zumaia, this species constitutes the *Rhomboaster-Discoaster araneus* association (Aubry, 2001), a short-lived species association characterized by nannoliths with anomalous structure. The stratigraphic distribution of the *Rhomboaster-Discoaster araneus* association at Zumaia, as in other sections of the Tethys and North Atlantic region, is restricted to the PETM interval.

The FO of the remaining species of the *Rhomboaster-Tribrachiatus* lineage species have been observed starting from the top of the SU (Fig. 37). *Rh. bramlettei* var T has its FO 4.65 m above the CIE onset, approximately at the level where the carbonate content begins to reach pre-P/E boundary values. *T. digitalis*, used to mark the base of the Subzone NP10b (Aubry, 1996) and proposed as a good datum to correlate deep marine successions with the base of the Ypresian stage (Aubry, 1996), has been detected only in two samples (in meter 8.25 and 9.25). In Zumaia, the occurrence of this taxon is very rare and discontinuous; it has not been found by Schmitz *et al.* (1997), whereas Monechi *et al.* (2000b) report it in only one sample. The FO of *Rh. contortus* was observed at meter 10.1, and its LO in meter 21.35, 50 cm above the FO of *T. orthostylus* at the top of the studied part of the succession (Fig. 37).

The quantitative analysis carried out in Zumaia allowed recognizing well defined different informal assemblage intervals that document the changes in the calcareous nannofossil during the PETM time interval.

Five assemblage (intervals) zones can be recognized: two below the onset of the CIE and three above it (Fig. 38).

The Paleocene assemblage (Zone A) is characterized by abundant *C. pelagicus*, *Tw. pertusus*, and the representatives of the genus, *Fasciculithus*, *Discoaster* and *Sphenolithus*, the former being quite rich in number of species. The calcareous nannofossil abundance and species richness is high (40 to 50 species and S-H higher than 2,1), indicative of general stable oceanic conditions with a stratiphied water column and low nutrient content.

The boundary between the Zone A and Zone B is marked in the studied sections by the beginning of a decreasing trend (drop) in *Fasciculithus* both in abundance and species richness and by a reverse increase in abundance of *Zygrhablithus*. Although the average diversity is still high (S-H values around 2), greater fluctuations can be observed in the number of species from sample to sample (from 36 to 51 species per sample). The calcareous nannofossil assemblage stability of zone A was broken; a prelude of the geochemical and oceanic changes at the PETM.

The base of the following assemblage Zone C coincides with the onset of the CIE and it is marked by an increase of *Discoaster* and *Fasciculithus* and by a sharply decrease in the abundance of *Zygrhablithus. Rhomboaster* is also recorded in connection with the onset of the CIE.

This zone includes the whole CIE and on the basis of the species recognized and of their fluctuations in abundanceit can be subdivided into two parts: a lower part Zone C-1 and an upper part Zone C-2. Zone C-1 (Figs. 38) corresponds approximately to the interval between the onset of the CIE and the BEE and it is discriminated at the base and top by two distinctive increases in abundance of *Discoaster*, mainly *D. multiradiatus* (rare *D. nobilis*, *D. delicatus* and *D. falcatus*), and of *Fasciculithus*, mainly *F. tympaniformis* (rare *F.* cf. *hayii* and *F. thomasii*). *Zygrhablithus* shows a significant abundance decrease and an inverse correlation with *Fasciculithus*. The calcareous nannofossil richness shows a slight decrease throughout this zone (Zone C-1) both in the number of species per sample (e.g. from 40 to 30 in Zumaia) and in the S-H index values (from 2 to 1,3). These changes in Zumaia occurs parallel to the abrupt turnover of calcareous nannofossils in high latitudes and can be explained as a sudden expansion of the low latitude warm water assemblages to high latitudes probably due to an abrupt increase of the surface water temperatures. The calcareous nannofossil assemblage fluctuations recorded in Zumaia could be the reflection of different pulses of that expansion, with the deterioration of the stable water column conditions of the upper Paleocene and the establishment of more unstable conditions.

The upper part of Zone C, Zone C-2, corresponds to the interval between the BEE (around the minimum carbon isotope values) up to the level where δ^{13}C returns to pre-excursion values. It is distinguished by abundant *Fasciculithus* (which shows a marked decreasing trend across this zone) and by the absence of *Zygrhablithus*. Zone C-2 shows lower calcareous nannofossil abundances and species diversity, but the most important feature that characterizes this interval is the presence of the so-called *D. araneus-Rhomboaster* assemblage, an event restricted to the PETM of the Tethys-Atlantic province (Kahn & Aubry, 2004).

In Zumaia the RD-assemblage appears a bit higher than the onset of the CIE, at the base of Zone C2, approximately at the level where the benthic foraminifera extinct around the minimum carbon isotope values, and disappear when the δ^{13}C return to pre-boundary values.

D. araneus is regularly present in the entire Zone but its maximum abundance is registered in the middle part of it (very low δ^{13}C values). As reported in Figure 38 *R. bramlettei* long arms has the same range as *D. araneu*s, whereas *D. spineus* only appears sporadically. *R. bramlettei* short arms, that has its FO before the other representatives of the RD-assemblage, shows its highest abundance in connection with δ^{13}C low values, and becomes less common with the extinction of the other representatives of the RD-assemblage. In fact, when δ^{13}C return to pre-excursion values *D. araneus*, *Rhomboaster spineus*, and *Rhomboaster bramlettei* long arms disappear and *R. bramlettei* short arms become less abundant. The origin and paleoecologic affinities of most of the representatives of the R-D assemblage is unknown, especially those of the *Rhomboaster-Tribrachiatus* lineage. According to the shape and morphological features of the some forms included in the R-D assemblage, we suggest that at least *D. araneus* , *D. anartios* and *Rh. bramlettei* long arms and *Rh. spineus*, as well as *Discoaster* with a big stem, should be interpreted as malformed forms produced by the increase in CO_2 concentration. The presence (abundance) of these taxa in different areas could indicate the proximity to the zones where the methane was released.

The following Zone D correlates with the return of the δ^{13}C to pre-excursion values and records important changes in calcareous nannofossil assemblages. *Zygrhablithus* reappears and increases considerable, becoming one of the most abundant taxa together with *C. pelagicus*, *Toweius* and minor *Sphenolithus*, whereas *Discoaster* and *Fasciculithus* experiment an important relative abundance decrease. The RD-assemblage is no longer present and only some representative of *R. bramlettei* short arms can be found.

A relative abundance increase of *Neochiastozygus*, *Hornibrookina* and *Braarudosphaera bigelowii* can also be found. Together with all these changes and in parallel to the δ^{13}C increase there is a relatively fast raise in the calcareous nannofossil abundance and species richness (S-H 1,75 and 25-32 species per sample). These changes are indicative of an important turnover in the ocean water column structure.

5.6. Planktic foraminifera across the P/E boundary interval (X. Orue-Etxebarria, E. Apellaniz, F. Caballero)

Planktic foraminifer assemblages are rich; the number of specimens per gram of dried residue ranges from 19.000 to 55.000, occasionally reaching up to 80.000 specimens in samples from a few meters below the SU. In contrast, from most samples of the SU less than 200 specimens per gram were obtained. The proportion of planktic foraminifers versus total foraminifers is usually larger than 90% and in many occasions exceeds 95%. Preservation of tests is generally good; some details of the ornamentation are normally less clearly developed than in tropical specimens and may occasionally ble-blurred due to secondary incrustations and/or recrystallization. A total of 37 species of planktic foraminifera have been recognized. The stratigraphic distribution of these species is shown in figure 39.

Planktic foraminifera of the Zumaia section had already been examined by several previous authors (Hillebrandt, 1965; Canudo & Molina, 1992; Arenillas & Molina, 1995; Canudo *et al.*, 1995; Pardo *et al.*, 1996; Schmitz *et al.*, 1997; Arenillas & Molina, 2000 and Orue-Etxebarria *et al.*, 2001).

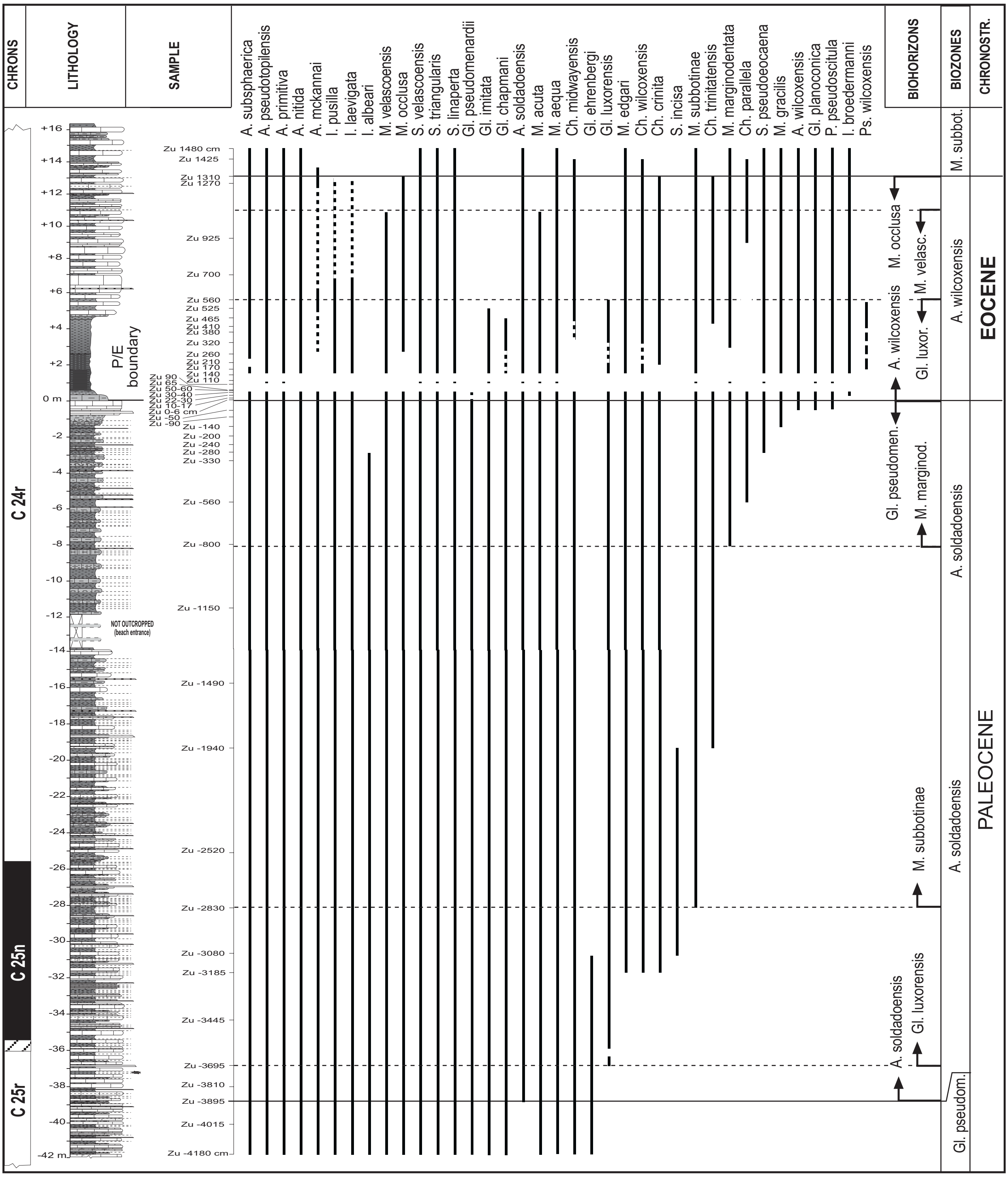


Fig. 39. Planktonic foraminiferal species ranges across the Paleocene/Eocene transition in the Zumaia section. Main biohorizons.

