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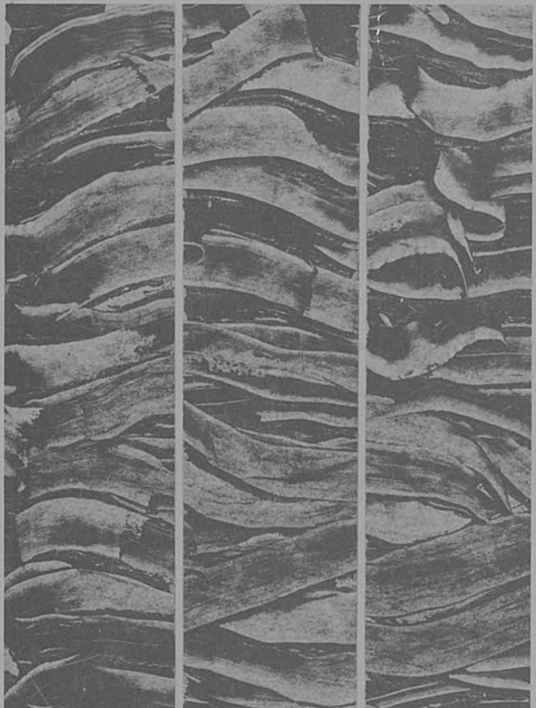
LA LIMITE PALÉOCÈNE – ÉOCÈNE EN EUROPE :
ÉVÉNEMENTS ET CORRÉLATIONS

Séance spécialisée S.G.F. / C.F.S.

Paris, 19 – 20 janvier 1998



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STRATA

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LA LIMITE PALEOCENE - EOCENE EN EUROPE : EVENEMENTS ET CORRELATIONS

Résumés des Communications et Posters

Conception et réalisation
- Jacques Rey, Françoise Ronchini -

La Limite Paléocène/Eocène en Europe : Evénements et Corrélations

The Palaeocene/Eocene Boundary in Europe: Events and Correlations

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Paris, 19 - 20 Janvier 1998

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THE UPPER PALEOCENE-LOWER-EOCENE STRATIGRAPHIC PUZZLE

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Great advance has been achieved over the last decade in documenting the long-term and short-term events that occurred during the late Paleocene and early Eocene epochs. In particular, much attention has been given to the unexpectedly prominent and abrupt events that occurred in the latest Paleocene and which are best epitomized by an unprecedented land mammal turnover (e.g., Gingerich, 1980), the extinction of the abyssal/ bathyal *Stensioina beccariiformis* fauna, regarded as the largest deep water benthic foraminiferal extinction of the Cenozoic and late Mesozoic (Thomas, 1992), and a distinct sudden decrease in the isotopic composition of dissolved inorganic carbon in the oceans (Kennett and Stott, 1991). In addition, a conspicuous generic and family turnover among the calcareous nannoplankton (Aubry, 1998a) and accelerated rates of evolution among the planktonic foraminifera (Kelly *et al.*, 1996) have been described. While these abrupt events have been related to tectonic (e.g., Beck *et al.*, 1995), volcanic (e.g., Thomas and Eldhom, 199; Knox, 1998), atmospheric (e.g., Rea *et al.*, 1990; Dickens *et al.*, 1995) and paleoceanographic (e.g., Miller *et al.*, 1987; Thomas and Shackleton, 1996) changes, there is no clear explanation as to the origin of the late Paleocene global biotic, climatic and oceanographic changes, and attempts at recreating the latest Paleocene ambient conditions through modeling have failed (e.g. Sloan and Rea, 1995; Sloan and Thomas, 1998). The release of greenhouse gas through the dissociation of methane hydrates following the warming of bottom waters (Dickens *et al.*, 1996) is currently seen as the best explanation for the Paleocene/Eocene events, but the cause of the initial warming remains unknown. The challenge in explaining the change in the course of evolution that occurred in the latest Paleocene, and which truly constitutes the beginning of the modern world as we know it, is multiple, and needs to be approached from different directions. However, I believe that it includes foremost the establishment of a correct stratigraphic framework of correlations between all sections that serve as support for late Paleocene-early Eocene geological reconstruction. Problems of correlation between deep sea sections became quickly obvious as it was recognized that the position of the $\delta^{13}C$ excursion relative to the NP9/NP10 zonal boundary and the Chron C25n/C24r and C24r/C25n reversal boundary varied between sections. Based on magnetobiochemostratigraphic correlations, it was proposed that the discrepancies did not reflect diachrony, but instead were best explained by unconformities occurring at different levels in different sections. DSDP Sites 550, and 549 and ODP Site 690 were essential in this demonstration (Aubry *et al.*, 1996, Stott *et al.*, 1996). It was subsequently suggested that some excursions recorded in deep sea sections were in fact pseudoevents as documented for DSDP Site 577 (Aubry, 1998b) rather than real events . Based on the stratigraphic record of Sites 550 and 690, a (virtual) composite stratigraphic section representative of Chron C24r was established that served to estimate the ages of the main events that occurred during this chron, and in particular of the biochronal markers and the $\delta^{13}C$ excursion (Aubry *et al.*, 1996; Berggren and Aubry, 1996). A major problem in this reconstruction was the fact that the Sites 690 and 550 sections probably do not overlap based on chemostratigraphy, with the implication that a second excursion or a shift should occur in upper Chronozone NP9 or lowermost Chronozone NP10.

