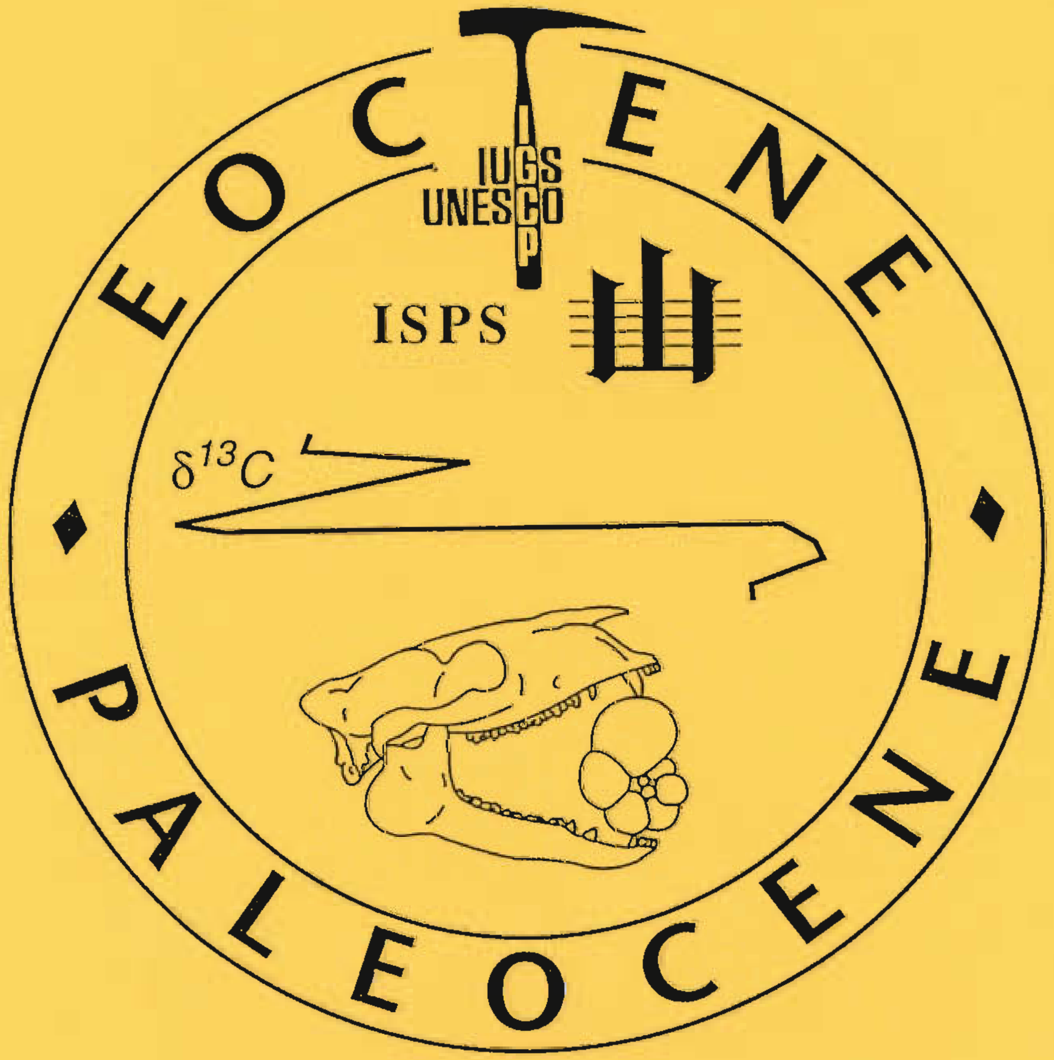
Early Paleogene Stage Boundaries

INTERNATIONAL MEETING AND FIELD CONFERENCE
IN
ZARAGOZA (SPAIN)
June 24 - 29, 1996



Abstracts and Field Trip Guides

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Organized
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Field trip guide to the Paleocene and Early Eocene of Zumaya section

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INTRODUCTION

The stratigraphic section chosen for the field trip is easily accesible since exposure occurs along San Telmo beach, just north of the township of Zumaya (Guipúzcoa province, northern Spain). (Fig. 1). Lithologically the lower Paleocene consist of about 50 metres of limestones with intercalated thin marl layers, the upper Paleocene is composed of about 120 metres of marls with interbedded limestones and calcarenite strata, and marls in the uppermost part. Turbidite strata are rare across the Paleocene stage boundaries and Paleocene/Eocene boundary, but become common in the upper part of the section from the *M. subbotinae* Biozone (Early Eocene) and in the upper Cretaceous.

The Paleocene and Early Eocene sedimentation in the Pyrenean Basin appears relatively continuous although there are differences between the sections. The Zumaya section was deposited in the deep part of a basin which outcrops along the beaches at the northern coast range of the iberian peninsula. The lithology and geologic setting of the Zumaya section was described in detail by Hillebrandt (1965) and Pujalte *et al.*, (1989). The depositional sequences have been established by Pujalte *et al.* (1993; 1994; 1995). Orue-Etxebarria and Lamolda (1985) studied the paleogeography of the Basque-Cantabrian basin by means of planktic foraminifera from the Paleocene to Middle Eocene and they concluded that deposition ocurred in a tropical to temperate province during the Paleogene. Futhermore, Orue-Etxebarria *et al.* (1996) have published an integrated biostratigraphy in other sections of the Basque Country (Ermua and Trabakua Pass), which are reference sections for the Paleocene/Eocene boundary. Excellent reference sections are also the Campo and Tremp sections in the area of the Ilerdian stratotype (Molina, 1994; 1996; Molina *et al.*, 1992; 1995; 1996).

Hillebrandt (1965) first published a significant biostratigraphic analysis of the planktic foraminiferal fauna from the Zumaya section but did not discuss details across the Paleocene stage boundaries and Paleocene/Eocene boundary. Planktic foraminifera and calcareous nannofossils are the major fossil components in the sediments across the Paleocene-Eocene boundary at Zumaya section. Benthic foraminifera and ostracodes occur in low percentage (<10%). Larger foraminifera and other macrofossils are very rare. Reworked fossils are extremely rare and only occur in the turbiditic levels, which are very well localized.

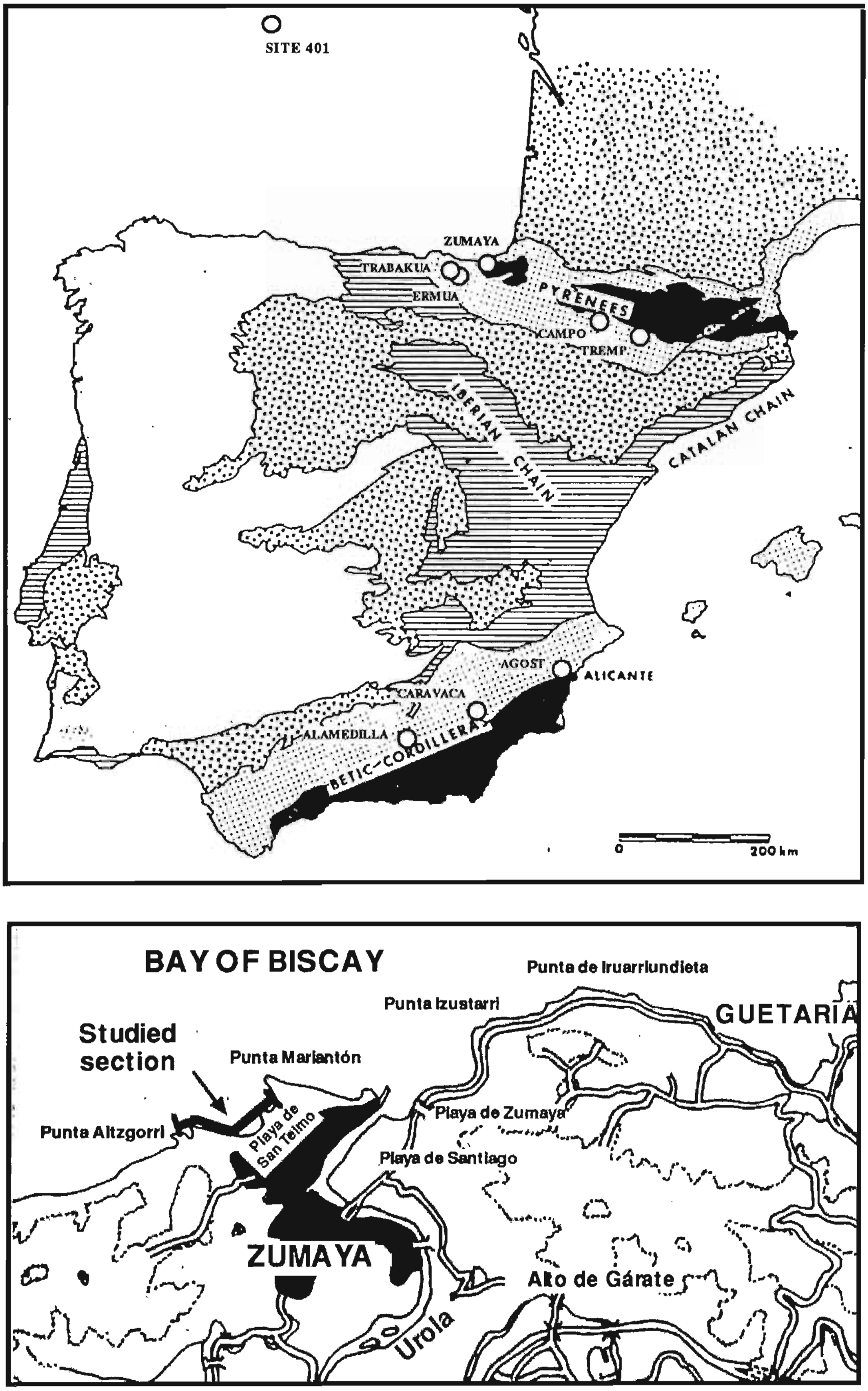

Fig. 1.- Location of the Zumaya section

PALEOCENE STAGE BOUNDARIES

One of the most promising sections for providing GSSP's for the Paleocene stage boundaries is the Zumaya section (Molina, 1994; 1996, Schmitz *et al.*, 1996). This section consists of an expanded about 160 metres thick and biostratigraphically complete Paleocene record (Fig. 2). The Zumaya section is an important link between the historical stratotype sections and the expanded Paleocene sections at low-middle latitudes in the Tethys and Atlantic Ocean.

According to Rosenkrantz (1924) (*fide* Schmitz *et al.*, 1996) the original basal Selandian in Denmark was defined at a major lithological shift, from greyish white, oxic limestone/chalk to suboxic greensand overlain by grey marl and clay. The limestone/chalk deposition prevailed throughout the entire Maastrichtian and Danian. After the Danian-Selandian transition, grey marls and clays dominated the sedimentation in Denmark for several million years. In terms of calcareous nannoplankton zonations, the basal Selandian in Denmark falls in the upper part of Zone NP4 or lower NP5(Thomsen, 1994).

Consequently, at Zumaya the base of the Selandian appears to correspond to a package of red marls forming a 3.4 metres thick transition interval from reddish limestone-marl couplets below to grey marls above, about 55 metres above the Cretaceous/Tertiary boundary. This lithological shift is one of the most prominent in the Zumaya section. Thick reddish limestones with some marl layers dominate the lower 45 metres of the Danian. They are followed by some 10 metres of reddish limestone-marl couplets. The overlying 3.4 metres of red marls change abruptly into grey marls, which thereafter persist for some 25 metres. A varied grey marl-limestone lithology characterizes the remainder of the Paleocene.

The entire Paleocene of Zumaya section has been analyzed for stable isotopes in whole-rock samples (0.25 to 0.5 metres spacings). In the interval in which the basal Selandian may be defined, there is a negative $\partial^{13}C$ shift of about 1 per mil (Schmitz *et al.*, 1996). The Danian/Selandian boundary is a prominent event in term of $\partial^{13}C$, nannofossils and foraminifera (Fig. 3).

Planktic foraminifera were studied by Arenillas *et al.* (1993), Arenillas and Molina (1995; 1996) and Arenillas (umpublish Ph. D. thesis). (Fig. 4). Near or at the base of the red marls four species have their last appearance (*Eoglobigerina trivialis, Acarinina indolensis, Acarinina kubanensis* and *Morozovella tadjikistanensis*), and two species have their first occurrence (*Morozovella crosswickensis* and *Chiloguembelina crinita*). In the red marls, or just above these, *Subbotina velascoensis* has its first appearance. Genus *Morozovella* strongly decreases at the Danian/Selandian transition that is characterised by three levels in which the percentage of planktic foraminifera decreases and the percentage of the small benthic foraminifera increases (Arenillas and Molina, 1996). In coincidence with the upper two levels there is an increase of the genus *Chiloguembelina* which proliferates when oxygen minimum zones are well developed (Arenillas and Molina, 1995). Consequently, these data support the correlation of the Zumaya grey marls with the basal Selandian stratotype grey suboxic clay/marl. As the upper part of *Morozovella angulata* Biozone (P3a) does not correspond to the Danian we have used a new biozone called *Morozovella crosswicksensis* to divide the *M. angulata* Biozone. The base of the new biozone approximately coincides with the base of the red marls. This is the main event in term of planktic foraminifera, but one meter above, in coincidence with the appearance of the nannofossil *F. tympaniformis*,

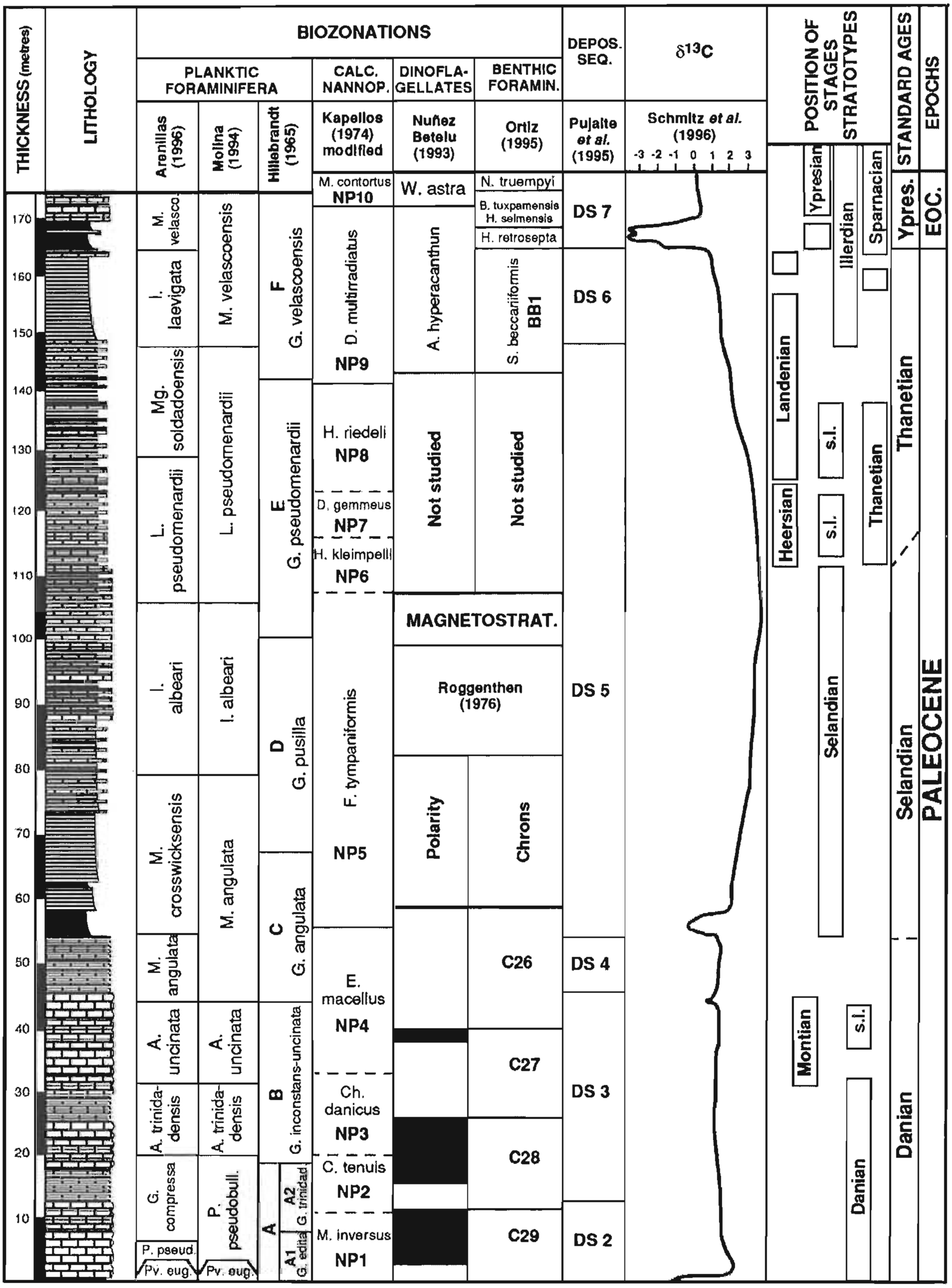

Fig. 2.- Integrated stratigraphy of the Paleocene at Zumaya section (modified from Molina, 1994)

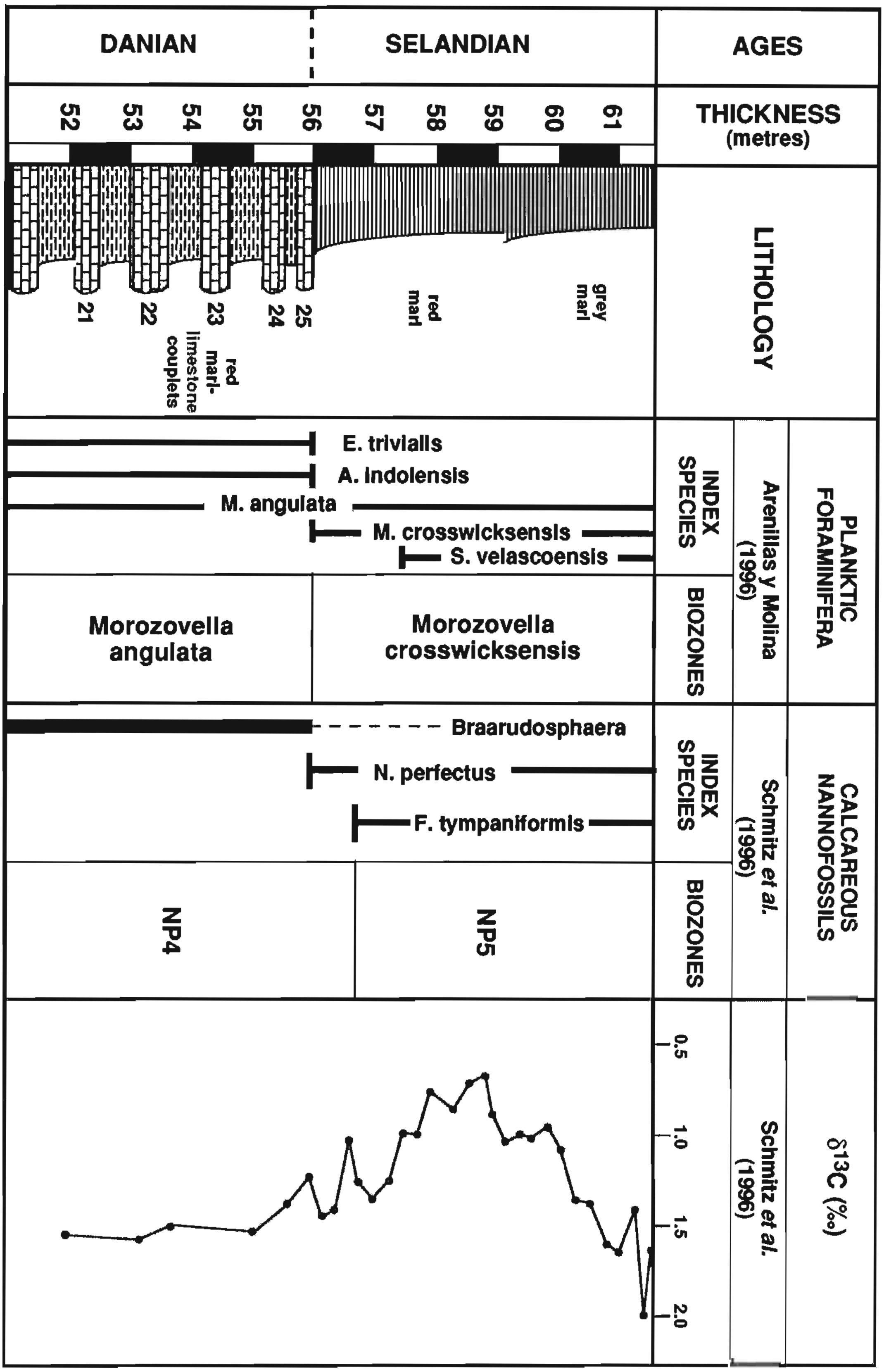

Fig. 3.- Integrated stratigraphy across the Danian/Selandian boundary at Zumaya section (modified from Schmitz *et al.*, 1996)

Subbotina velascoensis has its first appearance, which could be also used to define the Danian/Selandian boundary.