Most of our results are in general agreement with these earlier studies, although a detailed comparison is hampered by differences in their respective logs of the section. Since a detailed discussion of the distribution of the planktic foraminifers in the Zumaia section would be somewhat redundant, we focus only on those results which are at odds with previous studies, either at Zumaia or elsewhere.

1.- Within Biozone P4, *Globanomalina pseudomenardii* and the first, small size, specimens of

Morozovella subbotinae (Fig. 39) coexist; a fact that has been pointed out before in other sections situated westward (Orue-Etxebarria *et al.*, 1992; Orue-Etxebarria *et al.*, 1996).

2. - In the Zumaia section, *Gl. pseudomenardii* ranges up to the P/E boundary (Fig. 39). This extended range is in contrast to the indications of other authors who place the last occurrence (LO) of this species at the boundary between zones P4 and P5 (Canudo y Molina, 1992; Canudo *et al.*, 1995; Lu *et al.*, 1998; Molina *et al.*, 1999; Berggren *et al.*, 2000; Arenillas & Molina, 2000). According to Blow (1979) this species extends until the *Globorotalia* (*Acarinina*) *wilcoxensis berggreni* Biozone of the basal Lower Eocene. From its first occurrence (FO) up to 11,50 m below the P/E boundary, the specimens of *Gl. pseudomenardii* are rather frequent and reach a maximum diameter of 0,45 mm. However, above this level they become smaller and with a less distinct keel. Within 10 m below the CIE, they are very rare and reache a maximum size of 0,25 mm. This reduction in size and frequency does not coincide with any facies change. The persistent presence of *Gl. pseudomenardii* in all samples of Zone P4 indicates that they are unlikely to correspond to reworked tests.

3.- The first specimens of *Gobanomalina luxorensis* occur in the lower part of the *Acarinina soldadoensis* Biozone, (Fig. 39). However they are very rare and become more common only from sample Zu -280 cm onward. This is in contrast to the observations of Speijer & Samir (1997) in the eastern Mediterranean Region that *Gl. luxorensis* is very rare below the BEE (< 0,1%), but becomes very frequent (up to 10% of total planktic foraminiferal assemblage) above this level where it is the dominant species within the *Globanomalina* lineage. In Zumaia, it always constitutes < 1% of the total planktic foraminiferal assemblages and the most representative of the genus *Globanomalina* in the topmost Paleocene is *Gl. imitata*.

4.- Nine out of 12 samples examined from the 4 m thick Siliciclastic Unit (Unit V), above the P/E boundary have provided planktic foraminifers, whereas other authors reported them from only one sample (Canudo & Molina, 1992; Canudo *et al.*, 1995; Schmitz *et al.*, 1997; Molina *et al.*, 1999 and Arenillas & Molina, 2000). Even after a very careful examination of all samples above the CIE onset, we have not found any specimens of *Acarinina sibaiyaensis*, *Morozovella allisonensis* and *M. africana*, species restricted to the PETM which were described by Kelly *et al.* (1996) and found by Lu *et al.* (1998) throughout the Tethys realm. These species have neither been found in other sections of the Pyrenees (Orue-Etxebarria *et al.*, 1996; Orue-Etxebarria *et al.*, 2001 and Pujalte *et al.*, 2003). They may be characteristic of tropical latitudes (Kelly *et al.*, 1996), or that *M. allisonensis* and *A. sibaiyaensis* are ecophenotypical variants (Kelly *et al.* 1998). Pardo *et al.* (1999) *A. sibaiyaensis* and *A. africana* neither in the Zumaia section, nor in Site 401 of the Bay of Biscay.

5.- *Pseudohastigerina wilcoxensis* is very rare in the Zumaia section and occurs very sporadically at the lower part of the SU, approximately 1 meter above the onset of the CIE (Fig. 39).

6.- In Zumaia, we could not observe a significant increase in the proportion of acarininids in connection with the CIE (Fig. 40). Such an increase has been reported by Kelly *et al.* (1996, 1998) from tropical regions, by Lu *et al.* (1998) and Pardo *et al.* (1999) from sections of the Tethys realm.

Representatives of the genera *Subbotina, Morozovella* and *Acarinina* dominate the planktic foraminiferal assemblages below the P/E boundary (Fig. 40). In contrast to Schmitz *et al.* (1997), Molina *et al.* (1999) and Arenillas & Molina (2000) we find a significant increase in the proportion of acarininids, from 15-20% to 30-40%, only from sample Zu 465 cm, i.e. well above the CIE onset. However, most samples from the SU contain too few specimens (< 200 per gram) of planktic foraminifera to allow significant analyses of the proportions of the planktic foraminiferal genera.

7.- In the planktic foraminiferal assemblages of the Zumaia section, no drastic changes at species level occur in conjunction with the P/E boundary. The same species dominate below and above it. Above the SU, LOs of species such as *Globanomalina chapmani, Gl. imitata, Igorina pusilla and*

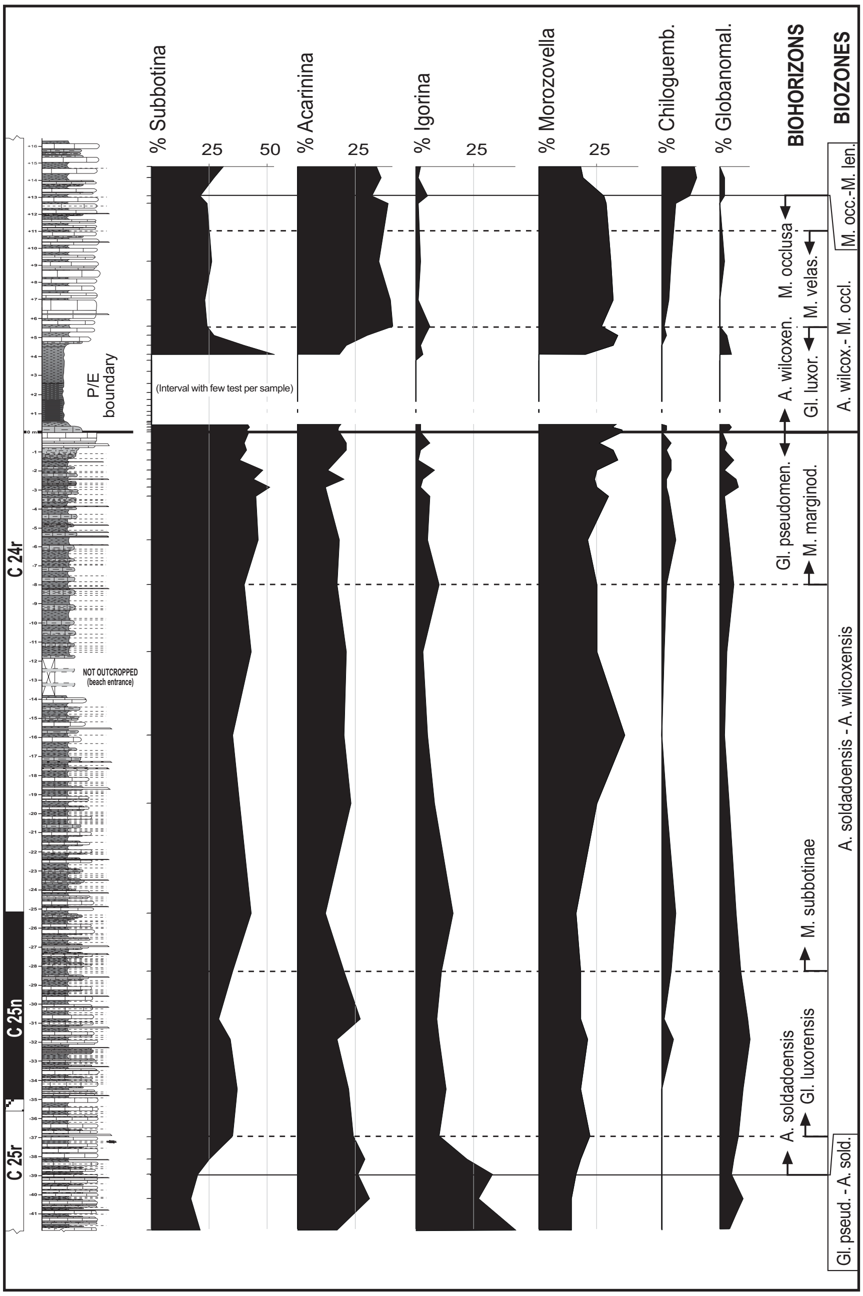
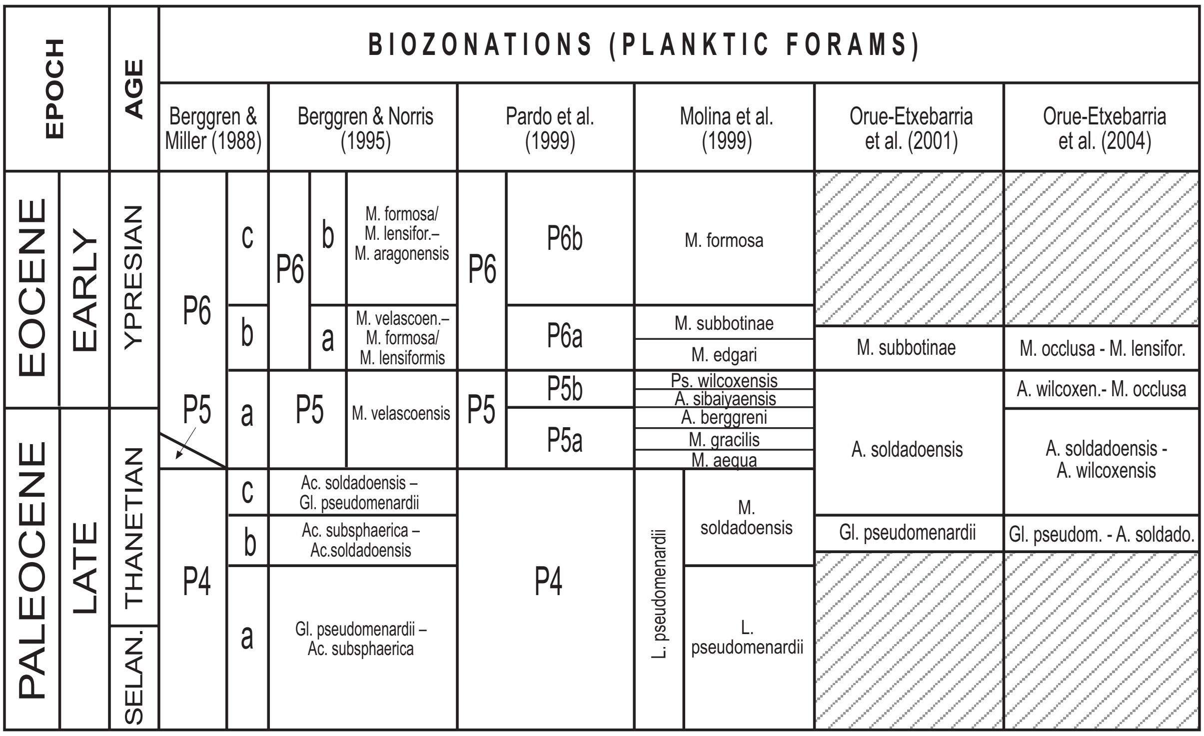