The Gebel Awaina section in the Nile Valley (Egypt), has been seen to constitute an essentially complete record of Chron C24r (Schmitz *et al.*, 1996; Aubry, unpublished), but failure to find a land or deep sea section with a comparable record of events between the $d^{13}C$ excursion and the NP9/NP10 (chron)Zonal boundary has precluded until now demonstration that that was the case. The excellent upper Paleocene-lower Eocene Bass River section (Cramer *et al.*, 1997) recovered on the New Jersey Coastal Plain provides the evidence that validates this view, when interpreted in the light of the stratigraphy of the adjacent Island Beach Borehole (Miller *et al.*, 1996). Unconformities occur in the upper Paleocene - lower Eocene sections recovered from the Island Beach and Bass River coreholes where the Manasquan Formation (lower Eocene) rests unconformably on the Vincentown Formation (upper Paleocene). In the Island Beach Corehole, the unconformity between the upper Paleocene Vincentown and the Manasquan formations result in Subzone NP10b resting unconformably on the lower part of Zone NP9 (the $d^{13}C$ excursion in this core is a pseudoevent). In the Bass River Corehole, Subzone NP10d rests on upper Zone NP9 and the $d^{13}C$ excursion reflects the same event as the excursion in ODP Hole 690B). The physical evidence (unconformities) in two adjacent sections rules out the possibility that the discrepancies between paleontological, magnetostratigraphic and chemostratigraphic events in the two sections are the result of diachrony, and confirms that many upper Paleocene — lower Eocene sections are riddled with unconformities. Other sections yield a discontinuous record of events, not because of the presence of unconformities, but because of the occurrence of a barren interval possibly reflecting a regional dissolution event. This is true of the sections located in Spain, and in particular for the Alamedilla section. The carbon isotope excursion recorded in this section (Lu *et al.*, 1996) is a pseudoevent, as is the benthic foraminiferal extinction. The latter occurs just below a 20 cm-thick barren interval whereas the former occurs immediately above it. While non-carbonate-bearing intervals have a genetic history different from that of unconformities, the two must be treated in the same fashion when correlating between events in different sections. The upper part of Zone NP9 is present in the Alamedilla section and of comparable paleontologic content as in the Bass River section, again confirming the non-diachronous nature of at least the main paleontological events over long distance.

We are now in a position 1) to establish a comprehensive temporal interpretation of upper Paleocene — lower Eocene sections over the world for sound correlations between events upon which proxies of oceanographic/atmospheric changes (such as $d^{13}C$, $d^{18}O$) can be compared, 2) to refine the chronology of events through Chron C24r in our composite reference section, and 3) to select stratigraphic intervals which may yield the most critical geological evidence. We must now search for the section other than Gebel Awaina which will provide the best representation of Chron C24r so as to locate precisely the NP9/NP10 biochronal boundary in the chron, a prerequisite for determining accurately the rates at which changes occurred during this time. A critical understanding that we have gained from studying the upper Paleocene — lower Eocene interval is that stratigraphic sections should not be viewed as approximate representations of a given interval of geological time. Upper Paleocene-lower Eocene stratigraphic sections constitute windows of various extent on late Paleocene - early Eocene time. For an essentially complete representation of this latter, upper Paleocene-lower Eocene stratigraphic sections must be assembled like pieces of a puzzle.