The calcareous nannofossils at Zumaya were studied by Kapellos (1974) and restudied by Von Salis in Schmitz *et al.* (1996). The first representatives of the genera *Sphenolithus* and *Fasciculithus* were found in the red limestone-marl couplets, that also include occasional *Ellipsolithus macellus*, and thus can be assigned to the upper part of NP4 Zone. The lowermost *Fasciculithus tympaniformis*, the marker for the base of NP5 Zone, was observed 75 centimetres above the lithologic change to the red marls, while the lowermost occurrence of typical *Neochiastozygus perfectus* was noted at the very base of the red marls. A sharp decrease of *Braarudosphaera bigelowi* s.l. accompanies the lithologic shift from the reddish limestone-marl couplets to the red marl. No major change in nannofossil flora was found at the shift from red to grey marl, however, a sharp increase in the abundance of microscopic dolomite rhombs was observed (Fig. 3).

Consequently, there are a number of arguments in favor of that the transitions from dominantly oxic limestones to grey suboxic marls or clays in Denmark and Spain reflect the same event. The nannofossil data show that at both sites the lithological shifts occur near or at the trasition from Zone NP4 to NP5. Moreover, the finding of *Neochiatozygus perfectus* at the base of the red marls at Zumaya concurs with its first appearance in Denmark in the uppermost metres of the Danian (Thomsen, 1994). Pujalte *et al.* (1994) suggest that one of the most important sequence boundaries in the Zumaya section occurs in the interval in which we think that the basal Selandian occurs. This is the boundary between depositional sequences DS4 and DS5 according to their sequence boundary as the one marking the unconformity at the Danian-Selandian boundary in Denmark, and which represents the Se-1 sequence boundary in the chart of Hardenbol (1974). The sequence boundaries may not be correlatable woldwide, it is likely that at least regionally, along the northeastern Atlantic coast the same sea-level changes are registered by the strata. Magnetostratigraphy shows that the upper part of the limestone-clay couplets at Zumaya belongs to the Chron C26r (Roggenthen, 1976), which is the same chron preliminarily assigned for the basal Selandian in Denmark (Ali *et al.*, 1994). It is noteworthy that the basal Selandian interval at Zumaya is situated about 10 metres above the boundary between the *P. uncinata* (P2)/*M.angulata* (P3a) planktic foraminiferal zones, which in the general stratigraphic sheme of Berggren *et al.* (1995) is considered to correspond to the Danian/Selandian boundary.

The base of the Thanetian in its original type area is an unconformity dated to the upper part of Zone NP6 (Aubry, 1994; Knox, 1994). The preliminary nannofossil data suggest that the NP6 zone corresponds approximately to the interval around 100 metres above the Cretaceous/Terciary boundary at Zumaya section. In term of planktic foraminifera the base of the Thanetian is been classically placed in the lower part of *P. peudomenardii* Zone (P4).

Apparently, in Zumaya there is no important sequence boundary near or within this interval (Pujalte *et al.*, 1994). The closed sequence boundaries are the above mentioned DS4/DS5(Danian/Selandian) boundary and the DS5/DS6 boundary at about 145 metres above the Cretaceous/Tertiary boundary. The DS5/DS6 boundary occurs in NP9 Zone above the Thanetian in England and corresponding to the basal Ilerdian in Spain. According to Arenillas and Molina

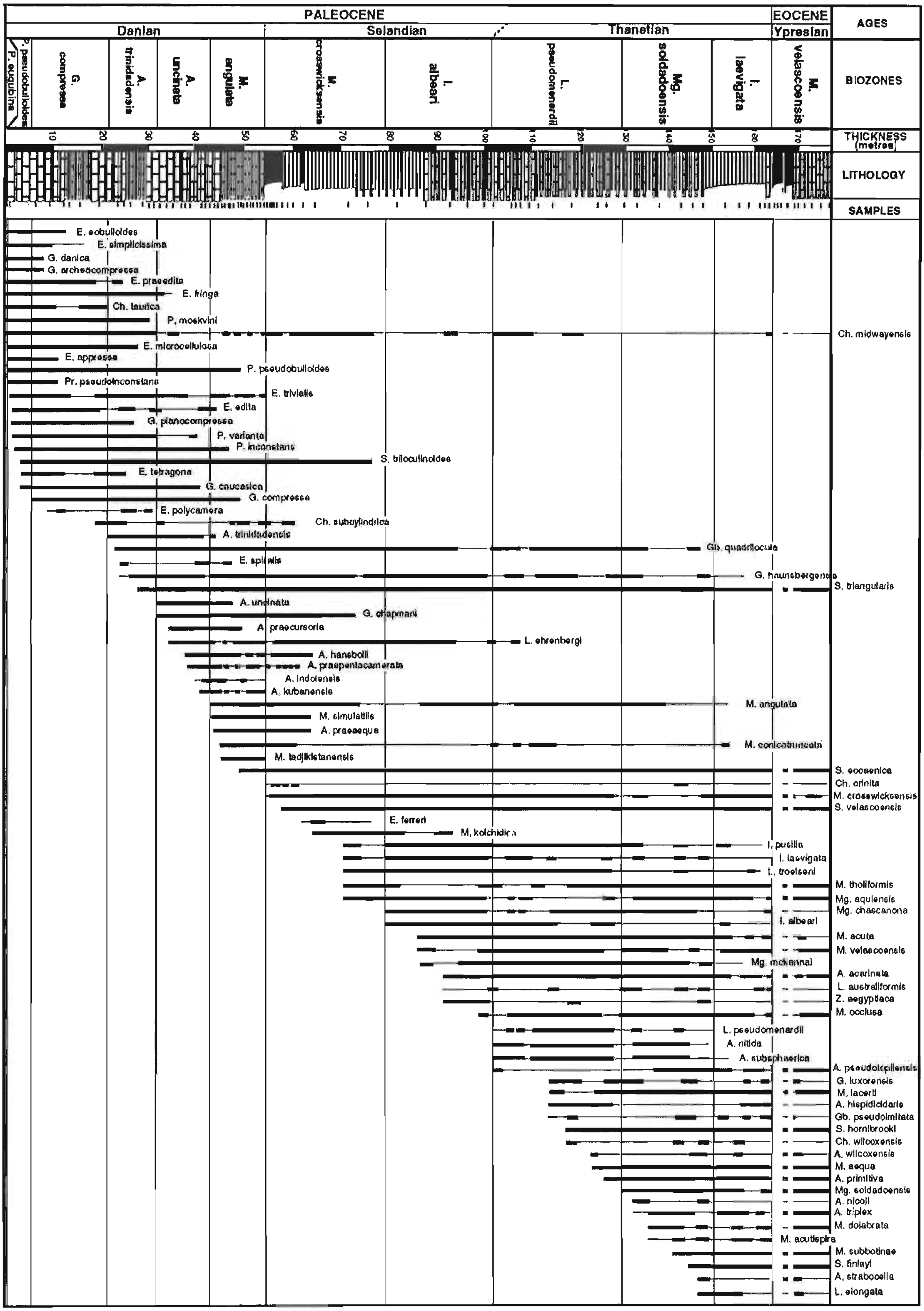

Fig. 4.-Planktic foraminiferal biostratigraphy of the Paleocene at Zumaya section (modified from Arenillas, umpubl. Ph. D. thesis)

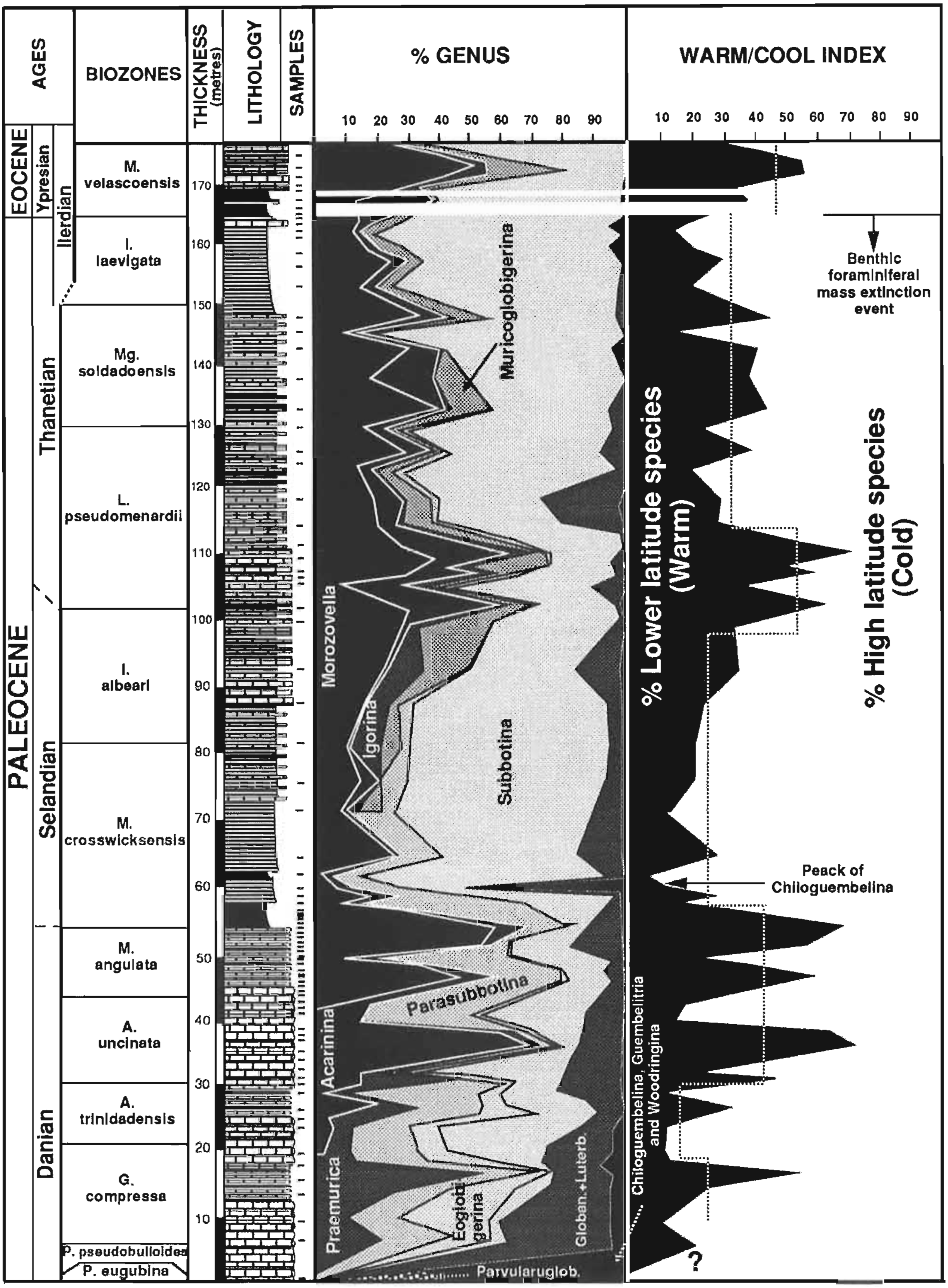

Fig. 5.-Planktic foraminiferal quantitative analysis of the Paleocene at Zumaya (modified from Arenillas and Molina, 1995)

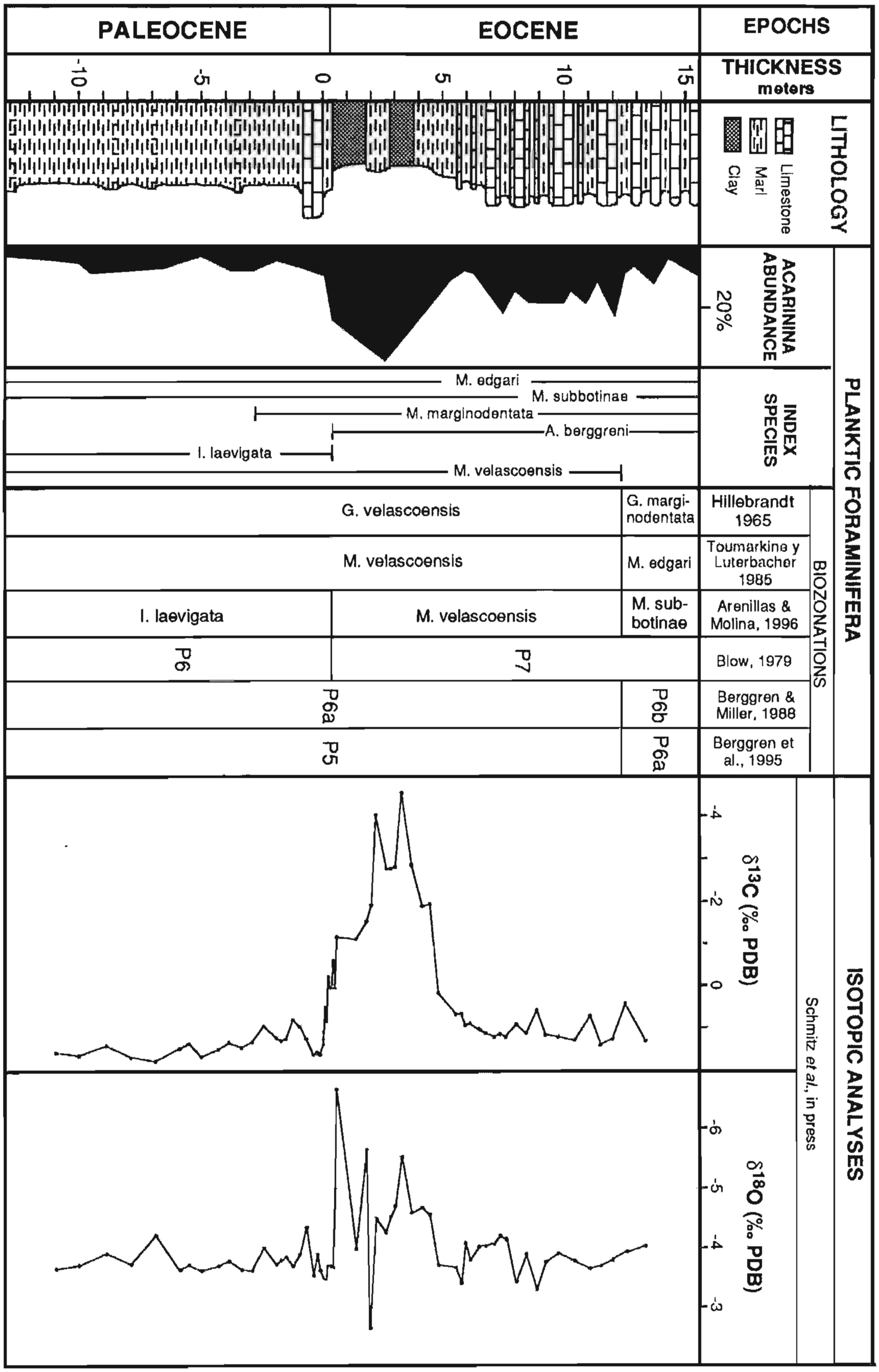

Fig. 6.-Planktic foraminiferal and isotopic analyses across the Paleocene/Eocene boundary at Zumaya section (modified from Schmitz *et al.*, in press)

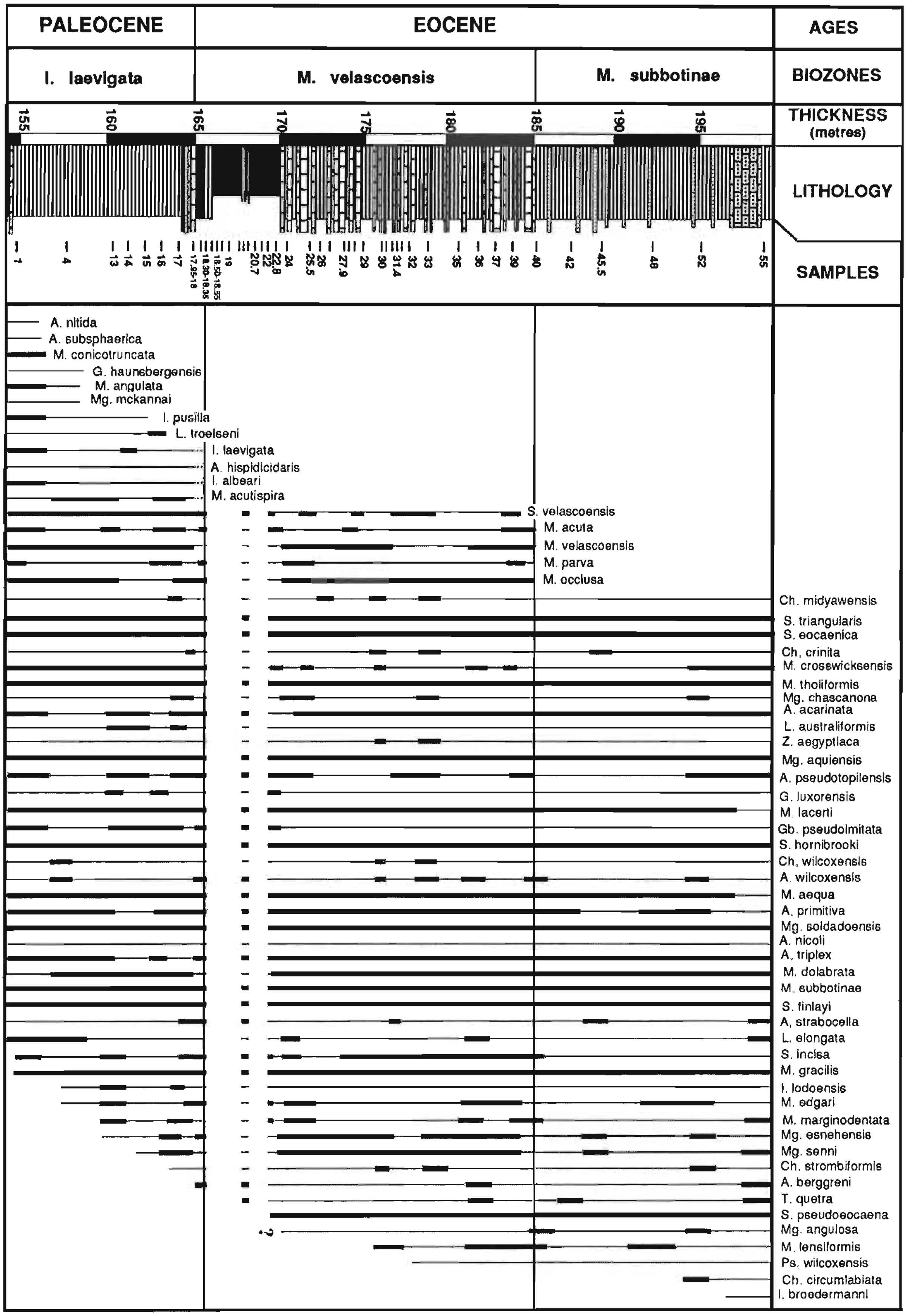

Fig. 7.- Planktic foraminiferal biostratigraphy across Paleocene/Eocene boundary at Zumaya section (modified from Canudo y Molina, 1996, and Arenillas, umpubl. Ph. D. thesis)

(1995) and Arenillas (umpublish Ph. D. thesis) in the lower part of *L. pseudomenardii* Biozone there are very relevant changes in the planktic foraminifera warm/cold index at Zumaya section, indicating an increase in temperature across the Selandian/Thanetian boundary (Fig. 5).