Fig. 40.- Relative abundance of planktic foraminiferal genera across the Paleocene/Eocene transition in the Zumaia section. Note the significant increment of *Acarinina*'s proportions take place around 6,5 m above the P/E boundary.

I. laevigata are gradual (Fig. 39). The occurrence of the latter species in samples above the P/E boundary casts some doubts on the use of their LOs as markers for the P/E boundary, which according to Arenillas & Molina (1996) coincide with the FOs of *A. sibaiyaensis* and *A. africana.*

8.- The planktic foraminiferal biozonation of the interval around the P/E boundary has experienced several modifications since Berggren *et al.* (1995). The LO of *Parasubbotina variospira* instead of the LO of *Acarinina subsphaerica* is used by Berggren *et al.* (2000) to draw the boundary between subzones P4a and P4b of the *Globanomalina pseudomenardii* Zone (P4).

The *Morozovella velascoensis* Zone (P5) has been subdivided into two subzones making use of the LO of *Igorina laevigata* (Arenillas & Molina, 1996). Another subdivision of Zone P5 uses the FOs of *Acarinina sibaiyensis* and/or of *A. africana* which is thought to coincide concide with the LO of



EPOCH		AGE	BIOZONATIONS (PLANKTIC FORAMS)										
			Berggren & Miller (1988)		Berggren & Norris (1995)			Pardo et al. (1999)		Molina et al. (1999)		Orue-Etxebarria et al. (2001)	Orue-Etxebarria et al. (2004)
EOCENE	EARLY	YPRESIAN	P6	c	P6	b	M. formosa/ M. lensifor.– M. aragonensis	P6	P6b		M. formosa	▨	▨
				b		a	M. velascoen.– M. formosa/ M. lensiformis		P6a		M. subbotinae		
											M. edgari	M. subbotinae	M. occlusa - M. lensifor.
			P5	a	P5		M. velascoensis	P5	P5b		Ps. wilcoxensis	A. soldadoensis	A. wilcoxen.- M. occlusa
											A. sibaiyaensis		
PALEOCENE	LATE	THANETIAN							P5a		A. berggreni		A. soldadoensis - A. wilcoxensis
											M. gracilis		
											M. aequa		
			P4	c	P4	c	Ac. soldadoensis – Gl. pseudomenardii	P4		L. pseudomenardii	M. soldadoensis		
				b		b	Ac. subsphaerica – Ac.soldadoensis					Gl. pseudomenardii	Gl. pseudom. - A. soldado.
				a		a	Gl. pseudomenardii – Ac. subsphaerica				L. pseudomenardii	▨	▨
		SELAN.											

Fig. 41.- Planktic foraminiferal zonation across the Paleocene/Eocene transition in the Zumaia section and comparison with other autors.

Igorina laevigata and the BEE (Lu *et al.*, 1998; Pardo *et al.*, 1999). Subsequently, Molina *et al.* (1999) differentiated 5 subzones within Zone P5 using the FOs of *Morozovella gracilis, Acarinina berggreni, Acarinina sibaiyaensis* and *Pseudohastigerina wilcoxensis.*

In this study, we have identified 4 interval biozones, marked by the FOs of *Acarinina soldadoensis* and *A. wilcoxensis* as well as the LO of *Morozovella occlusa* (Fig. 41):

a) Globanomalina pseudomenardii-Acarinina soldadoensis Biozone: Interval from the FO of *Globanomalina pseudomenardii* to the FO of *Acarinina soldadoensis.* In this study, we have only analyzed the uppermost part of this biozone, which corresponds to the final part of P4b of Berggren *et al.* (1995) (Fig. 41).

b) Acarinina soldadoensis - Acarinina wilcoxensis Biozone: Biostratigraphic interval from the FO of *A. soldadoensis* to the FO of *A. wilcoxensis*. This biozone includes Subzone P4c and part of Zone P5 of Berggren *et al.* (1995). The FO of *A.wilcoxensis* corresponds to the P/E boundary (we use a wide definition of *A.wilcoxensis* including in it *A. wilcoxensis berggreni*).

c) Acarinina wilcoxensis-Morozovella occlusa Biozone: Interval from the FO of *A. wilcoxensis* to the LO of *M. occlusa.* This biozone corresponds to the upper part of Zone P5 of Berggren *et al.* (1995). In the Zumaia section, its lower part is within the SU. The FOs of *Gl. planoconica* and *Igorina broedermanni* coincide more or less with the FO of *A. wilcoxensis* s.l. followed by the FO of *Pseudohastigerina wilcoxensis* which is a rare and sporadic form. *Gl. chapmani*, *Gl. imitata* and *Gl. luxorensis* disappear above the SU. *Igorina pusilla, I. laevigata, Morozovella velascoensis* and *M. acuta* and finally *M. occlusa*, the most abundant representative of the genus *Morozovella* in the Basque Basin have their LOs in the upper part of the biozone.

d) Morozovella occlusa-Morozovella lensiformis Biozone: Interval between the LO of *M. occlusa* and the FO of *M. lensiformis*. In this study we have only analyzed the lowest part of the biozone which corresponds to the lower part of P6a of Berggren *et al.* (1995).

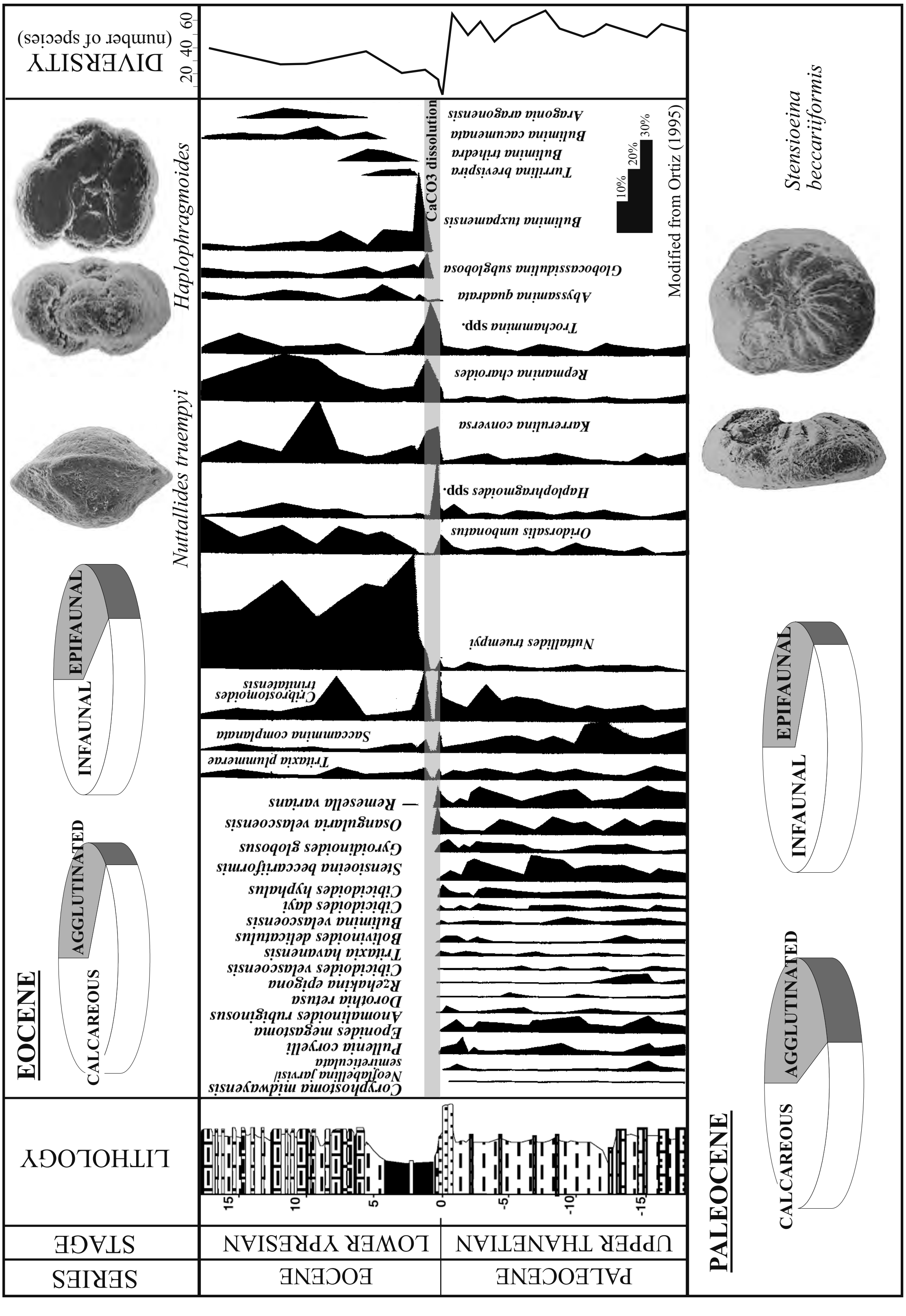


Fig. 42.- Benthic foraminiferal assemblagles across the Paleocene/Eocene transition in the Zumaia section

5.7. Benthic foraminifera from the Paleocene/Eocene transition *(Laia Alegret and Silvia Ortiz)*

Benthic foraminiferal assemblages from the Paleocene-Eocene transition at Zumaia have been studied by Ortiz (1995) and Kuhnt and Kaminski (1997), among others. Assemblages contain abundant representatives of the Velasco-type fauna and from the agglutinated flysch-type fauna, and they indicate a middle to lower bathyal depth of deposition (Fig. 43).

Upper Paleocene benthic foraminiferal assemblages consist of highly diversified communities that contain 30-50% of agglutinated taxa. A wide range of morphotypes is recorded in the upper Paleocene, including biconvex trochospiral (e.g., *Anomalinoides rubiginosus*, *Stensioeina beccariiformis*), planoconvex trochospiral (*Cibicidoides hyphalus, C. velascoensis*), rounded trochospiral (e.g., *Cribrostomoides trinitatensis, Gyroidinoides globosus*), rounded planispiral (e.g., *Pullenia coryelli*), cylindrical tapered (e.g., *Bulimina velascoensis, Dotothia retusa, Clavulinoides amorpha*), or flattened tapered morphotypes (e.g., *Hormosina velascoensis*), among others. Assemblages contain both infaunal and epifaunal morphogroups, and they are dominated by oligotrophic, deep-water taxa that are common under well-oxygenated conditions.

A global extinction event of deep-sea benthic foraminifera (BEE) has been identified in coincidence with the P/E boundary not only at Zumaia, but also at many other sites and sections worldwide (Tjalsma and Lohmann, 1983; Thomas, 1998, 2003). The P/E boundary at Zumaia is marked by a dramatic decrease in diversity of the benthic foraminiferal assemblages, and the extinction of ~30% of the species. The extinction of *Stensioeina beccariiformis* is significant, because the extinction of its shallow-water type (*Angulogavelinella avnimelechi*) allows correlation to the extinction event at shallower settings.

A sharp increase in the relative abundance of small, thin-walled agglutinated taxa such as *Haplophragmoides* spp., *Repmanina charoides* or *Trochammina* spp. (up to 80% of the assemblages), is also recorded in a dissolution interval located in the lowermost Eocene. Such an increase in the percentage of agglutinated foraminifera has been observed to occur at the beginning of the CIE-interval not only in Zumaia and in other bathyal sections from the Basque Basin (Orue-Etxebarria *et al.*, 1996), but also in many sections across the world (e.g., Thomas, 1998), including the stratotype section at Dababiya (Egypt; Alegret *et al.*, 2005).

Above the dissolution horizon, the calcareous oligrotrophic species *Nuttallides truempyi* and *Oridorsalis umbonatus* dominate the assemblages, and agglutinated species such as *Karrerulina conversa* or *Repmanina charoides* are also common (Ortiz, 1995; Kuhnt and Kaminski, 1997). Diversity of the Eocene assemblages is lower than in the Paleocene, suggesting that benthic foraminiferal assemblages from the lowermost 15 m of the Eocene had not completely recovered from the Paleocene/Eocene event. Further studies are needed to describe in more detail the recovery of the benthic foraminiferal assemblages after the P/E boundary, and to look into the causes of the dissolution interval in the lowermost Eocene at Zumaia.

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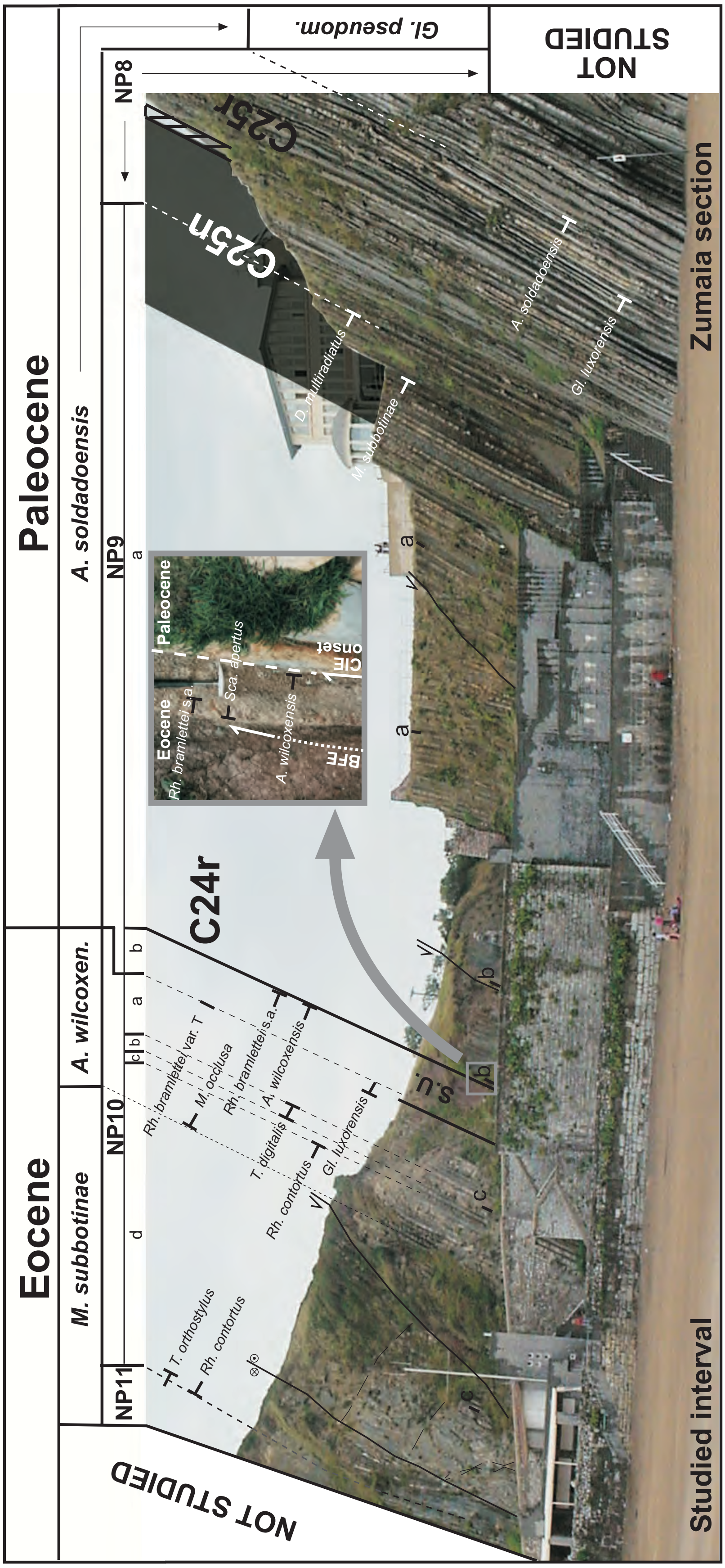

Fig. 43.- Zumaia P/E boundary transition picture showing the precise location of the principal tectonic structures and magnestostratographic and calcareous plankton events.

REFERENCES

Adatte, T., Bolle, M.P., de Kaenel, E., Gawenda, P, Winkler, W. and von Salis, K. (2000): Climatic evolution from Paleocene to earliest Eocene inferred from clay-minerals: a transect from northern Spain (Zumaya) to southern (Spain, Tunisia) and southeastern Tethys margins (Israel, Negev). GFF, 122: 7-8.

Agnini C., Fornaciari E., Giusberti L., Backman, J., Capraro, L., Grandesso P., Luciani V., Muttoni, G., Rio D. and Tateo F., (2005): The early paleogene of the Valbelluna. Early cenozoic extreme climates (ODP Leg 208 Post Cruise Meeting) Field trip Guidebook.

Alegret, L., Ortiz, S., Arenillas, I. and Molina, E., (2005): Paleoenvironmental turnover across the Paleocene/Eocene boundary at the Stratotype section in Dababiya (Egypt) based on benthic foraminfiera. Terra Nova 17, 526-536.

Ali, J.R. and Hailwood, E.A., (1998): Magnetostratigraphic (re) calibration of the Paleocene/Eocene boundary interval in boles 550 and 549, Goban Spur, eastern North Atlantic. Earth Planet. Sci. Lett., 161, 201-213.

Ali, J.R., Kent, D.V. and Hailwood, E.A., (2000): Magnetostratigraphic reinvestiga-tion of the Palaeocene/Eocene boundary interval in Hole 690B, Maud Rise, Antárctica. Geophys. J. Int., 141, 639-646.

Alvarez, L.W., Alvarez, W., Asaro, F. and Michel, H.V. (1980): Extraterrestrial cause for the Cretaceous-Tertiary extinction. Science 208, 1095-1108.

Alvarez, W., Alvarez, L.W., Asaro, F. and Michel, H.V. (1982): Current status of the impact theory for the terminal Cretaceous extinction. Geological Society of America Special Paper, 190, 305-315.

Alvarez, W., Alvarez, L.W., Asaro, F. and Michel, H.V. (1984): The end of the Cretaceous: Sharp boundary or gradual transition. Science 223, 1183-1186.

Angori, E. and Monechi, S. (1996): High-resolution nannofossil biostratigraphy across the Paleocene-Eocene boundary at Caravaca (southern Spain). Israel Journal of Earth Sciences., 44: 197-206.

Apellaniz, E. (1998): Los foraminíferos planctónicos en el tránsito Cretácico-Terciario: análisis de cuatro secciones en depósitos de cuenca profunda de la Región Vasco-Cantábrica. Tesis Doctoral, Univ. País Vasco, 390 p.

Apellaniz, E. (1999): Los foraminíferos planctónicos en el tránsito Cretacico-Terciario. Análisis de 4 secciones en depósitos de cuenca profunda en la Región Vasco-Cantábrica. Tesis doctoral. UPV/EHU, 1-391.

Apellaniz, E., Lamolda, M.A. y Orue-Etxebarria, X. (1983): Posición estratigráfica de las "Calizas del Danés", País Vasco. Revista española de Micropaleontología, 15, 447-455

Apellaniz, E., Baceta, J.I., Bernaola, G., Núñez-Betelu, K., Orue-Etxebarria, X., Payros, A., Pujalte, V., Robin, E. and Rocchia, R. (1997): Analysis of uppermost Cretaceous-lowermost Tertiary hemipelagic successions in the Basque country (Western Pyrenees): evidence of a sudden extinction of more than half planktic foraminifer species at the K/T boundary. Bull. Soc. Geol. France, 168: 783-793.

Apellaniz, E., Orue-Etxebarria, X. and Luterbacher, H. (2002): Evolution of the Early Paleocene planktonic foraminifera: A Basque point of view. N.JB. Geol. Paläont. Abh, 225(2), 157-194

Arenillas, I. (1996): Los foraminíferos planctónicos del Paleoceno-Eoceno inferior: sistemática, bioestratigrafía, cronoestratigrafía y paleoceanografía. Tesis Doctoral, Univ. Zaragoza, 513 p.

Arenillas, I. y Molina, E. (1995): Análisis cuantitativo de los foraminíferos planctónicos del Paleoceno de Zumaya: implicaciones paleoambientales y eventos paleoceanográficos. - Geogaceta, 17: 23-26.