PALEOCENE/EOCENE BOUNDARY

The lower part of the Paleocene-Eocene (P-E) transition at Zumaya consists of about 17 metres of homogenous grey marls; the upper part consists of more than 30 metres of grey marls alternating with grey limestone beds. A prominent one-meter-thick grey sandy limestone bed interrupts the marl sequence. A turbidite (8 centimetres thick) is found in the lower part of the grey limestone bed. Such turbidites are very rare in the Paleocene-Eocene interval. The limestone is overlain by 45 centimetres thick dark marl layer, which is succeeded by a 4 metres interval of red clay that marks the benthic foraminiferal mass extinction and $\delta^{13}C$ shift. The clay interval is devoid of original calcite except for a thin calcareous interval in the middle part.

Placement of the Paleocene/Eocene (P/E) boundary has not been officially defined. By convention, many biostratigraphers used to place this boundary at the extinction of *Morozovella velascoensis* that marks the P5/P6a boundary and is assigned an age estimate of 54.7 Ma (Berggren *et al.*, 1995). Since the last occurrence of *M. velascoensis* appears diachronous and this species is not present in high latitudes or in shallow sections, this datum is a very poor marker for a major series boundary. The *Pseudoshastigerina* "datum" has also been used by some authors to define the Paleocene/Eocene boundary, but the *Pseudohastigerina* first appearance is very diachronous (Molina *et al.*, 1992, Canudo & Molina, 1992; Pardo *et al.*, 1994). The *Pseudohastigerina wilcoxensis* first appearance is not a reliable datum, since it appears later in high latitudes and because of the difficulty in discriminating it from its ancestor *Globanomalina ovalis*. The $\delta^{13}C$ excursion, which has been determined to be globally synchronous and coincides with a major mass extinction in benthic foraminifera (Kennett and Stott, 1991; Kaiho, 1991; Pak and Miller, 1992) appears to be the best and most reliable marker horizon for the Paleocene/Eocene boundary. Consequently, we have placed the P/E boundary at this $\delta^{13}C$ excursion and extinction in benthic foraminifera (BB1/BB2 Zone of Berggren and Miller, 1989).

The Zumaya section contains the most biostratigraphically complete Paleocene-Eocene transition known to date, with about 30 metres representing Zones P5 and P6a of Berggren and Miller (1988), which correspond to Zone P5 of Berggren *et al.* (1995) and to the *Igorina laevigata* and *Morozovella velascoensis* Biozones of Arenillas and Molina (1996). This represents the most expanded interval for this zone known to date in deep sea sections. By comparison, Zone P6a is represented only by 1.2 m in the Pacific Ocean deep-sea Site 577 (Miller *et al.*, 1987), by 6.4 m in the Bay of Biscay Site 401 (Pardo *et al.*, 1995; 1996; Arenillas *et al.*, 1996) and by 6.5 m for Zones P5 and P6a combined (Pak and Miller, 1992).

Ranges of selected index species and the planktic foraminiferal biozonations are illustrated in Figure 6 along with the $\delta^{13}C$ and $\partial^{18}O$ curves. The $\delta^{13}C$ excursion coincides with a 45 centimetres thick dark grey shale layer, increased carbonate dissolution and the extinction of over 50% of the benthic foraminifera (Ortiz, 1993; 1995; Canudo *et al.*, 1995; Schmitz et al., in press). No significant planktic foraminifera species turnover is apparent across the clay. The most important change across the P/E boundary is the strong increase in the relative abundance of *Acarinina*, reaching the highest values just above the benthic extinction in the marl interval interbedded in the clay. It is notable that the *Acaninina* increase is already apparent 20 centimetres below the benthic extinction.

The ranges and relative abundances of all planktic foraminiferal species are illustrated in Figure 7. The Paleocene-Eocene transition is characterized by a gradual species turnover with about 30% (17 species) of the taxa terminating and about 20% (12 species) originating primarily in this interval. The lower 10 metres (*I. laevigata* Biozone) is characterized by the gradual disappearance of 8 species and the appearance of 6 species. The Paleocene/Eocene boundary is marked by the simultaneous desappearance of 4 species. The upper 20 metres (*M. velascoensis* Biozone) is characterized by the gradual appearance of 6 species and the desappearance of 5 species at the top of this zone (Fig. 7).

Biostratigraphic interpretation of Zumaya is based on planktic foraminifera following numerical zonations of Blow (1979), Berggren and Miller (1988) and the last revision of Berggren *et al.* (1995). In order to avoid confusion inherent to sucesive changes in numerical zonations we recongnized the classical biozonations of Hillebrandt (1965), Tourmarkine and Luterbacher (1985) and the new biozonation of Arenillas and Molina (1996). It has been proposed that Zone P5 be eliminated because the defining datum events, *L. pseudomenardii* LO and *M. subbotinae* FO, may even overlap due to diachroneity (Blow, 1979, Canudo and Molina, 1992a). This may give the erroneous impression of a hiatus with Zone P5 missing. Since the Spanish sections must be correlated to coeval sections worldwide where the P-zonation of Berggren and Miller (1988) has been used, our biostratigraphic interpretation here also follows this zonation. Nevertheless, in order to improve the biostratigraphical ressolution we have further divided the classical *M. velascoensis* Biozone: *Igorina laevigata* Biozone and *Morozovella valascoensis* Biozone (Arenillas and Molina, 1996). This division has the advantage of a higher resolution and the coincidence of the zonal boundary with the small benthic foraminiferal extinction.

Benthic foraminiferal data from Zumaya indicate that deposition across the P-E transition occurred probably in a middle or lower bathyal environment. This is indicated by the presence of many species which have upper depth limits between 600 m. and 1500 m. (e.g. *Spiroplectammina spectabilis, B. jarvisi, Cibicidoides hyphalus, Stensioina beccariiformis, Bulimina velascoensis, B. trinitatensis, B. tuxpamensis, C. pseudoperlucidus, Nuttallides truempyi, Oridorsalis umbonatus*, Morkoven et al., 1986). Planktic foraminifera represent more than 85% of the foraminiferal fauna. The presence of common *C. pseudoperlucidus* and *S. beccariiformis* in the *L. pseudomenardii* and *I. laevigata* Biozones prior to the $\delta^{13}C$ shift indicates well oxygenated waters (Kaiho, 1991) and the common occurrence of *Cibicidoides* spp. suggests a high influx of terrestrial organic carbon (Boersma, 1985, Ortiz, 1993;1994).

The extinction of over 50% of the benthic taxa coincides with the $\delta^{13}C$ excursion and deposition of dark grey to brown shale layers with low carbonate content and strong foraminiferal dissolution. At Zumaya, a low diversity agglutinated assemblage and high abundance of *Bulimina tuxpamensis* suggest low oxygen conditions and waters undersaturated in calcium carbonate during the $\delta^{13}C$ excursion. During the following interval of low $\delta^{13}C$ values, high abundance of *N. truempyi* (47%) suggests a return to more oxygenated waters (Ortiz, 1993; 1995, Canudo *et al.*, 1995).

Calcareous nannofossil were studied by Kapellos (1974) establishing the biozonation from the uppermost Cretaceous to the upper part of the Early Eocene. The Paleocene-Eocene transition is being studied in detail by Angori and Monechi and a preliminary biozonation was previously established by Aubry. According to Angori and Monechi the NP9/NP10 boundary is located 5 metres above the benthic foraminiferal mass extinction (Angori and Monechi, 1995, pers. comm.) and according to Aubry the same boundary is placed more than 20 metres above the benthic foraminiferal mass extinction (Aubry, 1990, pers. comm.) in coincidence with the CP8/CP9 boundary, but this boundary does not coincide with the NP9/NP10 boundary. This divergence is due to a different interpretation of the index species and not to a hiatus or condensation above the Paleocene/Eocene boundary at Zumaya section.

Dinoflagellates were studied by Nuñez-Betelu (1993, in press) establishing the biozonation of the Paleocene-Eocene transition at Zumaya section. The following zones were recognized: *A. hiperacanthum* Biozone across the Paleocene/Eocene boundary and *W. astra* Biozone above the dissolution interval in the lower Eocene. These results are similar and can be easily correlated to the nearby sections of Ermua and Trabakua Pass (Orue-etxebarria *et al.*, 1996).

Chemical element (Ir, Fe, Sc, Ni, Co, Se, Sb, Cr, Rb, Cs, Ce, Hf, and Th) concentration were analyzed with the Luis W. Alvarez-Iridium Coincidence Spectrometer. The stable isotope analyses were performed on bulk samples, using a VG Prism Series II mas spectrometer with an Isocarb automated carbonate preparation system (Schmitz et al., in press). It appears that at Zumaya the negative $\delta^{13}C$ shift associated with the benthic extinction begins to evolve gradually somewhere between 20 to 40 centimetres below the base of the dissolution clay (Fig. 6). Based on sedimentation rate calculations it appears that the benthic mass extinction event at Zumaya occurred about 6 to 25 kyrs after the initiation of the sea-water $\delta^{13}C$ shifts.

There are two paleontological arguments in favor of that the $\delta^{13}C$ gradual change reflect the initiation of the events leading to the benthic mass extinction. According to Speijer in Schmitz *et al.* (in press) at Zumaya there is a 15 centimetres offset between the LAD's of *G. beccariiformis* and the other benthics. Secondly, in the calcareous intervals of the dissolution clay an unusual planktic foraminiferal fauna exists, with acarinids, in particular *A. wilcosensis* and *A. acarinata*, showing major abundance peaks. These two species, however, already show their first strong abundance increase in a sample from the greenish marls at 20 centimetres below the benthic extinction. The $\partial^{18}O$ values lie relatively stable throughout the marl and limestone part of the section, with most values in the range -3.5 to -4 per mil. In the clay interval, however, $\partial^{18}O$ shows a widely scattered distribution, ranging between -2.6 to -6.6 per mil. These isotopic shifts has been atributed to a change in the source regions of global deep-water from the high latitudes to subtropical regions in connection with a high-latitude warming event. Warm, saline surface water began to spread all over the global see floor, leading to benthic faunal turnovers (Kennett and Stott, 1991; Schmitz *et al.*, in press) and producing a major sea level rise (Molina *et al.*, 1994).

In conclusion, the Zumaya section is located in a very accesible place, has a very continuos, expanded, well exposed, and very good paleontological and geochemical records. Consequently, this section is considered as one of the best candidates to define the Paleocene/Eocene boundary stratotype. Futhermore, the definition of the Paleocene stage boundaries (Selandian and/or Thanetian) in the same section represents a unique and advantageous opportunity.

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Field trip guide to the Paleocene-Middle Eocene of Agost, Caravaca and Alamedilla sections

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INTRODUCTION

The stratigraphic sections visited in the field trip are geologically located in the Betic Cordillera and geographically placed in the south-east of Spain. Agost is located near Alicante, Caravaca in Murcia province and Alamedilla in the north of Granada province (Fig 1).

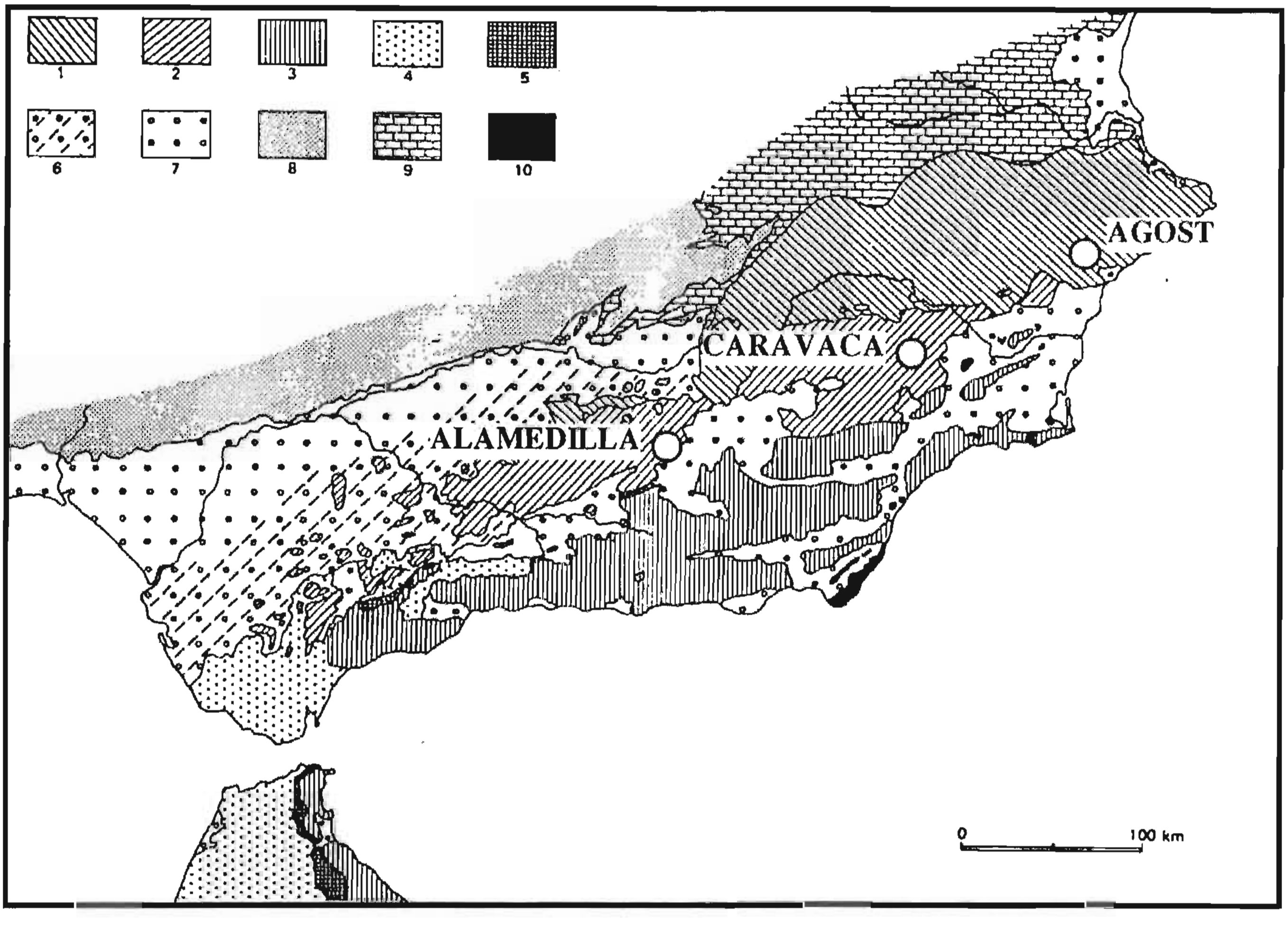

Fig. 1.- Geological location of the Agost, Caravaca and Alamedilla sections

AGOST SECTION

The section is located about 1 kilometre north-west of the town (Fig. 2). At Agost a very complete section from the upper Cretaceous to the Middle Eocene can be studied. Surface exposures of the K/T boundary can be found in a roadcutting near the 13 km marker post in the road to Castalla. This is probably the most complete and expanded K/T boundary section in Spain and have been studied by Groot *et al.* (1989), Canudo *et al.* (1991), Pardo *et al.* (1996), Molina *et al.* (1996), Arenillas (umpubl. Ph. D. thesis) and Arz (umpubl. Ph. D. thesis).

The Paleocene outcrops westward across Lomas de la Beata. The Paleocene/Eocene transition is not very well exposed and is characterized by a 50 metres thick series of turbidites containing debris of *Discocyclina* sp. and *Operculina* sp. from Paleocene platform deposits. The lower Eocene can be sampled in three small abandoned quarries at the north of Agost, and consists of pelagic marls interbedded with turbidites that contain abundant larger foraminifera.

The lower-middle Eocene transition can be sampled on the road from the abandoned quarries to a pottery factory. The section was first studied by Hillebrandt (1974) who tried to correlate the planktic and larger foraminifera. Both groups were also studied by Márquez (1979; 1983). The planktic foraminifera were studied by Cremades (1982) who established a biostratigraphical zonation from the Late Paleocene to the Middle Eocene. A more detailed study of the larger foraminifera was carried out by Isuman (1983) who recognized the following biozones: "small *Nummulites*" zone (Upper Cuisian), *Nummulites obesus* zone (Early Lutetian) and *Nummulites beneharnensis* zone (Middle Lutetian). Nevertheless, this author did not zoned the Y/L boundary and this biozonation should be revised and completed, since this section has a great potential for the correlation between planktic and larger foraminifera (Fig. 3).

The Early-Middle transition at Agost section is composed of marls with some prominent calcarenite intercalations, which are very well developed just at the Y/L boundary. The calcarenite strata contain abundant larger foraminifera at the base, indicating turbiditic sedimentation. The upper part of the section is composed predominantly of marls and contains abundant and well preserved planktic foraminifera of the *Hantkenina nuttalli* and *Globigerapsis subconglobata* zones. There is a sequence of gradual appearance of the following species: *Subbotina boweri, Subbotina frontosa, Morozovella dolobrata, Morozovella spinulosa, Globigerapsis subconglobata, Truncorotaloides rohri, Truncorotaloides topilensis, Hantkenina nuttalli, Hantkenina mexicana, Hantkenina dumblei, Morozovella lehneri* and *Globigerapsis index.*

Consequently, the Early/Middle Eocene boundary (Ypresian/Lutetian boundary) has not been studied in detail. Our recent sampling and the preliminary study allow us to locate the Ypresian/Lutetian boundary, using the first appearance of *Hantkenina* (Gonzalvo and Molina, in preparation), nevertheless a high resolution and integrated stratigraphical study is necessary to precise the exact position of the Y/L boundary (Fig. 4).