Arenillas, I. y Molina, E. (1996): Bioestratigrafía y evolución de las asociaciones de foraminíferos planctónicos del tránsito Paleoceno-Eoceno en Alamedilla (Cordilleras Béticas). Revista Española de Micropaleontología. 18, 75-96.

Arenillas, I., Arz, J.A. y Molina, E. (1998): El límite Cretácico-Terciario en Zumaia, Osinaga y Musquiz (Pirineos): control bioestratigráfico y cuantitativo de hiatos con foraminíferos planctónicos. Rev. Soc. Geol España, 11: 127-138.

Arenillas, I. y Molina, E. (2000): Reconstrucción paleoambiental con foraminíferos planctónicos y Cronoestratigrafía del tránsito Paleoceno-Eoceno de Zumaya (Guipúzcoa). Revista Española de Micropaleontología. 32(3): 283-300.

Aubry, M. P. (1994): The Thanetian Stage in NW Europe and its significance in terms og global events. GFF 116, 43-44.

Aubry, M.-P. (1996): Towards an Upper Paleocene-Lower Eocene High Resolution Stratigraphy Based on Calcarous Nannofossil Stratigraphy. Israel. J. of Earth Sc. 44, 239- 253.

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: 158-203; New York (Columbia University Press).

Aubry, M.-P. (2000): Where should the Global Stratotype Section and Point (GSSP) for the Paleocene/Eocene boundary be located? Bul!. Soc. Géol. 171,461-66.

Aubry, M.P. (2001): Provincialism in the photic zone during the LPTM. - In: Ash, A. M. & Wing, S. L. (eds): Climate and Biota of the Early Paleogene, Internacional meeting, Wyoming, USA, 6p.

Aubry, M.P. and Berggren, W.A. (2000): The Homeless GSSP: The dilemma of the Paleocene/Eocene boundary. Tertiary Research. 20, 107-112

Aubry, M.P, Requirand, C. and Cook, J. (2000): The Rhomboaster-Tribrachiatus lineage: A remarkable succession of events from 55.5 to 53.2 Ma. - GFF, 122: 15-18.

Baceta, J.I. (1996): El Maastrichtiense superior, Paleoceno e Ilerdiense inferior de la Región Vasco-Cantábrica: Secuencias Deposicionales, Facies y Evolución Paleogeográfica. Tesis Doctoral, Univ. País Vasco, 372 p.

Baceta, J.I., Pujalte, V. y Orue-Etxebarria, X. (1991): Rasgos erosivos y deposicionales en la zona proximal de un "slope-apron" carbonatado del Paleoceno Inferior (Eibar, Cuenca Vasca). Geogaceta, 10: 76-78.

Baceta, J.I., Pujalte, V., Dinarès-Turell, J., Payros, A., Orue-Etxebarria, X. and Bernaola, G. (2000): The Paleocene-Eocene boundary interval in the Zumaia section (Gipuzkoa, Basque basin): magnetostratigraphy and high-resolution lithostratigraphy. Rev. Soc. Geol. España, 13: 375-391.

Baceta, J. I.; Wright, V. P. and Pujalte, V. (2001): Palaeo-mixing zone karst features from Palaeocene carbonates of north Spain: criteria for recognizing a potentially widespread but rarely documented diagenetic system. Sedimentary Geology: 139, 205-216.

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

Baceta, J.I., Pujalte, V. and Bernaola, G. (2005): Paleocene coralgal reefs of the western Pyrenean Basin, N Spain: evidence concerning the post-Cretaceous recovery of reefal ecosystems. Palaeogeogr., Palaeoclimatol., Palaeoecol , 224: 117-143.

Barnolas, A., Robador, A., Serra-Kiel, J. and Caus, E. (1990): Introduction to the early Paleogene of the south Pyrenean basin (IGCP 286, Early Paleogene benthos, field-trip). ITGE, Madrid, 159 p.

Beerling, D. J. and Jolley, D. W. (1998): Fossil plants record and atmospheric $^{12}CO_2$ and temperature spike across the Palaeocene-Eocene transition in NW Europe. — Journal of the Geological Society, London. 155, 591-594.

Berggren, W.A., Kent, D.V., Swisher, C.C.III and Aubry, M.P. (1995): A revised Cenozoic geochronology and chronostratigraphy. In: Geochronology, time scales and global stratigraphic correlations (Berggren, W.A., Kent, D.V., Aubry, M.P. and Hardenbol, J., Eds.). S.E.P.M. spec. publ., 54: 129-212.

Berggren, W.A. and Aubry, M.-P., (1998). The Paleocene/Eocene Epoch/Scries Bounday: Chronostratigraphic Frame-work and Estimated Geochronology. In: Late Palaeocene-Early Eocene Climatic and Biolic Events in the Marine and Terrestrial Records (M.-P. Aubry *et al.*, eds), pp. 18-36. Columbia University Press, New York.

Berggren, W.A., Lucas, S. and Aubry, M.-P. (1998). Late Paleocene-Early Eocene Climatic and Biotic Evolution: An overview. In: Late Palaeocene-Early Eocene Climatic and Biotic Events in the Marine and Terre.itrial Records (M.-P. Aubry et al., eds), pp. 15-17. Columbia University Press, New York.

Berggren, W.A., and Miller, K.G., 1988. Paleogene tropical planktonic foraminiferal biostratigraphy and magnetobiochronology. Micropaleontology, 34:362-380.

Berggren, W. A., Aubry, M.P., Van Fossen, M., Kent, D. V, Norris, R. D. & Quillévéré, F. (2000): Integrated Paleocene calcareous plankton magnetobio-chronology and stable isotope stratigraphy: DSDP Site 384 (NW Atlantic Ocean) - Palaeogeography, Palaeoclimatology, Palaeoecology, 159: 1-51.

Berggren, W. A. and Ouda, K. H. (2003): Upper Paleocene - lower Eocene planktonic foraminiferal biostratigraphy of the Dababiya section, Upper Nile Valley (Egypt). Micropaleontology, Supplement. VO- 49; NO- 1; PG- 61-92

Berggren, W. A. and Pearson, P. N. (2005): A Revised Tropical To Subtropical Paleogene Planktonic Foraminiferal Zonation. Journal of Foraminiferal Research.2005; 35: 279-298.

Bernaola, G. (2002): Los nannofósiles calcáreos del Paleoceno en el dominio Pirenaico. Bioestratigrafía, cronoestratigrafía y paleoecología. Tesis Doctoral, Univ. País Vasco, 445 p.

Betzler, C. (1989): The upper Paleocene to middle Eocene between the rio Segre and the rio Llobregat (eastern south Pyrenees). Facies, stratigraphy and evolution. Tübinger Geowiss. Arb., Reihe A, 2: 133 p.

Blow, W.H., (1979): 'Cainozoic Globigerinida'. E.J. Brill, 3 Volumes, Leiden, 1413pp. Atlas.

Bown, P.R. (2005). Selective calcareous nannoplankton survivorship at the Cretaceous-Tertiary boundary. Geology.

Bralower, T. J. (2002): Evidence of surface water oligotrophy during the Paleo-cene-Eocene termal máximum: Nannofossil assemblage data from Ocean Dril-ling Program Site 690, Maud Rise, Weddell Sea. - Paleoceanography, 17(2): 1023.

Bralower, T. J., Zachos, J. C., Thomas, E., Parrow, M., Paull, C. K., Kelly, D. C., Premoly Silva, I., Sliter, W. V. and Lohmann, K. C. (1995): Late Paleocene to Eocene paleoceanography of the ecuatorial Pacific Ocean: Stables isotopes recorded at ODP Site 865, Allison Guyot. — Paleoceanography, 10, 841-865.

Bybell, L. M. and Self-Trail, J. M. (1995): Evolutionary, biostratigraphic, and ta-xonomic study of calcareous nannofossils from a continuous Paleocene/Eocene boundary section in New Jersey. U.S. Geol. Surv., Prof. Paper 1554, 36 pp.

Cande, S.C. and Kent, D.V. (1995): Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic, J. Geophys. Res. 100. 6093-6095.

Canudo, J.I. and Molina, E. (1992): Planktic foraminiferal faunal turnover and bio-chronostratigraphy of the Paleocene-Eocene boundary at Zumaya (Northern Spain). Revista de la Sociedad Geológica de España, 5, 145-157.

Canudo, J.I., Keller, G., Molina, E. and Ortiz, N. (1995): Planktic foraminiferal turnover and d^{13}C isotopes across the Paleocene-Eocene transition at Caravaca and Zumaya, Spain. Palaeogeogr. Palaeoclimatol. Palaeoecol., 114: 75-100.

Clemens, W.A. Archivald, J.D. and Hickey, L. J. (1981): Out with a whimper not a bang. Paleobiology 7, 293-298.

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

Courtillot, V., Féraud, G., Maluski, H., Vandamme, D., Moreau, M.G., and Besse, J. (1988): Deccan flood basalts and the Cretaceous/Tertiary boundary: Nature, v. 333, p. 843-846.

Cowie, J.W., Zieger, W., and Remane, J. (1989): Stratigraphic Commission accelerates progress, 1984-1989. Episodes, 112: 79-83.

Cramer, B.S. (2001): Latest Palaeocene-earliest Eocene cyclostratigraphy: using core photographs for reconnaissance geophysical logging. Earth. Planet. Sci. Leu., 186, 231-244.

Cremades, J. (1982): Contribución al conocimiento de los foraminíferos planctónicos y al estudio bioestratigráfico del Eoceno y Oligoceno del sector oriental de las Cordilleras Béticas. Tesis Doctoral. Universidad de Granada. 350 p.

Crouch, E.M., Heilmann-Clausen, C., Brinkhuis, H., Morgans, H.E.G., Rogers, K.M., Egger, H., and Schmitz, B. (2001): Global dinoflagellate event associated with the Late Paleocene Thermal Maximum. Geology, 29:315–318.

Cuevas, J.L. (1992): Estratigrafía del "Garumniense" de la Conca de Tremp, Prepirineo de Lérida. Acta Geol. Hisp., 27: 95-108.

Delacote, O., Renard, M., Laj, C., Perch-Nielsen, K., Premoli Silva, I. and Clauser, S. (1985): Magnétostratigraphie et biostratigraphie du passage Crétacé-Tertiaire de la coupe de Bidart (Pyrenees Atlantiques). - Bull. Soc. géol. Fr., (8), I, 3, 243-254.

Dinarès-Turell, J., Baceta, J.I., Pujalte, V., Orue-Etxebarria, X and Bernaola, G. (2002): Magnetostratigraphic and cyclostratigraphic calibration of a prospective Paleocene-Eocene stratotype at Zumaia (Basque basin, Northern Spain). Terra Nova, 34: 279-290.