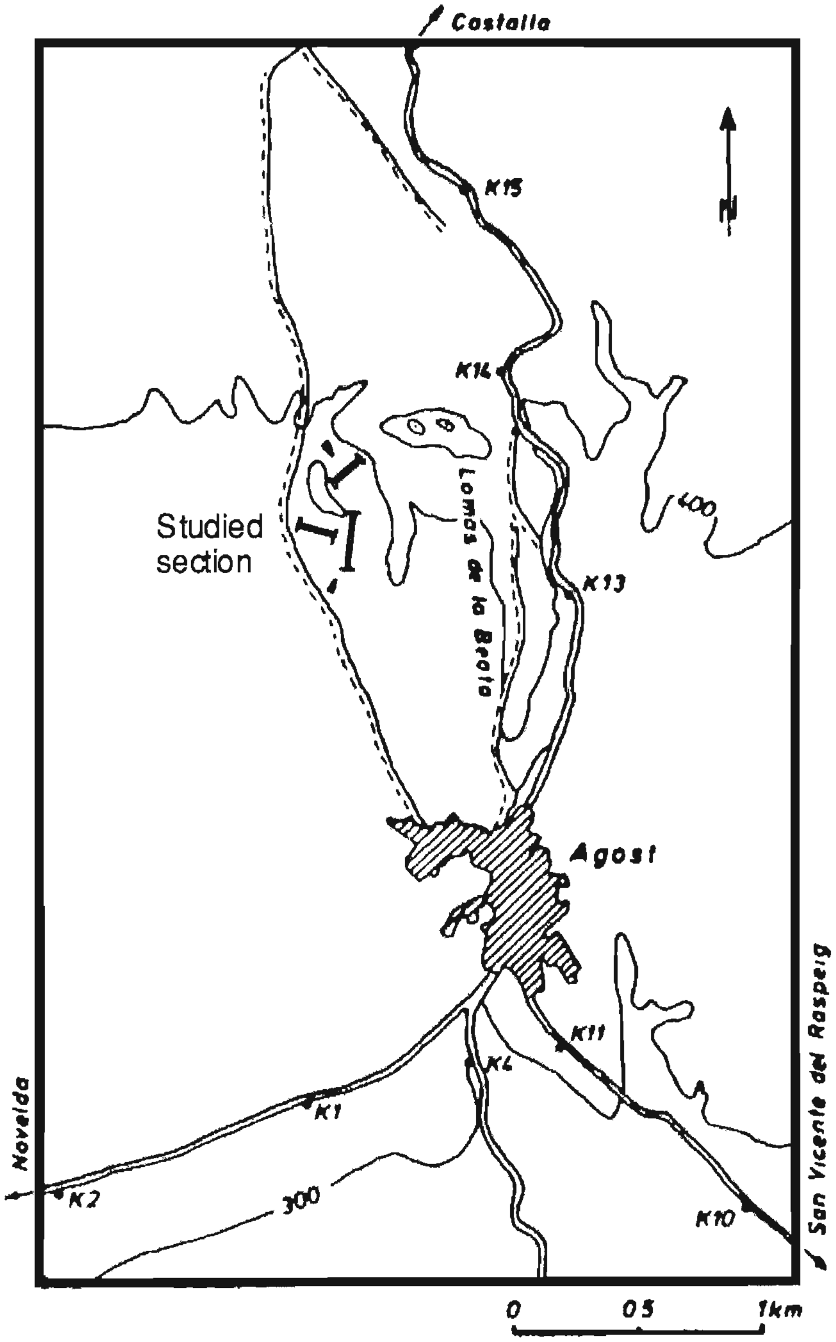

Fig.2.- Geographical location of the Ypresian/Lutetian transition at Agost

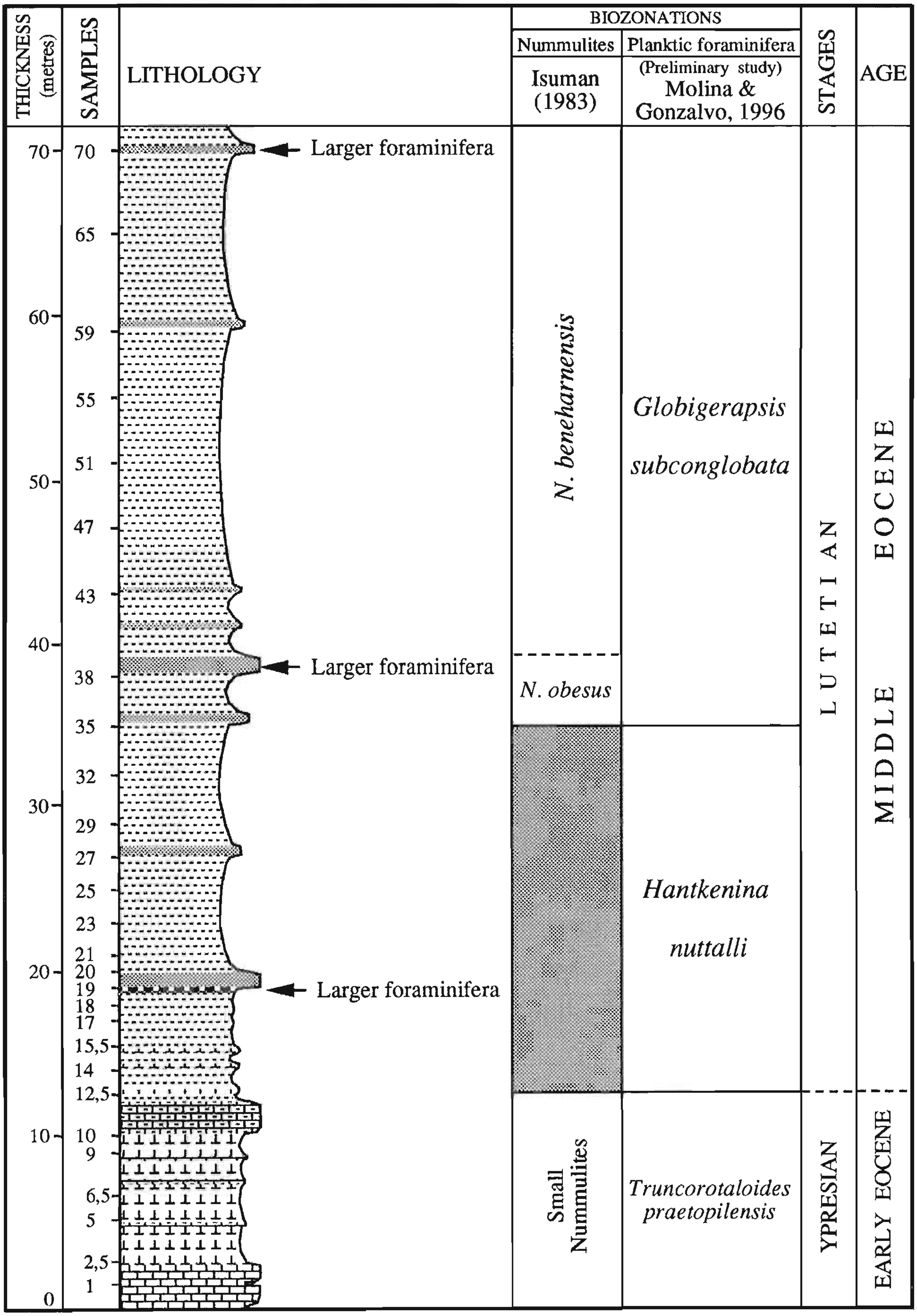

Fig. 3 : Planktic and larger foraminiferal biostratigraphy at Agost section.

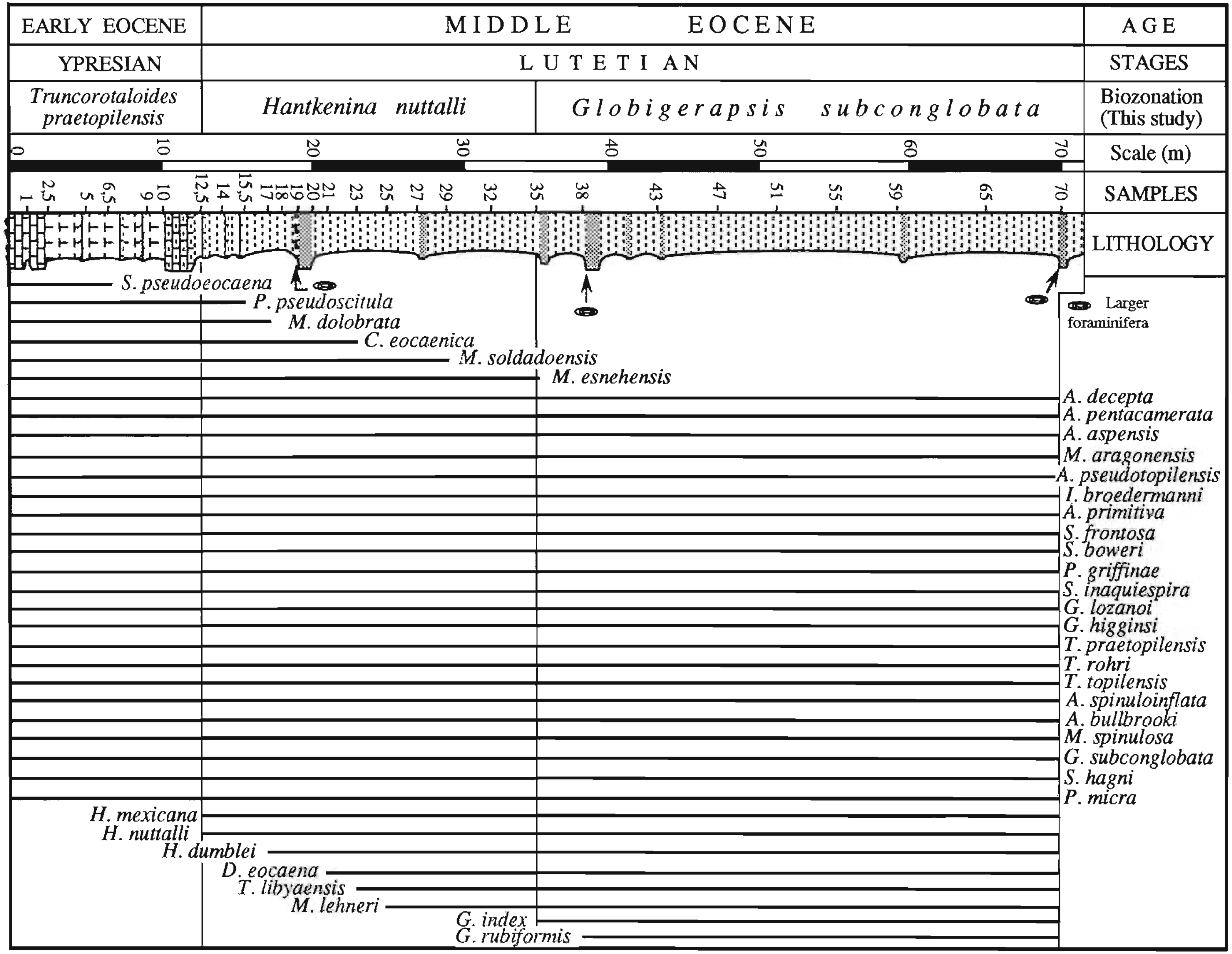

Fig. 4 : Planktic foraminiferal distribution and biozonation at Agost section.

CARAVACA SECTION

The classical Caravaca section, very well known world wide because of the Cretaceous/Tertiary boundary (Smit, 1977), contains a nearly continuous marine record spanning from the Late Cretaceous to the Middle Eocene. The section was initially studied by Durand Delgá & Magné (1958) and later in more detail by Von Hillebrandt (1974) based on foraminifera and by Romein (1979) based on calcareous nannoplankton. Recently, the K/T boundary has been studied by Canudo et al. (1991).

The Caravaca section is located in the Barranco del Gredero 4 km south of the town of Caravaca de la Cruz, Murcia province, southern Spain. The section is easily accessible by the road from Lorca to Caravaca (Fig. 5). Geologically, the section is located in the Subbetic Zone of the Betic Cordillera.

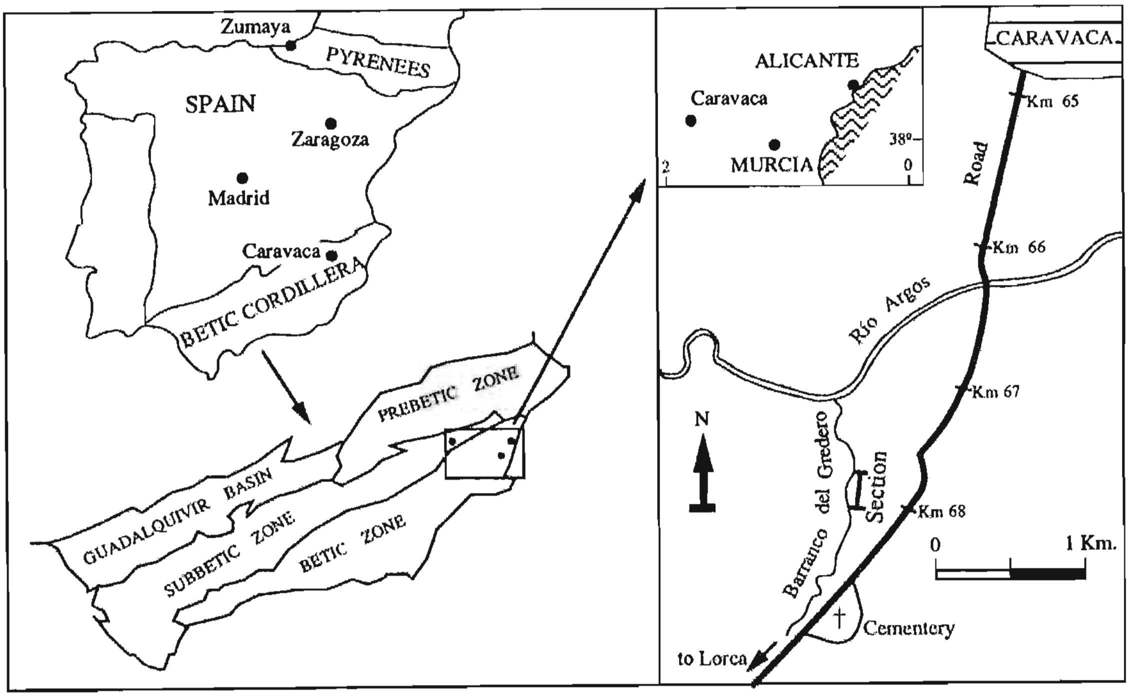

Fig. 5. Geographical and geological location of the Caravaca section

PALEOCENE STAGE BOUNDARIES

In eastern Spain, latest Senonian-Paleocene time was characterized by a wide-spread regression, which culminate in the emergence of large parts of the carbonate platform (De Ruig, *et al.*, 1991). The platform emergence was accompanied during the Paleocene by occasional sandy turbidites in the bathial environments such as Caravaca.

The Danian sediment consists of marls, clay, and marly limestone with abundant pelagic fauna, and few turbiditic strata. About 100 metres above the K/T boundary in the Selandian, pelagic sedimentation is interupted by a thick turbidite strata, containing glauconite-encrusted debris and larger foraminifera (*Ranikothalia* sp., *Miscellanea* sp., *Daviesina* sp., etc.) of Paleocene shallow water limestone.

The Danian-Selandian interval was first studied by Hillebrandt (1974) and by Romein (1979) that established the planktic foraminiferal and calcareous nannofossil biozonations respectively. Smit (1979) studied the K/T boundary and Smit (1982) accomplished the magnetostratigraphy of the Danian at Caravaca section (Fig. 6). The correlation of the Danian nannofossil zones established by Romein (1979) does not mach with the standard correlation of Berggren el al. (1995). The planktic foraminiferal biostratigraphy and the correlation with all the other data have been recently studied by Arenillas (umpubl. Ph. D. thesis) (Fig. 7 and 8)

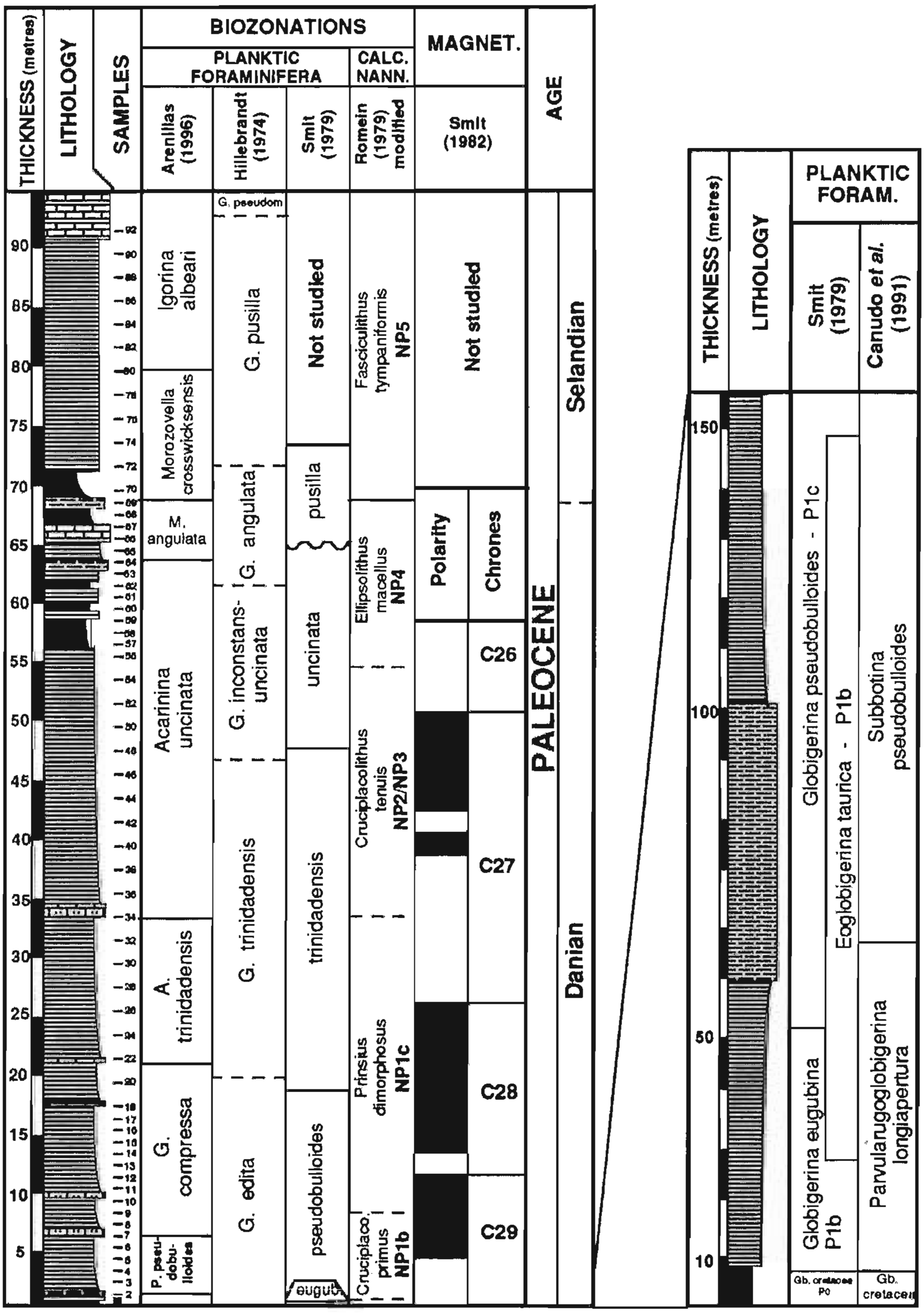

Fig. 6.- Integrated stratigraphy of the Danian and Selandian at Caravaca section (modified from Arenillas, umpubl. Ph. D. thesis)