Dinarès-Turell, J., Baceta, J.I., Pujalte, V., Orue-Etxebarria, X., Bernaola, G. and Lorito, S. (2003): Untangling the Palaeocene climatic rhythm: an astronomically calibrated Early Palaeocene magnetostratigraphy and biostratigraphy at Zumaia (Basque basin, northern Spain). Earth and Planetary Science Letters. 216, 483-500.

Eichenseer, H. (1988): Facies geology of late Maastrichtian to early Eocene coastal and shallow marine sediments, Tremp-Graus basin, northeastern Spain. Tesis Doctoral, Univ. Tübingen (Publicada en Arb. Inst. u. Mus. Geol. Pal. Universität Tübingen, 1, 273 p.

Farley, K. A., and S. F. Eltgroth (2003): An alternative age model for the Paleocene-Eocene thermal maximum using extraterrestrial 3He, Earth Planet. Sci. Lett., 208, 135– 148.

Garrido-Megías, A. y Ríos, L.M. (1972): Síntesis geológica del Secundario y Terciario entre los ríos Cinca y Segre (Pirineo Central de la vertiente surpirenaica, provincias de Huesca y Lérida). Bol. Geol. Min., 83: 1-47.

Gawenda, P., Winkler, W., Schmitz, B. and Adate, T (1999): Climate and bio-productivity control of carbonate turbidite sedimentation (Paleocene to earliest Eocene, Gulf of Biskay, Zumaia, Spain). - Journal of Sedimentary Research, 69(6): 1253-1261.

Gómez de Llarena, J. (1946): Revisión de algunos datos paleontológicos del Flysch cretáceo y nummulítico de España. Not. Com. Inst. Geol. Min. España, 15: 109-1665.

Gorostidi, A. (1993): Nanofósiles calcáreos y eventos del Cretácico Medio-Superior de la Cuenca Vasco-Cantábrica. Tesis Doctoral. Universidad del País Vasco / Euskal Herriko Unibertsitatea 331pp.

Gorostidi, A y Lamolda, M.A. (1991): El paso Cenomaniense-Turoniense de Menoyo (Alava): Variaciones de la nanoflora calcárea. Geogaceta, 10, 54-57.

Hallam, A. (1987): End-Cretaceous mass extinction event: Argument for terrestrial causation. Science 238, 1237-1242.

Herbert, T.D., Premoli-Silva, I., Erba, E. and Fischer, A.G. (1995): Orbital chro-nology of Cretaceous/Palaeoccne marine sediments. In: Geochronology Time. Scales and Global Stratigraphic Córrela tion. (W.A. Berggren *et al.*, eds). Spec. Publ. Soc. econ. Paleont. Miner., Tulsa, 54, 81-92.

Hillebrandt, A., von (1965): Foraminiferen - stratigraphie in alttertiár von Zumaya (provinz Guipúzcoa, NW Spanien) und ein Vergleich mit anderen Tethys-Gebieten. Bayr. Akad. Wiss. Math-Natur. Kl. Abh. NF, 123: 1-62.

Jenkins, D. G. and Lutherbacher, H. (1992): Paleogene stages and their boundaries. Neues Jahrbuch für Geologie und Paläontologie, Abh. 186, 1-5.Kahn, A. and Aubry, M. P. (2004): Provincialism associated with the Paleocene/Eocene thermal maximum: temporal Constraint. Marine Micropaleontology, 52, 117-131.

Kaminski, M.A. Gradstein, F.M., *et al.*, (2005): Atlas of Paleogene Cosmopolitan Deep-Water Agglutinated Foraminifera.Grzybowski Foundation Special Publication 10, 547pp.

Kapellos, CH. (1974): Ueber das Nannoplankton im Alttertiär des Profils von Zumaya-Guetaria (Provinz Guipuzcoa, Nordspanien). Eclogae geol. Helv. 67(2):435-444.

Keller, G. (1988): Extinction, survivorship and evolution of planktic foraminifers across the Cretaceous/Tertiary boundary at El Kef, Tunisia. Marine Micropaleontology, 13: 239-263.

Keller, G. and Lindinger, M., (1989): Stable isotope, TOC, and $CaCO_3$ record across the Cretaceous Tertiary boundary at El Kef, Tunisia. Paleogeogr. Paleoclimat, Paleoecol., 73: 243-265.

Keller, G., Adatte, Th., Stinnesbeck, W., Stüben, D., Berner, Z., Kramar, U. and Harting, M. (2004): More evidence that the Chicxulub impact predates the K/T mass extinction.Meteorites & Planetary Science 39 (7), 1127-1144

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

Kelly, D.C., Bralower, T.J. and Zachos, J.C. (1998): Evolutionary consequences of the latest Paleocene thermal máximum for tropical planktonic foraminifera. -Palaeogeography, Palaeoclimatology, Palaeoecology, 159: 1-51.

Kelly, D.C., Zachos, J.C., Bralower, T.J., and Schellenberg, S.A. (2005): Enhanced terrestrial weathering/runoff and surface-ocean carbonate production during the recovery stages of the Paleocene-Eocene Thermal Maximum: Paleoceanography, v. 20, PA4023, doi: 10.1029/2005PA001163.

Kennett, J.P. and Stott, L.D. (1991): Abrupt deep-sea warming, paleoceanographic changes and benthic extinctions. Nature, 353, 225-229.

Khan, A. and Aubry, M. P. (2004): Provincialism in the calcareous nannoplankton during the Paleocene-Eocene thermal maximum: constrain on timing and duration. Marine Micropalentology, 52: 117-131.

Knox, R. (1994): From regional stage to standard stage: implications for the historical Paleogene stratotypes of NW Europe. GFF 116, 56-57.

Knox, R. (1998): Kaolinite influx within Paleocene/ Eocene boundary strata of western Europe. Newsl. Stratigr., 36: 49-53.

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

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

Kuhnt, W. and Kaminski, M.A. (1993): Changes in the comunity structure of deep water agglutinated foraminifers across the K/T boundary in the Basque Basin (Northern Spain). Revista Española de Micropaleontología, 25, 57-92.

Kuhnt, W. and Kaminski, M. A., (1997): Cenomanian to lower Eocene Deep-water agglutinated foraminifera from the Zumaya section, Northern Spain. Annales Societatis Geologorum Poloniae 67, 257-270.

Kuhnt, W., Kaminski, M.A. and Moullade, M. (1989): Late Cretaceous deep-water agglutinated benthic foraminiferal assemblages of the upper Cretaceous North Atlantic and its marginal seas. Geologische Rundschau 78: 1121-1140 (DSDP Special Volume).

Lamolda, M.A., Orue-Etxebarria, X. and Proto-Decima, F. (1983). - The Cretaceous-Tertiary boundary in Sopelana (Biscay, Basque Country). - Zitteliana, 10, 663-670.

Lamolda, M.A., Mathey, B. and Wiedmann, J. (1988): Field-guide excursion to the Cretaceous-Tertiary boundary section at Zumaya (Northern Spain). En: M.A. Lamolda, E.G. Kauffman And O.H. Walliser (eds.): Paleontology and Evolution: Extinction Events. 2nd International Conference on Global Bioevents, Bilbao, Revista de Paleontología, nº extraordinario, p. 141-155.

Lamolda, M.A. and Gorostidi, A. (1992): Nannofossil stratigraphic record in upper Maastrichtian-Lowermost Danian at Zumaya. Mem. Ist. Geol. Miner. Univ. Padova, 43: 149-161.

Laskar, J. (1993): Orbital, precessional, and insoltaion quantities for the Earth from 320 Myr to +10 Myr, Astron. Astrophys. 270 522-533.

Laskar, J., Robutel, P., Joutel, F., Gastineau, M., Correia, A.C.M. and Levrard, B: (2004): A long term numerical solution for the insolation quantities of the Earth. Astronomical Solutions for Earth Paleoclimates. 428, 261-285.

Leroux, H., Rocchia, R., Froget, L., Orue-Etxebarria, X., Doukhan, J-C., and Robin, E. (1995): The K/T boundary of Beloc (Haiti) : compared stratigraphical distributions of the boundary markers. -Earth Planet. Sci. Lett., 131, 255-268.

Leymerie, A. (1863): Note sur la systeme Garumnien. Bull. Soc. Géol. France, 20: 483-488.

Lu, G., Keller, G. and Pardo, A. (1998): Stability change in Tethyan planktic foraminifera across the Paleocene-Eocene transition. - Marine Micropaleontology, 35: 203-233.

Luterbacher, H.P., Eichenseer, H., Betzler, C. and Van Den Hurk, A.M. (1991): Carbonate-siliciclastic depositional systems in the Paleogene of the South Pyrenean foreland basin: a sequence stratigraphic approach. In: Int. Ass. Sediment. Spec. Publ., 12: 351-407.

Luterbacher, H.P., Hardenbol, J. and Schmitz, B. (2000): Decisión of the voting members of the Internacional Subcommission on Paleogene stratigraphy on the criterion for the recognition of the Paleocene-Eocene boundary. - Newsletter of the Internacional Subcommission on Paleogene Stratigraphy, 9: 13.

Lyell, C. (1833): Principies of Geology, being an attempt to explain the forme r changes of the Earth's surface by reference to causes now in operation (vol. 3). John Murray, London: 398 p.

Mangin, J.Ph. (1959-1960): Le nummulitique sud-Pyrénéen a l'ouest de l'Aragon. Pirineos, 51-58: 1-659.

Martínez, F.J. (1977): Estudio micropaleontológico del Nummulítico de un sector comprendido entre Moreda-Piñar- Pedro Martinez (zona subbetica). Tesis Doctoral, Univ. Granada, Vol. I, 219 p., Vol.262 p.

Martini, E. (1971): Standard Tertiary and Quaternary calcareous nannoplankton zonation, In: Proceedings of the Second Planktonic Conference Roma 1970, (ed. A. Farinacci), Edizioni Tecnoscienza, Roma, 2, 739-785.

Mathey, B. (1982): El Cretácico superior del Arco Vasco. In: El Cretácico de España. Universidad Complutense de Madrid, 111-135.

Mathey, B. (1986). - Les flyschs Crétacé supérieur des Pyrénées Basques. Age, anatomie, origine du material, milieu de dépót et relations avec l'ouverture du golfe de Gascogne. - Université de Dijon, 403 p. (unpublished Ph. D. Thesis).

McLean, D.M. (1982): Deccan volcanism and the Cretaceous-Tertiary transition scenario: a unifying causal mechanism. Syllogeus 39, 143-144.

McLean, D.M. (1985): Deccan Traps mantle degassing in the terminal Cretaceous marine extintions. Cretaceous Research 6, 235-259.

Medus, J. y Colombo, F. (1991): Sucession climatique et limite stratigrafique Crétacé-Tertiaire dans le N.E. de l'Espagne. Acta Geol. Hisp., 26: 27-35.