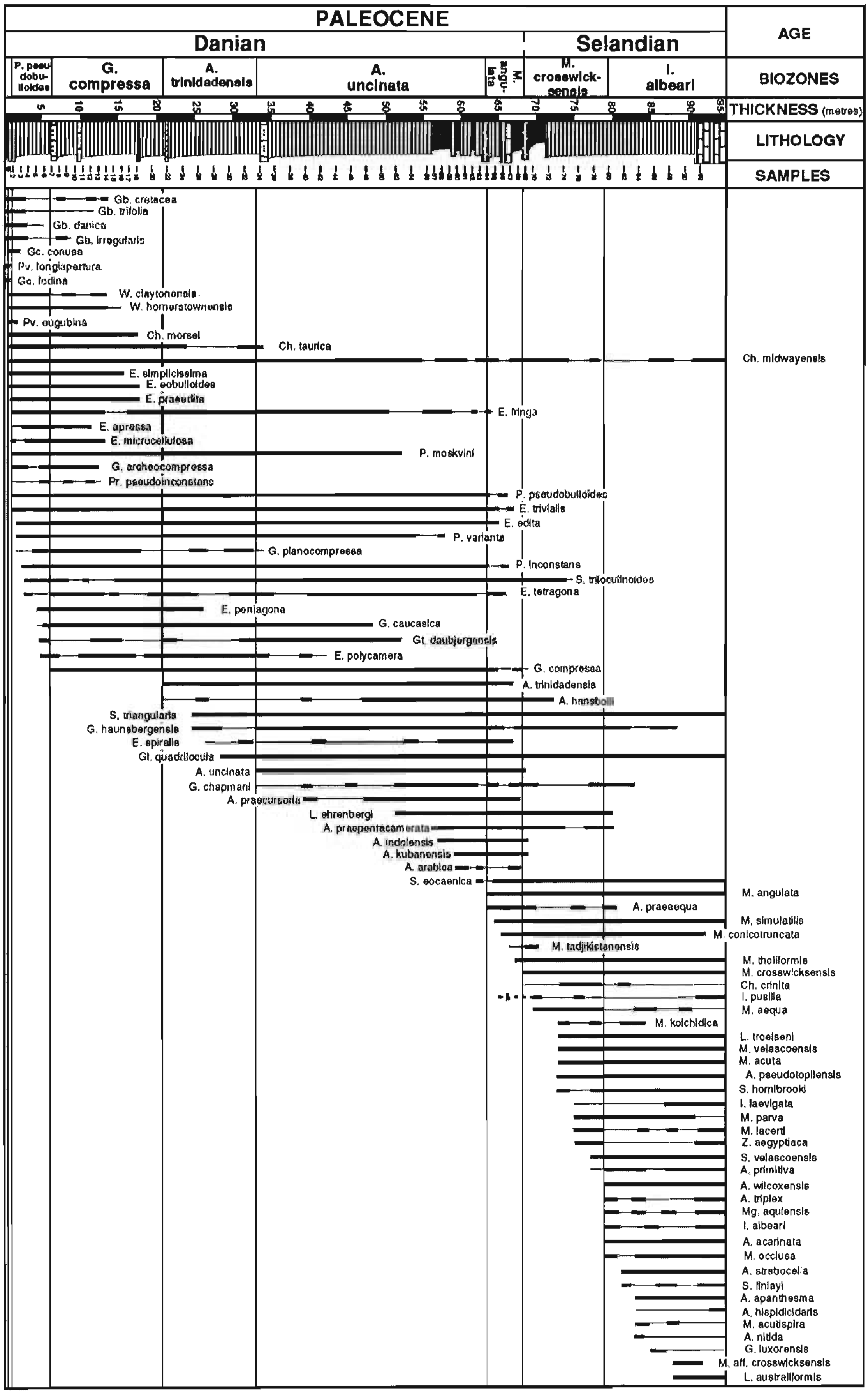

Fig. 7.- Planktic foraminifera stratigraphical distribution of the Danian and Selandian at Caravaca section (modified from Arenillas, umpubl. Ph. D. Thesis).

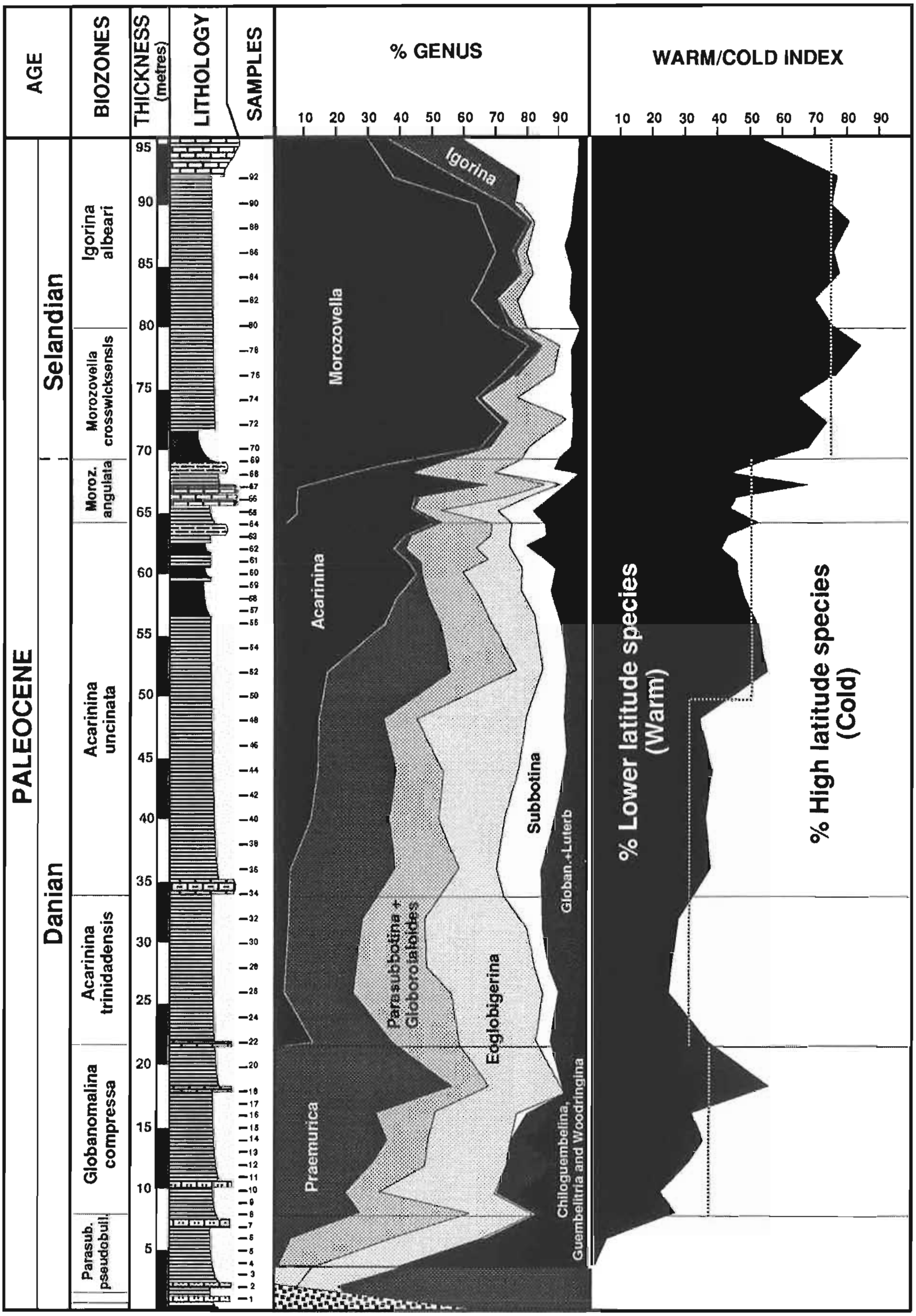

Fig. 8.-Planktic foraminifera quantitative analysis of the Danian and Selandian at Caravaca section (modified from Arenillas, umpubl. Ph. D. Thesis)

PALEOCENE/EOCENE BOUNDARY

High resolution sample analysis across the Paleocene-Eocene transition at Caravaca based on integrated microfossil and geochemical data allow to establish a high resolution stratigraphy. The study revealed a series of events (extinctions, originations and geochemical shifts) that allowed the establishment of a detailed chronostratigraphical framework. The text that follows was published by Molina, Canudo, Martinez and Ortiz (1994) in *Eclogae geologicae Helvetiae* and have been modified selecting and up-dating the most important aspects for the field trip guide.

Material and methods

The Paleocene-Eocene transition is within the Jorquera Formation (Van Veen, 1969), which consists of 225 m of marls interbedded with sandy limestones. The Paleocene part is predominantly marly whereas the Lower Eocene is calcarenitic with interlayers of marls and clays. Several stratigraphic horizons containing larger foraminifers (nummulitids, alveolinids, discocyclinids) are present in the calcarenitic sediments. These lithified strata frequently show cross bedding, convolute lamination and other sedimentological structures, indicating a high energy environment, with possible transport from an inner platform area.

The Paleocene-Eocene transition examined stratigraphically spans 80 m of sediments, of these 40 m spanning the P/E boundary were analyzed in closely spaced samples. The Upper Paleocene is composed of gray marls with two thin interbeds of calcarenite. The P/E boundary interval consists of 20 m of gray marls and clays with the clay interval containing two intervals of strong carbonate dissolution. At the base of the clay layer the sediment colour is dark grey to black and benthic taxa suggest an anoxic event. The Lower Eocene is composed primarily of yellow-gray limestones and sandy limestones intercalated with marls. The top of the section is predominantly marly . The marly intervals contain rich calcareous microfossil assemblages. Planktic foraminifera (Fig. 9) and calcareous nannoplankton are abundant and well preserved. Small benthic foraminifera (Fig. 10) are relatively few to common throughout the section, whereas larger foraminifera are present only in certain strata. Ostracods and siliceous microfossils are rare.

The Upper Paleocene to Lower Eocene interval was sampled at 30 cm to 50 cm with closer sample spacing of 10 cm to 20 cm intervals across the P-E transition (Zones P4-P6b). For foraminiferal isotope and faunal analyses, samples were disaggregated in tap water and then washed through a 63 µm sieve and dried at 50ºC. Isotopic analyses were conducted on the benthic foraminifera *Nuttallides truempyi* except for the $\partial^{13}C$ shift interval where *N.truempyi* is absent and *Lenticulina* spp. was analyzed. Specimens were picked from the 180-250 µm size fraction, ultrasonically cleaned to remove sediment infilling of chambers and roasted under vacuum at 380ºC. Isotopic measurements were conducted with a Finnigan MAT 251 linked to a Kiel carbonate extraction system at the stable isotope laboratory of the University of Michigan, Ann Arbor. Analytical error was measured at 0.05% for $\partial^{13}C$. The data are reported in values referenced to PDB. No species correction factors were applied to the data because species offset studies are still in flux.

The bulk sample mineralogy was studied by X-ray Diffraction at the Department of Mineralogy and Petrology of the University of Granada (Spain).

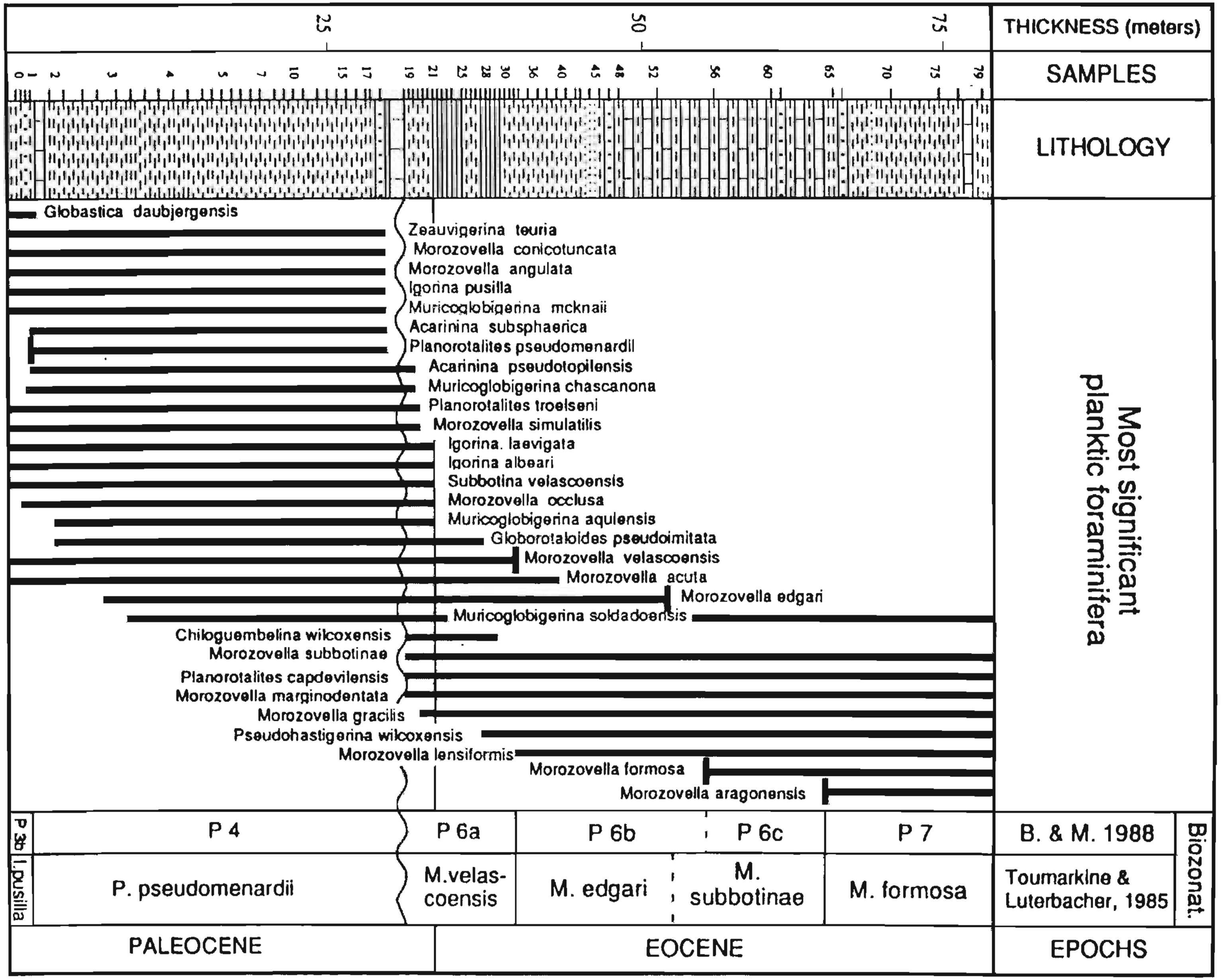

Fig. 9. Stratigraphical distribution of the most significant planktic foraminifera throughout the Paleocene-Eocene transition at Caravaca section. B. & M. 1988 means Berggren & Miller, 1988.

Geochemical analyses were carried out at the XRAL Laboratories in Ontario (Canada) by X-ray Fluorescence, Inductively Coupled Plasma and Neutron Activation. Elements data have been normalized on a carbonate-free basis (Fig. 11 and 12).

Biostratigraphy

Planktic foraminifera are very abundant in the marly intervals and were studied by Von Hillebrandt (1974), establishing a biozonation from the uppermost Cretaceous to the Middle Eocene, but he did not give any range chart. Recently, a very detailed quantitative study has been accomplished (Canudo et al., 1995), evaluating the planktic foraminiferal turnover across the Paleocene-Eocene transition. The following biozones have been recognized: *Igorina pusilla, Planorotalites pseudomenardii, Morozovella velascoensis, Morozovella edgari, Morozovella subbotinae* and *Morozovella formosa*, according to the Toumarkine & Luterbacher (1985) biozonation (Fig. 9). These biozones correspond to P3b, P4, P6a, P6b, P6c and P7, according to the Berggren & Miller (1988) biozonation. The biozone P5 and the lower part of P6a were not identified and the reason is the presence of a hiatus, as it is shown by the simultaneous disappearance of seven species (*Zeauvigerina teuria, Morozovella conicotruncata, Morozovella angulata, I. pusilla, Morozovella mcknaii, Acarinina subsphaerica* and *P. pseudomenardii*) and the subsequent appearance of four species (*M. subbotinae, Chiloguembelina wilcoxensis, Morozovella marginodentata,* and *Planorotalites capdevilensis*). This hiatus is placed 2.5 m below the dissolution interval that marks the Paleocene/Eocene boundary. Nine species (*Acarinina pseudotopilensis, Muricoglobigerina chascanona, Planorotalites troelseni, Morozovella simulatilis, Igorina laevigata, Igorina albeari, Subbotina velascoensis, Morozovella occlusa* and *Muricoglobigerina aquiensis*) disappear gradually at or below the boundary indicating that the planktic foraminifera were affected by the event but not so much as the small benthic foraminifera. Five species (*I. laevigata, I. albeari, S.velascoensis, M. occlusa* and *M. aquiensis*) seem to become extinct, but their distribution compared with other sections (Molina et al., 1992; Canudo *et al.*, 1995) shows that in Caravaca *I. albeary* and *M. occlusa* constitute local disappearances. Another species (*M. aquiensis)* also becomes extinct later in the area of the Ypresian stratotype in Belgium (Pardo *et al.*, 1994). Consequently, only *Igorina laevigata* and *Subbotina velascoensis* become extinct at the P/E boundary.

The small benthic foraminifera have been recently studied quantitatively (Ortiz, 1993, 1994, 1995) for the P-E transition evaluating the faunal turnover across the P/E boundary. In this paper the following biozones have been recognized: *Pyramidina rudita* , *Stensioina beccariiformis, Haplophragmoides retrosepta, Bulimina tuxpamensis/Tappanina selmensis, Nuttallides truempyi,* and *Cibicidoides subspiratus* (Fig. 10). This new biozonation is correlated with the Berggren & Miller (1989) biozonation. At the top of *S. beccariiformis* Biozone that coincides with the BB1/BB2 boundary, 50% of the bathyal and abyssal species disappeared and this can be considered a mass extinction; among them the more significant are: *S. beccariiformis, Anomalinoides rubiginosus, Cibicidoides velascoensis, Gyroidinoides globosus, Neoflabellina semireticulata, Osangularia velascoensis, Dorothia retusa, Eponides megastoma, Pullenia coryelli* and *Tritaxia globulifera.* The *H. retrosepta* Biozone is very poor in microfossils probably because of the dissolution. The uppermost part of the section can be dated as BB3 Biozone because an assemblage that characterized

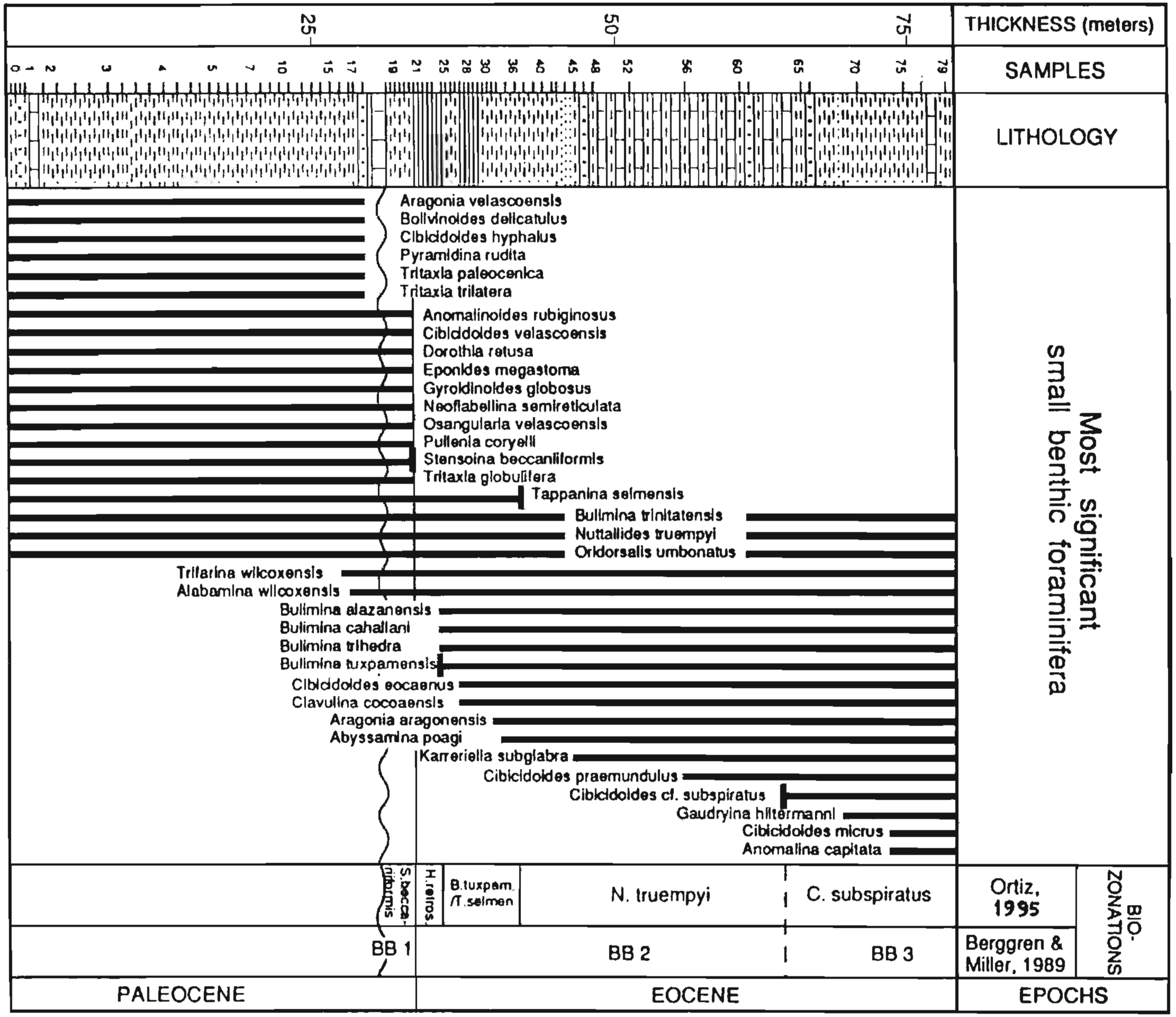

Fig. **10.** Stratigraphical distribution of the most significant small benthic foraminifera throughout the Paleocene-Eocene transition at Caravaca section.

this biozone (*Gaudryina hiltermanni, Cibicidoides micrus* and *Anomalina capitata*) and *Cibicidoides* cf. *subspiratus* were found. According to the small benthic foraminifera the depositional depth appears to have been upper bathyal to upper middle bathyal depth (P4 to P6b about 600 m and P6b to P7 about 300 m), as it is suggested by a buliminid dominated fauna including *N. truempyi, Bulimina cahallani, Bulimina trinitatensis and Anomalinoides capitatus.*

The larger foraminifera were studied by Von Hillebrandt (1974) who found four strata containing nummulitids, alveolinids and discocyclinids. In the lower level, placed at the boundary between P4 and P6a Biozones, an assemblage was identified that characterizes the *Operculina azilensis* Biozone. The second level contains an assemblage that characterizes the *Nummulites praecursor* Biozone and the *Alveolina moussoulensis* Biozone and the third level the *Nummulites involutus* Biozone and the *Alveolina trempina* Biozone. The fourth level was found in the P7 Biozone and contains an assemblage of the *Nummulites planulatus* Biozone. The biozonations established by Hillebrandt (1974), that were based on the alveolinids biozonation of Hottinger (1960) and on the nummulitids biozonation of Schaub (1951), have been modified according to our revision (Fig. 13). Our most detailed sampling allows to recognize an older new level, which is placed in the top of P3b, containing very rare and primitive nummulitids that according to Serra-Kiel (personal communication), are: *Ranikotalia soldadensis, Miscellanea* sp. and *Daviesina* sp. This assemblage has never been found before in the Spanish Betic Cordillera and they were previosly found mainly in the Paleocene of the French Guiana.

The calcareous nannoplankton was studied by Romein (1979) that established the vertical distribution of the Paleocene and Early Eocene species, defining a detailed biozonation. For the Paleocene-Eocene transition interval the following biozones were recognized: *Fasciculithus tympaniformis, Heliolithus kleinpellii, Discoaster mohleri, Discoaster multiradiatus, Tribrachiatus contortus, Discoaster binodosus* and *Tribrachiatus orthostylus* (Fig. 13). These biozones correspond to NP5, NP6, NP7, NP8, NP9, NP10, NP11 and NP12 of the biozonation of Martini (1971). According to Romein (1979) all samples, except those from the upper part of the *D. multiradiatus* Biozone and the lower part of the *T. contortus* Biozone contained rich nannofossil floras. *T. contortus* did not occur in the Early Eocene floras from this sequence and the boundary between the *T. contortus* Biozone and the *D. binodosus* Biozone was arbitrarily drawn between the appearance of *T. orthostylus* and the entry of *Discoaster barbadiensis.*

Geochemistry

Some aspects of the carbon and oxygen isotopes have been previously discussed by Canudo et al., (in press) correlating the planktic and benthic foraminiferal turnovers and the $\partial^{13}C$ and $\partial^{18}O$ isotopes across the Paleocene-Eocene transition at Caravaca and Zumaya, representing the first isotopic records from the western Tethys Seaway. The isotopic data are integrated in this paper with the bulk mineralogy (Fig. 11), the trace and major elements and the biostratigraphy, in order to clearly characterized the P/E boundary event.

The $\partial^{13}C$ isotope measured on *N. truempyi* shows a relatively constant positive value through P4 Biozone, indicating very stable conditions at the Caravaca section. Nevertheless, just above the hiatus in P6a Biozone there is a small excursion into negative values that coincides with deposition of an anoxic dark grey shale. The biggest shift occurred in the upper part of P6a Biozone

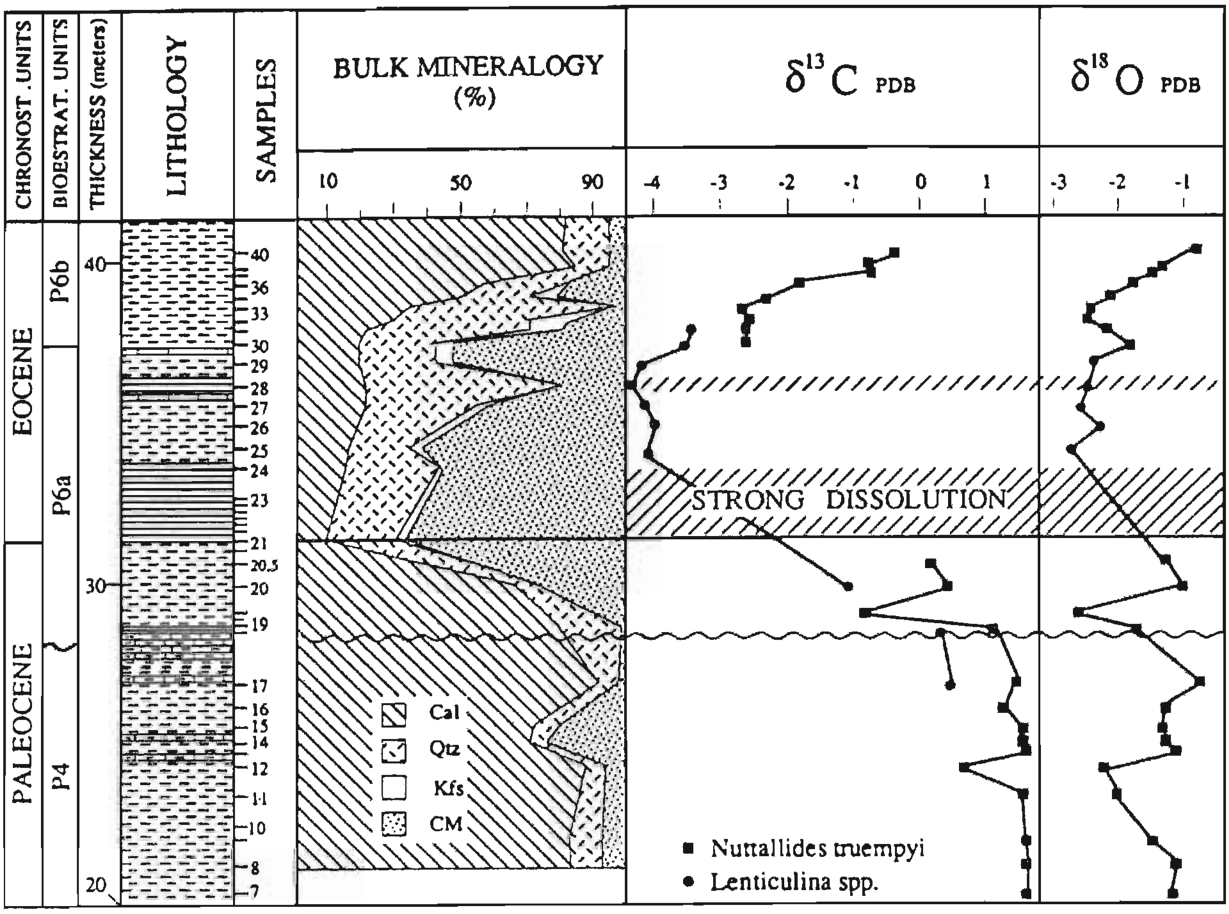

Fig.11. $\partial^{13}C$ and $\partial^{18}O$ isotopes and bulk sample mineralogy across the Paleocene/Eocene boundary at Caravaca section.

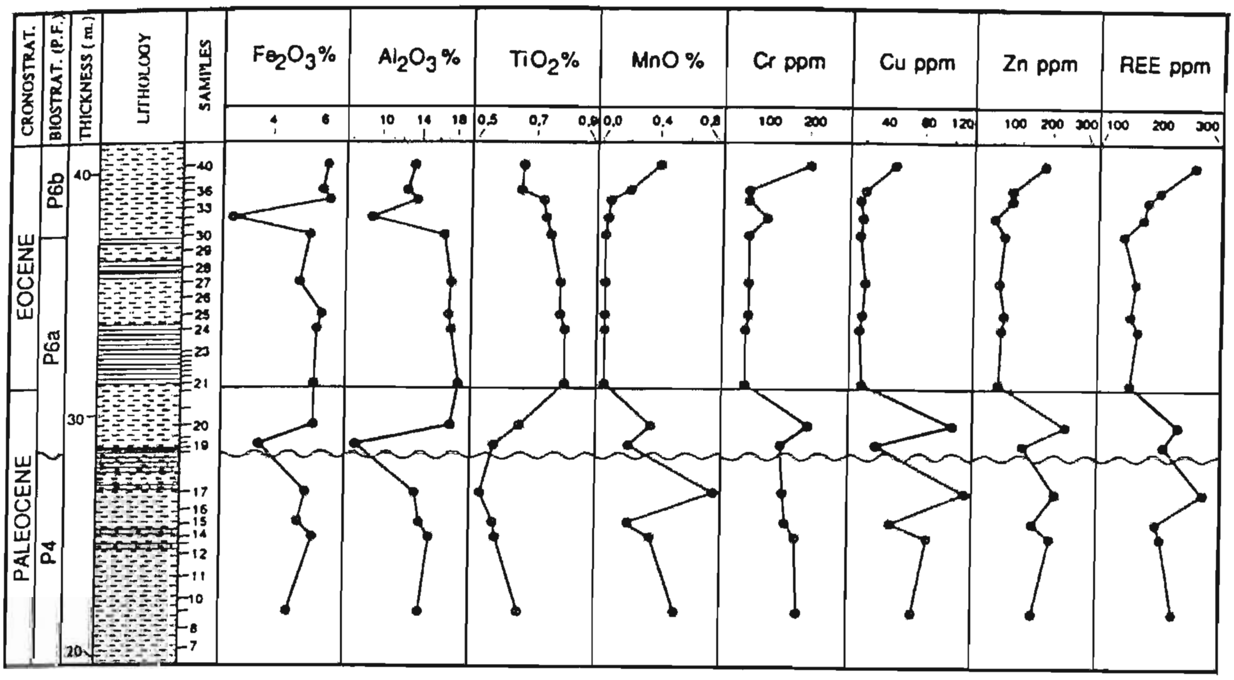

Fig.12. Geochemical analyses across the Paleocene/Eocene boundary at Caravaca section.

where there is a rapid decrease of 4 permil in the benthic *Lenticulina* values. This excursion coincides with the sudden extinction of 50% of the small benthic foraminifera and with a clay layer where most calcareous foraminifera are dissolved. At the top of P6a Biozone there is a second dissolution interval where *Lenticulina* values are stable around -4 permil and in the lower part of P6b Biozone *N. truempyi* gradually increase by 2.5 permil (Fig. 11).

The $\partial^{18}O$ isotope has also been measured on *N. truempyi* and where it was absent *Lenticulina* was analysed. In P4 Biozone there is a small shift that coincides with another in $\partial^{13}C$ and just above the hiatus another small shift is observed. Furthermore, at the transition between the P6a and P6b biozones a similar excursion, not so strong as in $\partial^{13}C$, was observed.

The mineralogy of the section consists mainly of calcite, phyllosilicates, feldspar and quartz. Above the Paleocene/Eocene boundary we have detected a significant decrease in carbonate content. This is also evident by the dissolution of most of the calcareous foraminifera, but in the middle part of this interval some foraminifers were preserved. The carbonate decrease is accompanied by an important increase in the quartz content that reaches high proportions in the Lower Eocene. Different elements were analyzed across the P/E boundary and the most significant changes are observed in Fe_2O_3, Al_2O_3, TiO_2, MnO, Cr, Cu, Zn and REE. In the interval above the P/E boundary we have detected higher contents in Fe_2O_3, Al_2O_3 and TiO_2, and lower contents in MnO, Cr, Cu, Zn and REE.

Sequence Stratigraphy

As the depositional depth of Caravaca appears to have been of upper bathyal to upper middle bathyal (intervals with small benthic foraminifera) and outer neritic (levels with larger foraminifera) depth (200-600 m.), the sedimentation reflects the sea level changes quite clearly. Based in the study of this section, as well as in regional observations, several depositional sequences have been recognized (Fig. 13). The highstands are represented by shallow marine sandstones containing larger foraminifera and the lowstands and trangressive-systems tracts by mudstones containing planktic microfossils. The marly intervals are thicker than the calcarenitic ones indicating deposition mainly on an outer platform.

The boundary between our depositional sequence 1 and 2 is clearly placed in the lower part of P 4 biozone just above a calcarenitic short interval, including larger foraminifera, where evidence of reworking exists. Our depositional sequence 2 seems to correlate with the TA2.1 of Haq et al. (1987) and clearly corresponds to the Thanetian stage. The hiatus at the base of depositional sequence 3 represents about 0.7 Ma missing the Lower Ilerdian, the TA2.2 and possibly the TA2.3 sequences of Haq et al. (1987). Our depositional sequence 3 could correspond to TA2.3 and mainly to TA2.4 and constitutes a very high sea level rise that coincides with the anoxic event which marks the P/E boundary. The rest of the sequences 4, 5 and 6 are not so clearly defined, due to the interference between the global eustatic changes and the local tectonic, but could correlate with TA2.5, TA2.6 and TA2.7 respectively. This correlation is tentative and provisional since the global depositional sequences for this time interval are under revision (Hardenbol, personal communication).

Discussion

Although the Caravaca section was studied biostratigraphycally by several authors (Durand Delgá & Magné, 1958; Von Hillebrandt, 1974; Romein 1979) none of them found the hiatus that has been recently discovered by means of planktic foraminifera (Canudo et al., in press) and it is also documented in this paper by the simultaneous disappearance of at least six small benthic foraminifera (*Aragonia velascoensis, Bolivinoides delicatulus, Cibicidoides hyphalus, Pyramidina rudita, Tritaxia paleocenica* and *Tritaxia trilatera*). This hiatus comprises from the uppermost part of P4 Biozone to the upper part of P6a Biozone and according to the calibration of the Berggren & Miller (1988) biozonation the hiatus could represent about 0.7 Ma, since the top of P4 Biozone was dated at 58.64 Ma and the ∂^{13}C excursion at 58.0 Ma by Pak & Miller (1992) at Site 577 where good chronostratigraphic control is available.

The hiatus is located 2.5 m below the small benthic foraminifera extinction and just above a calcarenitic strata containing nummulitids of the *O. azilensis* Biozone (Late Thanetian) (Fig. 13). About 17 m above Von Hillebrandt (1974) found another level with nummulitids of the *N. praecursor* Biozone (Middle Ilerdian) and he attributed the marly clay interbedded interval to the *N. fraasi* Biozone (Early Ilerdian). Nevertheless, because of the hiatus this biozone could not be represented and neither the alveolinids *A. cucumiformis* and *A. ellipsoidalis* Biozones (Early Ilerdian). When this section is compared with the Zumaya section (Canudo & Molina, 1992a) the 2.5 m between the P/E boundary and the base of the hiatus at the P4 Biozone in Caravaca correspond to about 19 m at Zumaya where the P/E transition is very well represented in continuous marine facies.