Molina, E. (1994): Paleocene sections in Spain: chronostratigraphical problems and possibilities. GFF, Geologiska Föreningens i Stockholm Förhandlingar. 116, 58-60.

Molina, E. (1996): El límite Paleoceno/Eoceno en España: características y posibilidades. Geogaceta, 20(7): 1668-1671.

Molina, E., Arenillas, I. and Pardo, A. (1999). High resolution planktic foraminiferal biostratigraphy and correlation across the Paleocene/Eocene boundary in the Tethys. Bulletin de la Société géologique de France, 170, 521-530.

Molina, E., Angori, E., Arenillas, I., Monechi, S. and Schmitz, B. (2000): In-tegrated stratigraphy across the Paleocene/Eocene boundary at Campo (Spain). -GFF, 122 (1): 106-107.

Monechi, S., Angori, E. and Speijer, R. P. (2000b): Upper Paleocene biostratigraphy in the Mediterranean región: Zonal markers, diachronism, and preservational problems. - GFF, 122: 108-110.

Monechi, S., Angori, E. and Von Salis, K. (2000a): Calcareous nannofossil turn-over around the Paleocene/Eocene transition at Alamedilla (southern Spain). -Bulletin de la Société Géologique de France, 171(4): 477-489

Norris, R. D. (1997): symbiosis as an evolutionary innovation in the radiation of Paleocene planktic foraminifera. Paleobiology, Vol. 22, p. 461-480.

Norris, R.D. and Róhl, U. (1999): Carbón cycling and chronology of chínate warming during the Palaeocene/Eocene transition. Nature, 401, 775-778.

Officer, Ch.B. and Drake, Ch.L. (1983): The Cretaceous-Tertiary transition. Science 219, 1383-1390.

Officer, Ch.B. and Drake, Ch.L. (1985): Terminal Cretaceous enviromental events. Science 227, 1161-1167.

Officer, Ch.B., Hallam, A. Drake, C.L. and Devine, J.D. (1987): Late Cretaceous and paroxysmal Cretaceous/Tertiary extinction. Nature 326, 143-149.

Olsson, R.K., Hemleben, C., Berggren, W.A., and Huber, B.T. (1999): Atlas of Paleocene planktonic foraminifera: Smithsonian Contributions to Paleobiology, v. 85, p. 252.

Ortiz, N. (1995): Differential patterns of benthic foraminiferal extinctions near the Paleocene/Eocene boundary in the North Atlantic and the western Tethys. Marine Micropaleont., 136: 17-30.

Orue-Etxebarria, X. (1982). - Los foraminíferos planctónicos del Paleó-geno del Sinclinorio de Bizkaia (Corte de Sopelana-Punta de la Galea). - Universidad del País Vasco, 271 p. (Unpublished Ph. D. Thesis).

Orue-Etxebarria, X. (1983): Los foraminíferos planctónicos del Paleógeno del Sinclinorio de Bizkaia (corte Sopelana-Punta Galea): Parte 1, Kobie, 13: 175-249.

Orue-Etxebarria, X. (1997). - El Kef blind test IV results. - Mar. Mi-cropaleontol., 29, 2, 85-88.

Orue-Etxebarria, X., Baceta, J. I., Pujalte, V., Núñez-Betelu, K., Serra-Kiel, J., Apellaniz, E. and Payros, A. (1992): Evaluating prospective Paleo-cene-Eocene boundary parastratotypes in the deep water Basque Basin, western Pyrenees: Ermua and Trabakua Pass sections. Biscay, Basque Basin - Zaragoza IGCP Project 308: Palaeocene/Eocene Boundary Events. International Meeting and Field Conference, Abstract Volume, p. 9 and 10 page addendum.

Orue-Etxebarria, X., Rocía, R., Pujalte, V., Apellaniz, E. and Baceta, J. I. (1991): Estudio preliminar del tránsito Cretácico/Terciario en la sección del monte Urko (Eibar, País Vasco). - / Congreso del Grupo Español del Terciario, Vic, 240-243.

Orue-Etxebarria, X., Apellaniz, E., Baceta, J. I., Coccioni, R., Di Leo, R., Dinarès-Turell, J., Galeotti, S., Monechi, S., Núñez-Betelu, K., Pares, J. M., Payros, A., Pujalte , V., Samso, J.-M., Serra-Kiel, J., Schmitz, B. and Tosquella, J. (1996): Physical and bioestratigraphic analysis of two prospective Paleocene-Eocene Boundary Stratotypes in the intermediate-deep water Basque Basin, western Pyrenees: The Trabakua Pass and Ermua sections. - N. Jb. Geol. Paláont. Abh., 201(2): 179- 242.

Pardo, A., Arenillas, I. and Canudo, J.I. (1996): Planktic Foraminifera and Paleo-geographic evolution across the Paleocene-Eocene boundary at Site 401, Zumaya and Alamedilla sections. - International Paleogene Stage Boundaries Meeting, Zaragoza (Spain), June 24-29 (abstr).

Orue-Etxebarria, X., Pujalte, V., Bernaola, G., Apellaniz, E., Baceta, J.I., Payros, A., Núñez-Betelu, K., Serra-Kiel, J. and Tosquella. J. (2001): Did the Late Paleocene Thermal Maximum affect the evolution of larger foraminifers?: Evidences from calcareous plankton of the Campo section (Pyrenees, Spain). — Marine Micropaleontology, 41, 45-71.

Pardo, A., Keller, G. and Oberhansli, H. (1999): Paleoecologic and paleoceano-graphic evolution of the Tethyan realm during the Paleocene-Eocene transition. -Journal of Foraminiferal Research, 29 (1): 37-57.

Orue-Etxebarria, X., Bernaola, G., Baceta, J. I., Angori, E., Caballero, F., Monechi, S., Pujalte, V. Dinarés-Turell, J., Apellaniz, E. and Payros, A., (2004): New constraints on the evolution of planktic foraminifers and calcareous nannofossils across the Paleocene-Eocene boundary interval: the Zumaia section revisited. N. Jb. Geol. Paläont. Abh., 234(1-3): 223-259.

Perch-Nielsen, K. (1979a): Calcareous nannofossils from the Cretaceous between the North Sea and the Mediterranean. IUGS Series A, 6, 223-72.

Perch-Nielsen, K. (1979b): Calcareous nannofossil zonation at the Cretaceous/Tertiary boundary in Denmark. In: T. Birkelund, R.G. Bromley (eds.), Proceedings Cretaceous-Tertiary Boundary Events symposium, Copenhagen, 2, 120-126.

Perch-Nielsen, K. (1981a): Les nannofossiles calcaires á la limite Crétacé-Tertiaire près de El Kef, Tunisie. Cahiers de Micropaleontologie 3, 25-37.

Perch-Nielsen, K. (1981b): Nouvelles observations sur les nannofossils calcaires a la limite Crétacé-Tertiaire prés de El Kef (Tunisie). Cahiers de Micropaléontologie 3, 25-36.

Perch-Nielsen, K. (1985a): Mesozoic calcareous nannofossils, In: H.M. Bolli, J.B. Saunders, K. Perch-Nielsen (Eds.), Plankton Stratigraphy, Cambridge University Press, 329-426.

Perch-Nielsen, K. (1985b): Cenozoic calcareous nannofossils In: H.M. Bolli, J.B. Saunders, K. Perch-Nielsen (Eds.), Plankton Stratigraphy, Cambridge University Press, 427-554.

Percival S.F. Jr. and Fischer A.G. (1977): - Changes in calcareous nan-noplankton in the Cretaceous-Tertiary biotic crisis at Zumaya, Spain. - Evol. Theory, 2, 1-35.

Plaziat, J.C. (1975): Signification paléogéografique dex "clacaires conglomérés des brèches et des niveaux a rhodophycées" dans la sédimentation carbonatée du bassin Basco-Béarnais à la base du Tertiaire (Espagne-France). Rev. Géogr. Phys. Geol. Dyn., 2-17: 239-258.

Plaziat, J.C. (1981): Late Cretaceous to late Eocene paleogeographic evolution of southwest Europe. Palaeogeogr. Palaeoclimatol. Palaeoecol., 36: 263-320.

Pluchery, E. (1995): Cycles de dépôts du continent àl'océan. Les séries d'âge Maastrichtien supérieur à Eocène moyen de la marge basco-cantabrique et de son arrière-pays ibérique (Espagne du Nord). Tesis Doctoral, Univ. Bourgogne-Dijon, 324 p.

Pospichal, J. (1996b): Calcareous nannoplankton mass extinction at the Cretaceous/Tertiary boundary: an update. In: G. Ryder, D. Fastovsky, S. Gartner (Eds.), The Cretaceous-Tertiary event and other catastrophes in Earth history., Geological Society of America, Special Pape, 307, 335-360.

Preissinger, A., Aslanian, S., Brandstatter, F., Grass, F., Stradner, H. and Summesberger, H. (2002): Cretaceous-Tertiary profile, rhythmic deposition, and geomagnetic polarity reversals Pap. 356. 213-229.

Prestwich, J. (1850a):. On the structure of the strata between the London Clay and the Chalk in the London and Hampshire Tertiary systems. Part I. Quaterly Journal of the Geological Society of London, 6, 252-281.

Prestwich, J. (1850b): On the structure of the strata between the London Clay and the Chalk in the London and Hampshire Tertiary systems. Part II. Quaterly Journal of the Geological Society of London, 8, 235-264.

Prestwich, J. (1854): On the structure of the strata between the London Clay and the Chalk in the London and Hampshire Tertiary systems. Part III. Quaterly Journal of the Geological Society of London, 10, 75-154.

Puigdefàbregas, C. and Souquet, P. (1986): Tecto-sedimentary cycles and depositional sequences of the Mesozoic and Tertiary from the Pyrenees. Tectonophysics, 129: 173-203.

Pujalte, V.; Robles, S.; Robador, A.; Baceta, J. I. and Orúe-Etxebarría, X. (1993): Shelf-to-basin Palaeocene palaeogeography and depositional sequences, western Pyrenees, north Spain. In: *Sequence Stratigraphy and Facies Associations* (Posamentier, H. W., Summerhayes, C. P., Haq, B. U. and Allen, G. P., Eds.). *Spec. Publ. Int. Ass. Sediment.*, 18, 369-395.

Pujalte, V., Baceta, J.I., Dinarès-Turell, J., Orue-Etxebarria, X., Parés, J.M. and Payros, A. (1995): Biostratigraphic and magnetostratigraphic intercalibration of the latest Cretaceous and Paleocene depositional sequences from the deep-water Basque Basin, western Pyrenees, Spain. Earth Planet. Sci. Letters, 136: 17-30.