The 2 m interval of clay with some dissolution resistant foraminifers between the two strong dissolution levels in Caravaca is represented by about 13 m at Zumaya (Canudo *et al.*, 1995), but at this interval no evidence of a hiatus exists, consequently the sedimentation has to be condensed at this interval at Caravaca where the ∂^{13}C excursion maintains the minimum values of -4 permil. At the base of the clay with dissolution the smaller benthic foraminifera suffer a dramatic extinction in coincidence with the carbon and oxygen isotopes shifts. Consequently, although condensed in comparison with Zumaya the P/E boundary event is well represented at Caravaca, which is an excellent expanded section compared with any deep-sea record known so far.

The mass extinction horizon of small benthic bathyal and abyssal foraminifera appears to be the more suitable event to mark the Paleocene/Eocene boundary, since it coincides with the ∂^{13}C excursion, which has been determined to be globally synchronous (Thomas, 1990; Kaiho, 1991; Kennett & Stott, 1991; Pak & Miller, 1992; Lu & Keller, 1993). Nevertheless, the placement of the P/E boundary has not been officially defined and many biostratigraphers still place this boundary at the extinction of *M. velascoensis* that marks the P6a/P6b boundary in accordance with Berggren *et al.* (1985). Some others place the P/E boundary at the extinction of *P. pseudomenardii* that defines the top of P4 Zone in accordance with Cavelier and Pomerol (1986). The *M. velascoensis* extiction is supposed to coincide with the *Pseudohastigerina wilcoxensis* appearance, but in Caravaca both species overlap. In contrast, in the Pyrenean sections (Zumaya and Campo) and in the area of the Ypresian stratotype (Knokke borehole) *P. wilcoxensis* appears later (Molina et al., 1992; Canudo & Molina 1992a,b; Pardo *et al.*, 1994). Consequently, these horizons are not isochronous, especially the *Pseudohastigerina* "datum", and are less reliable than the benthic extinction horizon. Furthermore, the base of the Ypresian stratotype appears to be closer to the mass extinction (about 50%) of the small benthic foraminifera than to the *M. velascoensis* extinction (Pardo et al., 1994). According to planktic foraminifera and the Spanish sections, the markers to identify the P/E boundary considered in this paper are the simultaneous extinctions of *Igorina laevigata* and *Subbotina velascoensis*. In addition, the $\partial^{13}C$ event is associated with a major planktic foraminiferal turnover marked by the gradual extinction and evolution of 33% and 18% of the species respectively.

Stable isotope $\partial^{13}C$ indicates that during the Paleocene-Eocene transition, which coincides with the small benthic foraminifera mass extinction, temperature increased reaching values comparable to the Late Cretaceous. This event has been previously documented in other regions mainly by Shackleton (1986),Thomas (1990), and Kennett & Stott (1991). In the Caravaca section we have found two slight shift below and above the hiatus and a strong shift coinciding with the small benthic extinction at the base of a clay interval with a strong carbonate dissolution. The negative $\partial^{13}C$ excursion can be dated very precisely at Caravaca in the middle P6a Biozone persisting up to the P6a/P6b boundary and declining at the lower part of the P6b Biozone, just above a second strong dissolution interval. The $\partial^{13}C$ shows a similar contemporary trend indicating an increase in temperature. These can be considered high resolution records since the entire shift and recovery interval is represented by 8 m at Caravaca compared with less than 20 cm at DSDP Site 577 (Pak & Miller, 1992). Even so, this interval is less expanded than in the Zumaya section (Pyrenees, Northern Spain) where it is represented by 18 m (Canudo *et al.*, 1995).

In the geochemically studied Paleocene/Eocene transition interval, a significant change in the concentration of different elements suggests a change in palaeoceanographic conditions. Above the P/E boundary the increase in Fe_2O_3, Al_2O_3 and TiO_2 (Fig. 12), suggests an important increase in the detrital accumulation around this boundary, which is also indicated by the increase in the quartz content (Fig. 11). Although hydrothermal activity is a potential source of Fe, detrital input plays a more important role in the supply of Fe to the ocean sediments. In addition, typical continent derived-elements such as Al and Ti also increase their concentration. Statistical analyses of the obtained data reveal a good correlation among Fe-Al-Ti (r=0.96 for 16 samples). These data suggest a common terrestrial source for these elements.

The sharp decrease in MnO content across the boundary could indicate a change in the paleoenviromental conditions. The main source for this element is volcano-hydrothermal activity (Klinkhammer, 1980; Andrianiazy & Renard, 1984). In this sense, changes in the tectonic regime of mid-ocean ridges have been documented at the P/E boundary (Aubry et al. 1988; White & MacKenzie, 1984) and as a consequence, an increase in the hydrothermal activity (Olivarez & Owen, 1989). Therefore an enrichment in Mn can be expected in the area of the oceanic ridges due to hydrothermal input (Lyle, 1976). However the influence of oceanic ridges as source of Mn is less important in continental margin environments, such as at Caravaca, where a decrease is observed across the boundary and not the expected increase as a consequence of hydrothermal activity. Redox conditions seem to be the main control of Mn concentrations. The decrease in Mn across the boundary suggests a significant change to reductive conditions in which Mn decreases because it would be essentially soluble.

Other elements, such as Cr, Cu, Zn or REE, are usually associated with the clay fraction, in addition organic material also plays an important role in the Cu and Zn concentrations. In reductive conditions, similar to those of the interval above the P/E boundary (Irwin et al., 1977; Schmitz, 1985; Schmitz et al., 1988), an increase in these elements could be expected; however the significant increase in siliciclastic input could have masked such enrichment. The decrease in REE also indicated siliciclastic accumulation since higher size fractions than silt present less REE concentration due to the dilution effect of quartz (sea Cullers et al., 1988).

In summary, trace minerals and major elements reveal an important increase in the detrital input as a probable consequence of the tectonic instability across the P/E transition, as well as a change across this boundary to more reductive conditions.

Regarding the sequence stratigraphy, the P/E boundary approximately coincides with a sea level change (Haq et al., 1987). The P/E boundary event placed in the lower part of our depositional sequence 3 seem to be related to a strong sea level rise. According to Leinfelder & Seyfried, 1993 during very long-lasting greenhouse episodes the sea level is very high, climate and circulation systems are stable and biotic crises often develop as a consequence of oxygen depletion. The P/E boundary event is mainly characterized by the extinction of the small bathyal and abyssal foraminifera in coincidence with a strong rise in temperature and an anoxic interval, as it is evident at Caravaca. Consequently, the interrelated rise in sea level and temperature, together with a stabilisation in oceanic deep water circulation, because of the temporal cessation of polar cool water supply that caused a sharp deep-sea warming at the P/E boundary, could produce oxygen deplection in the deep sea, causing partial collapse of the benthic bathyal and abyssal marine ecosystems.

Conclusions

The high resolution study and the integrated approach allow us to establish a detailed stratigraphy of the Paleocene/Eocene transition at Caravaca. The chronostratigraphical study spans from the Middle Selandian to the Middle Ypresian. A short hiatus has been detected between the P4 and P6a biozones and is evaluated at about 0.7 Ma. missing the Early Ilerdian (nummulitid *N. fraasi* Biozone and alveolinids *A. cucumiformis* and *A. ellipsoidalis* Biozones). Above the hiatus we have detected significant changes in the concentration of several elements (TiO_2, MnO, Cr, Cu, Zn and REE), coinciding with a carbonate dissolution interval, a dark anoxic level, a major increase in the quartz content and excursions in $\partial^{13}C$ and $\partial^{18}O$. A high increase in temperature, a very high sea level rise, volcanism and hydrothermal activity could have caused the physical and chemical changes that caused the biotic effects observed in different groups of foraminifers. The geochemical and mineralogical shifts coincide with a mass extinction in the small benthic bathyal foraminifera at a horizon that is very reliable to place the P/E boundary because of its possibilities of global correlation and its proximity to the base of the Ypresian stratotype. The Paleocene/Eocene boundary falls whitin the middle part of the P6a (*M. velascoensis*) Biozone, the upper part of NP9 (*D. multiradiatus*) Biozone, the top of BB1 (*S. beccariiformis*) Biozone, the base of *N. praecursor* and *A. moussoulensis* Biozones. In conclusion, the Caravaca section is not optimal as potential Global Stratotype Section and Point for the Paleocene/Eocene boundary, mainly because of the hiatus near the boundary event, but is a very good reference section and it allows the integration of geochemistry and biostratigraphy based on different groups of microfossils.

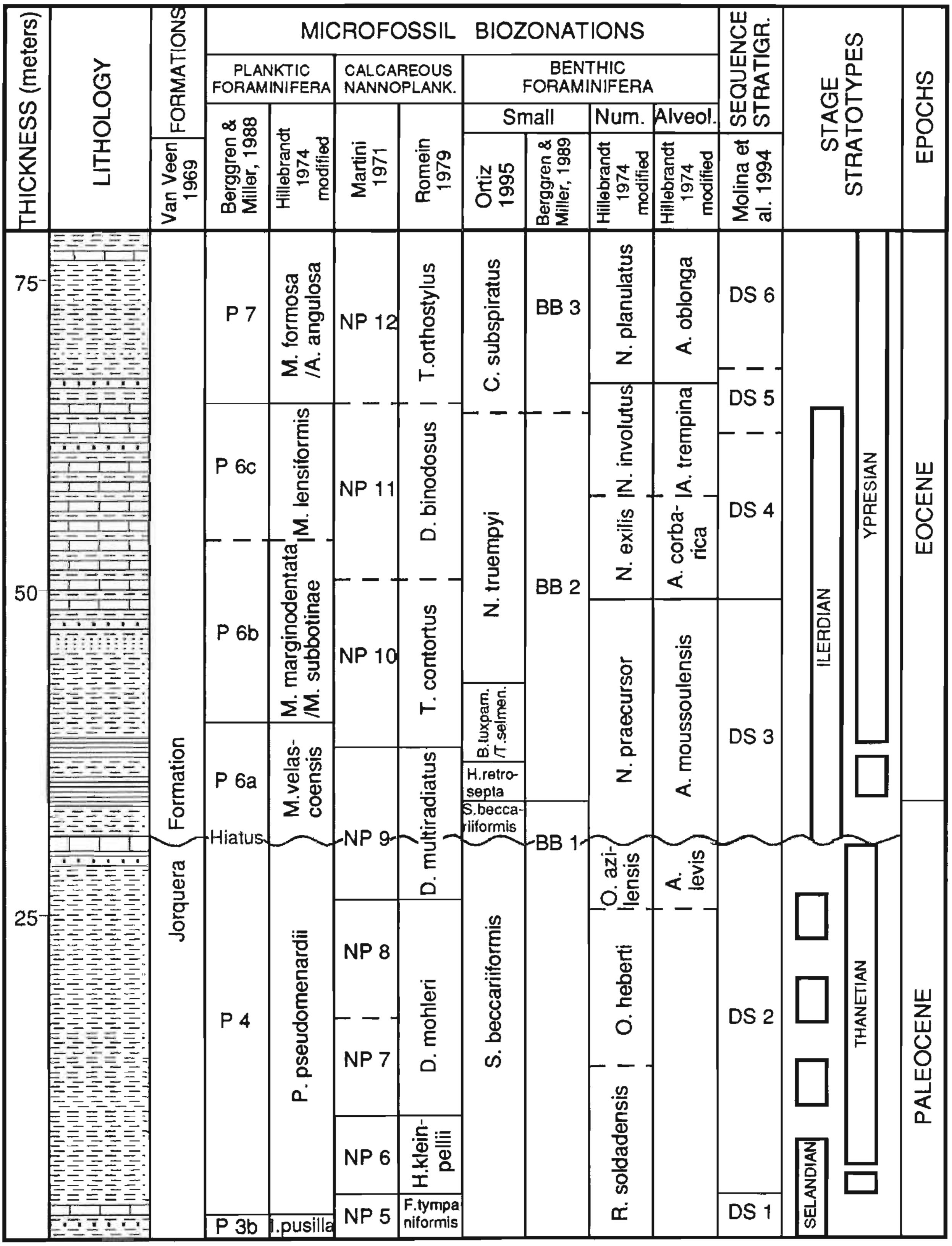

Fig. 13. - Integrated stratigraphy throughout the Paleocene-Eocene transition at Caravaca section (modified form Molina *et al.*, 1994).

ALAMEDILLA SECTION

The Alamedilla section is located in the Barranco de Valencianos about 2 kilometres at the South of Alamedilla town in the north of Granada province in southern Spain (Fig. 14). Geologically, the section belongs to the central Subbetic zone of the Betic Cordillera. The pelagic sedimentation from the Late Cretaceous to the Early Eocene was characterized by red marls and marly limestones, known as "capas rojas" facies. The Paleocene/Eocene boundary consists of a very prominent dark red clay and 2 meters below a 9 centimetres bioturbated calcarenite, which could be the equivalent to the Zumaya turbidite layer intercalated in the prominent limestone strata below the P/E boundary. Between the calcarenite and the clay an interval of grey marls, very dark in its base, is present.

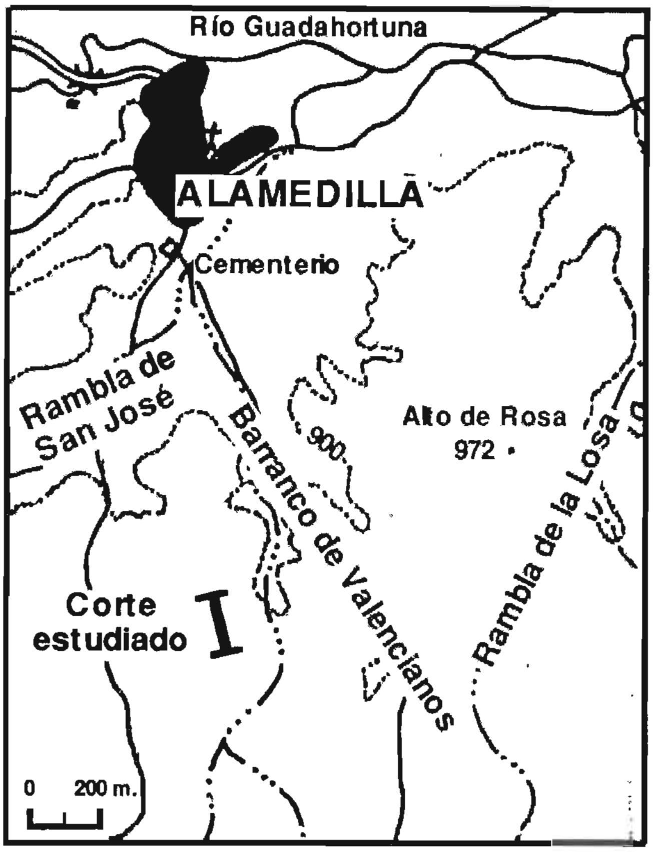

Fig. 14. - Geographical location of the Alamedilla section

THE PALEOCENE /EOCENE BOUNDARY

The sediment accumulation rate at Alamedilla section is several times higher than in the deep-sea. This expanded section provide the opportunity for building a high resolution time scale and examining in detail the regional climatic, oceanographic and biotic changes across the P/E boundary. Two papers dealing with the Paleocene/Eocene boundary at Alamedilla section are in press: Arenillas and Molina (1996) in Revista Española de Paleontología and Lu *et al.* (1996) in Terra Nova.

The planktic foraminiferal biostratigraphy has been established by Arenillas and Molina (1996) and a new biozone was defined: *Igorina laevigata* Biozone. The top of this biozone coincides with the P/E boundary, dividing the classical *M. velascoensis* Biozone into two biozones. (Fig. 15). The quantitative analysis was also accomplished indicating a major increase of tropical compresed acarininids just above the boundary clay (Fig. 16 and 17). The compressed acarininids are composed of three species: *Acarinina africana, Acarinina sibaiyaensis* and *Acarinina berggreni.*

Benthic foraminifera were studied by Ortiz *in* Lu et al. (1996) indicating that deposition across the P/E boundary at Alamedilla occurred in a lower bathyal to abyssal environment. The fauna contains abundant (up to 27%) abyssaminids and many species that have their upper depth limits between 1000 and 2000 metres (e.g., *Abyssamina poagi, Bulimina valascoensis, Aragonia aragonensis* and *Tritaxia paleocenica*). The benthic foraminiferal mass extinction event was located at 13.45 m. as marked by de disappearance of *Stensioina beccariiformis*, among other species. Aproximately 53% of the Late Paleocene species survived into the Early Eocene.

Stable isotope measurements were conducted on *Morozovella subbotinae*, a surface dwelling planktic foraminiferal species. The results demonstrate little change in $\partial^{13}C$ values across the P/E boundary, but a distinct negative excursion in $\partial^{13}C$ values that peaks in the interval of maximum foraminiferal turnover (Fig. 18). The beginning of the rapid excursion in planktic $\partial^{13}C$ values occurred 20 cm below the benthic foraminiferal extinction and 35 cm below the onset of the planktic foraminiferal turnover. A pre-excursion gradual decrease in $\partial^{13}C$ values began well below faunal changes and has not been observed at other continental margin or deep-sea sites outside the Tethys region.

Sediments are composed of calcite, phyllosilicates and detritus and were studied by Adatte *in* Lu *et al.* (1996). Whole-rock compositional analyses show that calcite content averages 75-80% over the lower 10 m of the section. A gradual but fluctuating decrease in calcite content starts at 10.30 me and accelerates at 13.45 m. Coincident with the $\partial^{13}C$ excursion and planktic foraminiferal turnover, calcite content drop to as low as 2% and averages 44% between 13.45 m and 15.00 m. At 15.00 m, calcite content rapidly returns to 75-85% and remains high through the upper portion of the section. Detrital content, measuring the relative abundance of detritital quartz, K-feldspar and plagioclase, averages 2-4% through most of the section except for the $\partial^{13}C$ excursion interval between 13.45 and 15.00 m. Within this interval detrital content increases to a high of 12% and averages 9%. Changes in calcite and detrital contents are compensated by changes in phyllosilicate contents (Fig. 18).

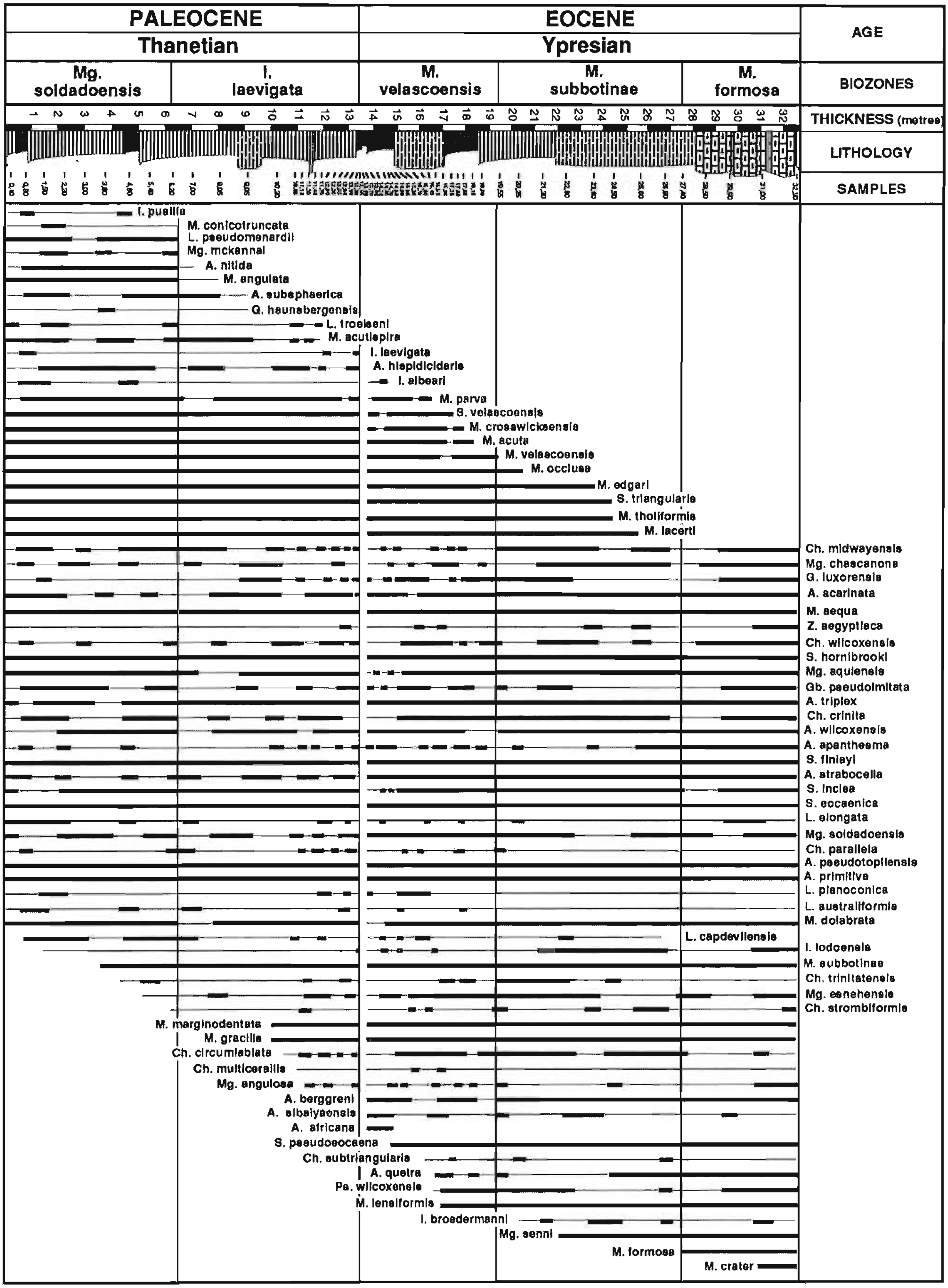

Fig. 15.- Planktic foraminifera stratigraphical distribution of the Paleocene-Eocene transition at Alamedilla section (modified from Arenillas y Molina, 1996)

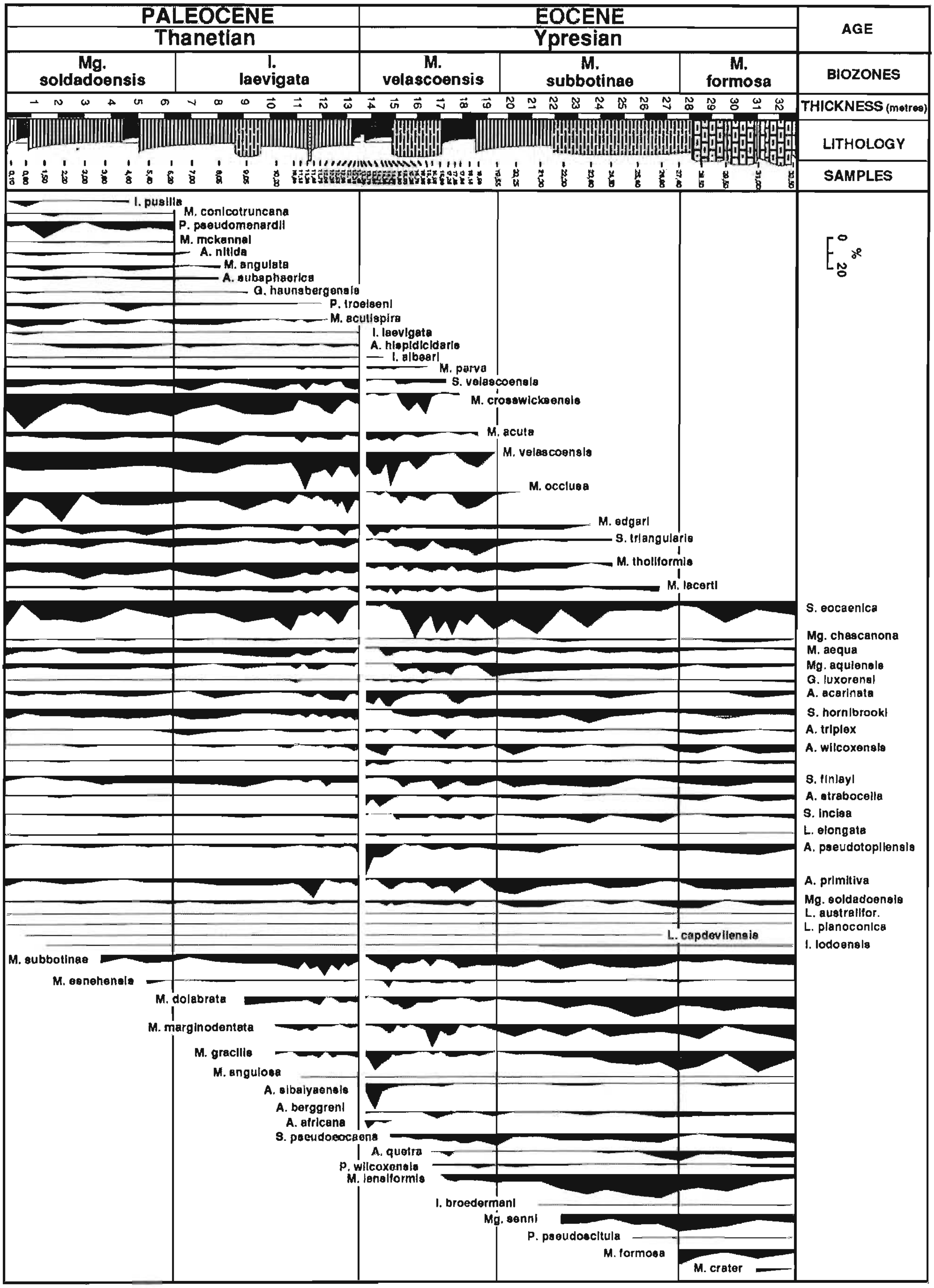

Fig. 16.- Planktic foraminifera relative abundances of the Paleocene-Eocene transition at Alamedilla section (modified from Arenillas y Molina, 1996)

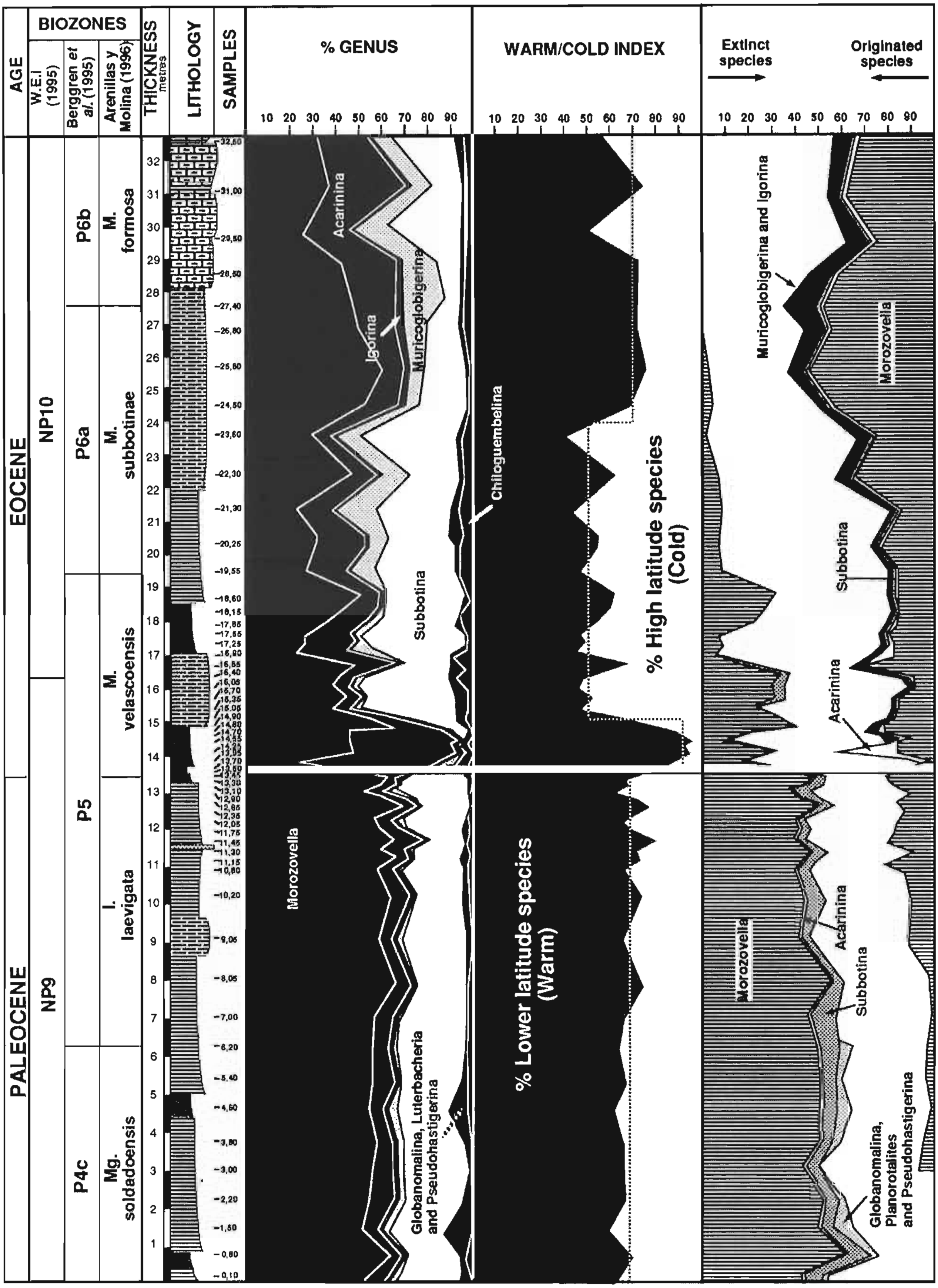

Fig. 17.- Planktic foraminifera quantitative analysis of the Paleocene-Eocene transition at Alamedilla section (modified from Arenillas y Molina, 1996)

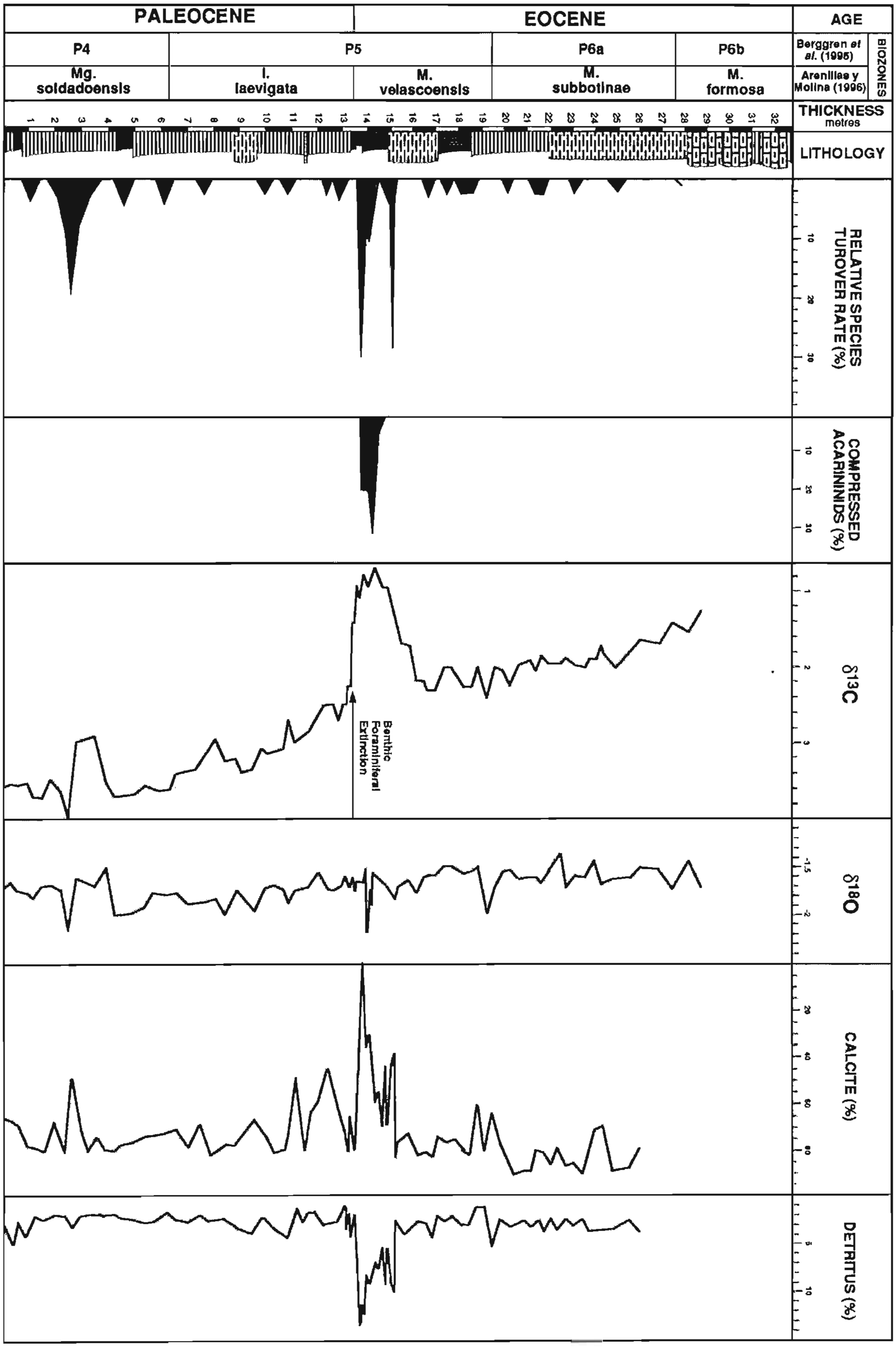

Fig.18.- Faunal, isotopic and sedimentary changes across the Paleocene/Eocene boundary a Alamedilla section (modified from Lu *et al.*, 1996)

THE YPRESIAN/LUTETIAN BOUNDARY

The sediements of the Ypresian-Lutetian transition at Alamedilla section were deposited in a deep basin of the Subbetic zone in the Betic cordillera. The lithology is composed of marls, red and more calcareous at the Early Eocene and green and very argilaceous at the Middle Eocene. The great abundance of planktic foraminifera has allowed to establish the quantitative distribution of species along the Y/L transition. The change in colour of the sediments coincides with the first appearance of *Subbotina boweri, Subbotina frontosa* and *Morozovella dolobrata* at the uppermost part of the Early Eocene. The Y/L boundary is characterized by the *Hantkenina* appearance and the apparently simultaneous first appearance of *M. lehneri, T. rohri* and *T. topilensis.* Consequently, the Y/L transition seems to be less expanded at Alamedilla than at Agost. (Fig. 19).

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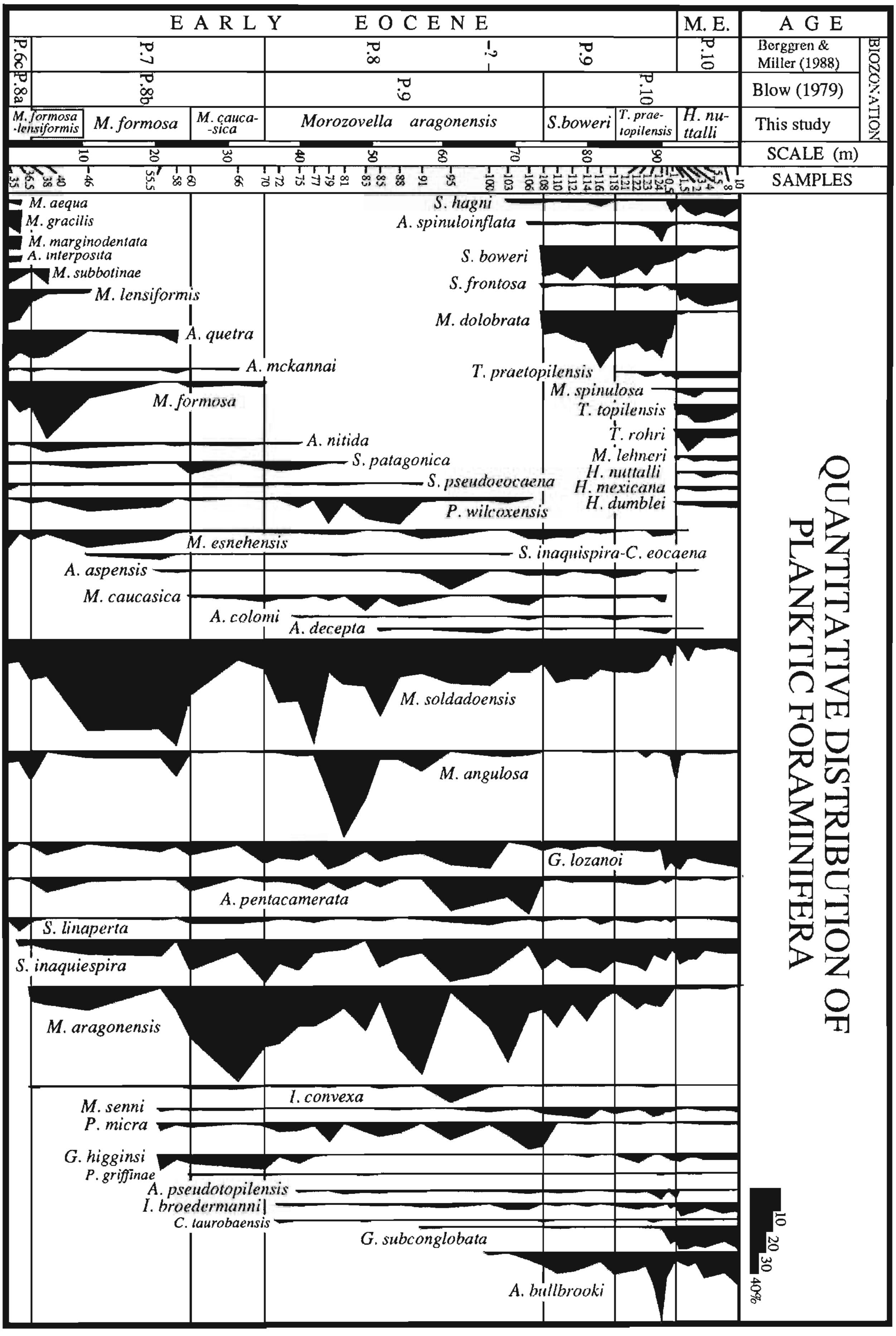

Fig. 19 : Quantitative distribution of Planktic Foraminifera at Alamedilla section (Betic Cordillera, Spain)

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