Pujalte, V., Baceta, J.I., Apellaniz, E., Orue-Etxebarria, X., Payros, A. and Serra-Kiel, J. (1998a): Paleocene of the Pyrenees and the North Sea Basin: differences and similarities. Newsletter on Stratigraphy, 36: 43-47.

Pujalte, V., Baceta, J.I., Orue-Etxebarria, X. and Payros, A. (1998b): Paleocene strata of the Basque Country, western Pyrenees, northern Spain: Facies and sequence development in a deep-water starved basin. En: Mesozoic and Cenozoic Sequence Stratigraphy of European Basins (P.C. de Graciansky, J. Hardenbol, T. Jacquin y P.R. Vail, Eds.), Soc. Econ. Paleont. Miner., Sp. Publ., 60: 311-325

Pujalte, V., Baceta, J.I., and Payros, A. (2002): Tertiary: Western Pyrenees and Basque-Cantabrian region. En: The Geology of Spain (W. Gibbons y T. Moreno, Eds.), Geol. Soc. (London), 293-301.

Pujalte, V., Orue-Etxebarria, X., Schmintz, , B., Tosquela, , J., Baceta, J. I., Payros, A., Bernaola, G., Caballero, F. and Apelleniz, E. (2003): Basal Iler-dian (earliest Eocene) turnover of larger foraminifera: Age constraints based on calcareous plankton and 8^{13}C isotopic profiles from new southern Pyrenean sections (Spain). - Geological Society of America, Special Paper, 369: 205-221.

Remane, J., Bassett, M.G., Cowie, J.W., Gohrbandt, K.H., Lane, H.R., Michelsen, O. and Naiwen, W., (1996): Revised guidelines for the establishment of global chronostratigraphic standards by the International Commission on Stratigraphy (ICS). Episodes, 19(3), 77-81.

Rocchia, R., Robin, E., Cachier, T., Lim, B., Leroux, H., Apellaniz, E. and Orue-Etxebarria, X. (1996). Impact remains in K-T boundary sediments from the Basque Country. - Meeting on the Cretaceous-Tertiary boundary: biological and geological aspects, París, Abstracts.

Roggenthen, W.M. (1976): Magnetic stratigraphy of the Paleocene. A comparision between Spain and Italy. Mem. Soc. Geol. Italia, 15: 73-82.

Röhl, U., Bralower, T.J., Norris, R.D. and Wefer, G., (2000): New chronology for the late Paleocene thermal maximum and its environmental implications. Geology, 28(10): 927-930.

Romein, A.J.T. (1979): Lineages in early Paleogene calcareous nannoplankton. Utrecht Micropalaeontol. Bulletins, 22, 1-230.

Rosell, J., Linares, R. y Llompart, C. (2001): El Garumniense prepirenaico. Rev. Soc. Geol. España, 14: 47-56.

Roth, P.H. and Thierstein H. (1972): Calcareous nannoplankton: Leg 14 of the Deep Sea Drilling Project. Initial Reports of the Deep Sea Drilling Project, XIV, 421-485.

Schimper, W.P. (1874): Traite de Paléontologie vegetóle, ou la flore du monde primitif dans ses rapports avec les formations géologiques. JB Bailliére et Fils, Paris.

Schmitz, B. and 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., Asaro, F., Molina, E., Monechi, S., Von Salis, K. and Speijer, R. (1997): High-resolution Iridium, d^{13}C, d^{18}O, foraminifera and nannofossil profiles across the lates Paleocene benthic extinction event at Zumaia, Spain. Palaeogeogr. Palaeoclimatol. Palaeoecol., 133: 49-69.

Schmitz, B., Molina, E. and Von Salis, K. (1998): The Zumaya section in Spain: a possible global stratotype section for the Selandian and Thanetian stages. Newsletters on Stratigraphy, 36: 35-42.

Schmitz, B., Aubry, M.P. and Zachos, J.C. (2000): Preface & summary. GFF, 122: 4-6.

Serra-Kiel, J., Canudo, J.I., Dinarès-Turell, J., Molina, E., Ortíz, N., Pascual, O., Samso, J.M. y Tosquella, J. (1994): Cronoestratigrafía de los sedimentos marinos del Terciario inferior de la Cuenca de Graus-Tremp (Zona Central Surpirenaica). Rev. Soc. Geol. España, 7: 273-297.

Seyve, C. (1990): Nannofossil biostratigraphy of the Cretaceous-Tertiary boundary in the French Basque Country. - Bull. Centres Rech. Expiar. - Prod. Elf-Aquitaine, 14, 2, 553-572.

Smit, J. (1982): Extinction and evolution of planktonic foraminifera after a major impact at the Cretaceous/Tertiay boundary. Geological Society of America, Special Papers, 190: 329-352.

Smit, J. (1990): Asteroid impact, extinctions and the Cretaceous-Tertiary Boundary. Geologie en Mijnbouw, 69: 187-204.

Smit, J. and Ten Kate, W.G.H.Z. (1982): Trace-element patterns at the Cretaceous/Tertiary boundary: consequences of a large impact. Cretaceous Research, 3, 307-332.

Speijer, R.P. (1994): Extinction and Recovery Patterns in Benthic Foraminiferal Paleocommunities across the Cretaceous/Paleogene and Paleocene/Eocene Boundaries.Geologica Ultraiectina 124, 191 pp. (Utrecht University, Netherlands).

Speijer, R.P., Van der Zwaan, G.J. and Schmitz, B. (1996): The impact of PaleoceneI/Eocene boundary events of middle neritic benyhic foraminiferal assemblages fron Egypt. — Marine Micropaleontology, 28, 99-132.

Speijer, R. P. and Samir, A. M. (1997): Globanomalina luxorensis, a Tethian bio-stratigraphic marker of latest Paleocene global events. - Micropaleontology, 43 (1): 51-62.

Speijer, R.P., Schmitz, B. and Luger, P. (2000): Stratigraphy of late Palaeocene events in the Middle East: implications for low- to middle-latitude successions and correlations. Journal of the Geological Society of London 157, 36-46.

Stott, L.D., Kennett, J.P., Shackleton, N.J. and 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: Proceedings of the Ocean Drilling Program, Scientific Results, (eds. P.F. Barker, J.P. Kennett, *et al.*). 113, 849-863.

Ten Kate, W.G. and Sprenger, A. (1993): Orbital cyclicities above and below the Cretaceous/Paleogene boundary at Zumaya (N Spain), Agost and Relleu (SE Spain). Sedim. Geol., 87: 69-101.

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

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

Thomas, E. (2003): Extinction and food at the sea floor: A high-resolution benthic foraminiferal record across the Initial Eocene Thermal Maximum, Southern Ocean Site 690, in Causes and Consequences of Globally Warm Climates of the Paleogene, GSA Spec. Pap., vol. 369, edited by S. Wing *et al.*., 319-332, Geol. Soc. of. Am., Boulder, Colo.

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

Thomsen, E. (1994): Calcareous nannofossil stratigraphy across the Danian-Selandian boundary in Denmark, GFF 116, 65-67.

Thomsen, E. & Heilmann-Clausen, C. (1985). The Danian-Selandian boundary at Svejstrup with remarksks on the biostratigraphy of the boundary in Western Denmark. Bulletin of the Geological Society of Denmark, 33, 341-362.

Tjalsma, R.C. and Lohman, G.P. (1983): Paleocene-Eocene bathyal and abyssal benthic foraminifera from the Atlantic Ocean. Micropaleontology, Spec. Publ., 4, 1-90.

Torrence, C. and Compo, G.P. (1998): A practical guide to wavelet analysis, Bull. Am. Meteorol. Soc. 79, 61-78.

Van de Velde, J. (1967): Geology of the Ordesa region. Estudios Geol., 23: 163-201.

Van der Hurk, A.M. (1990): Eustatic and tectonic controls on carbonate and siliciclastic Paleogene depositional systems in the South Pyrenean foreland basin (Esera, Cinca and Cinqueta valleys; prov. Huesca, Spain). Tesis Doctoral, Univ. Autónoma de Barcelona, 292 p.

Van Morkhoven, F. P. C. M., Berggren, W. A. and Edwards, A. S. (1986): Cenozoic cosmopolitan deep-water benthic foraminifera. Bulletin des Centres Recherches Exploration-Production Elf-Aquitaine, Memoir, 11: 1-423.

Van Vliet, A. (1982): Submarine fans and associated deposits in the lower Tertiary og Guipuzcoa /northern Spain) PhD. Thesis, Univ. Utrecht: 45 p.

Varadi, F., Runnegar, B., and Ghil, M. (2003): Successive refinemements in long-term integrations of planetary orbits, Astrophys. J. 592 620-630.

Vergés, J. (1993):. Estudi geologic del vessant sud del Pirineu oriental i central, Evolució cinematica en 3D, PhD thesis, University of Barcelona, Barcelona.

von Hillebrandt, A. (1964): El Terciario bajo de la zona del Monte Perdido (Huesca). Not. Com. Inst. Geol. Min. España, 73: 61-98.

von Hillebrandt, A. (1965): Foraminiferen-Stratigraphie im Alttertiär von Zumaya (Prov. Guipuzcoa, NW Spanien) und ein Vergleich mit anderen Tethys-Gebieten. Abh. Königl. Bayer. Akad. Wissench. München, 123: 1-63. Young

Von Salis, K., Monechi, S., Bybell, L.M., Self-Trail, J. and Young, J.R. (2000): Remarks on the calcareous nannofossil genera Rhomboaster and Tribrachiatus around the Paleocene/Eocene Boundary. GFF, 122: 138-240.

Ward, P., Kennedy, K.G., McLeod, K.G. and Mount, J. (1991): End-Cretaceous molluscan extinction patterns in the Bay of Biscay K/T boundary sections: two different patterns. Geology, 19: 14-81.

Ward, P. and Kennedy, K.G. (1993): Maastrichtian ammonites from the Biscay region (France, Spain). Jour. Paleontol., 67: 1-58.

Wiedmann J. (1986): Macro-invertebrates and the Cretaceous-Tertiary boundary. In: Global Bio-Events. (Ed. O. Walliser) Lecture Notes in Earth Sciences, 8, 397-409.

Wiedmann, J. (1988): The Basque coastal sections of the K/T boundary: a key to understanding "mass extinction" in the fossil record. En: Paleontology and evolution: Extinction events (M.A. Lamolda, E.C. Kauffman y O. Walliser, Eds.), Rev. Esp. Paleontología, (nº extr.), 127-140.

Wood, H. E., II, Chaney, R.W., Clark, J., Colbert, E. H., Jepsen, G. L., Reeside, J. B. and Stock, C., (1941): Nomenclature and correlation of the North American continental Tertiary: Geological Society of America Bulletin, v. 52, p. 1–48.

Zachos, J. C., Arthur, M. A. and Dean W. E. (1989): Geochemical evidence for suppression of pelagic marine productivity at the Cretaceous/Tertiary boundary. Nature 337, 61– 64.