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ABRUPT CLIMATIC, OCEANOGRAPHIC AND ECOLOGIC CHANGES NEAR THE PALEOCENE-EOCENE TRANSITION IN THE DEEP TETHYS BASIN : THE ALAMEDILLA SECTION : SOUTHERN SPAIN

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The Tethys is a critical region for investigating the mechanism(s) of the Paleocene-Eocene global change, because of its potential in producing warm saline water masses, a possible driving force for the deep ocean warming at this time. Expanded sedimentary records from the Tethys reveal unique faunal and isotopic changes across the Paleocene-Eocene (P-E) transition.

To examine climatic, oceanographic and ecologic changes in the deep Tethys basin, we conducted high resolution faunal, isotopic and mineralogic analyses across the P-E transition at the Alamedilla section (paleodepth between 1000 m and 2000 m) in southern Spain (Fig.1). At this location, foraminiferal $\delta^{18}O$ values show little temperature change in surface waters, but a $4^{o}C$ warming in bottom waters (Fig.2). Comparison with deep-sea sites indicates that Antarctic intermediate water was consistently colder than Tethys bottom water. During the course of the P-E global change, however, the temperature difference between these two water masses was reduced from a previous $5^{o}C$ to $3^{o}C$. Clay mineralogic analyses at the Alamedilla section indicate increased aridity in the Tethys region that contrasts with a humid episode on Antarctica during high-latitude warming. Foraminiferal $\delta^{13}C$ values at Alamedilla show a negative excursion of 1.7‰ in both surface and bottom waters with little change in the vertical $\delta^{13}C$ gradient. Accumulation of organic and inorganic carbon in sediments decreased significantly, suggesting changes in the size and structure of the oceanic carbon reservoir (Fig.1). Associated with these climatic and oceanographic changes is a reorganization of the Tethys ecosystem, a benthic foraminiferal mass extinction, and planktonic foraminiferal assemblages marked by increased species turnover rates and high relative abundance of short-lived, opportunistic species that suggest increased instability. Benthic foraminifera suffered a mass extinction as shown by a 47% reduction in species richness, a magnitude similar to open oceans. Decreased influx of organic particles to the sediment/water interface might have contribute to the benthic foraminiferal extinction in the Tethys.

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Unlike in the open oceans, this part of the Tethys exhibits a gradual decrease of 1.5‰ in $\delta^{13}C$ values prior to the rapid $\delta^{13}C$ excursion (Fig.2). The existence of a prelude decrease in $\delta^{13}C$ values has been observed in other sections (Zumaya, Trabakua, northern Spain) and suggests that the P-E $\delta^{13}C$ excursion may have occurred in two steps and over a few hundred thousand years, rather than as one step over a few thousand years as previously suggested. This slower excursion rate is readily explained by changing organic carbon weathering or burial rates and avoids the need of invoking *ad hoc* scenarios.

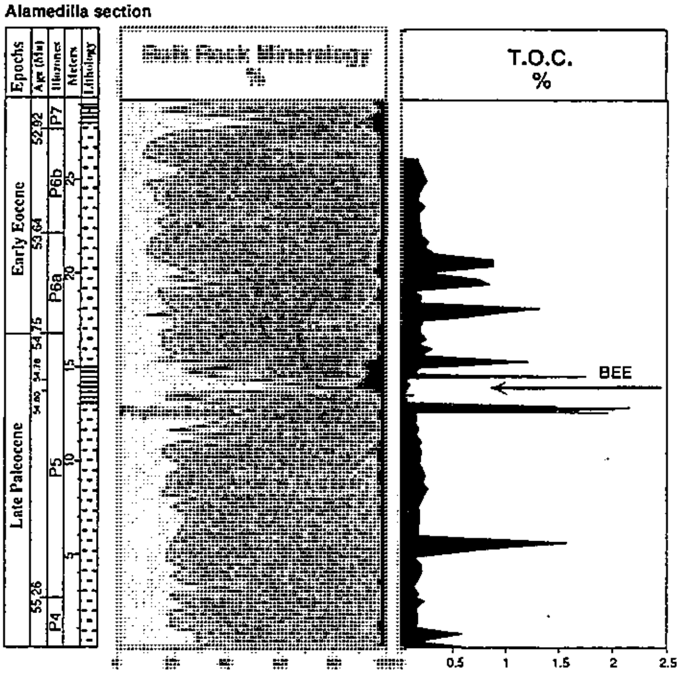

Figure 1 : Bulk rock composition and organic carbon acumulation at the Alamedilla section. Note the decrease in T.O.C. content during the BEE

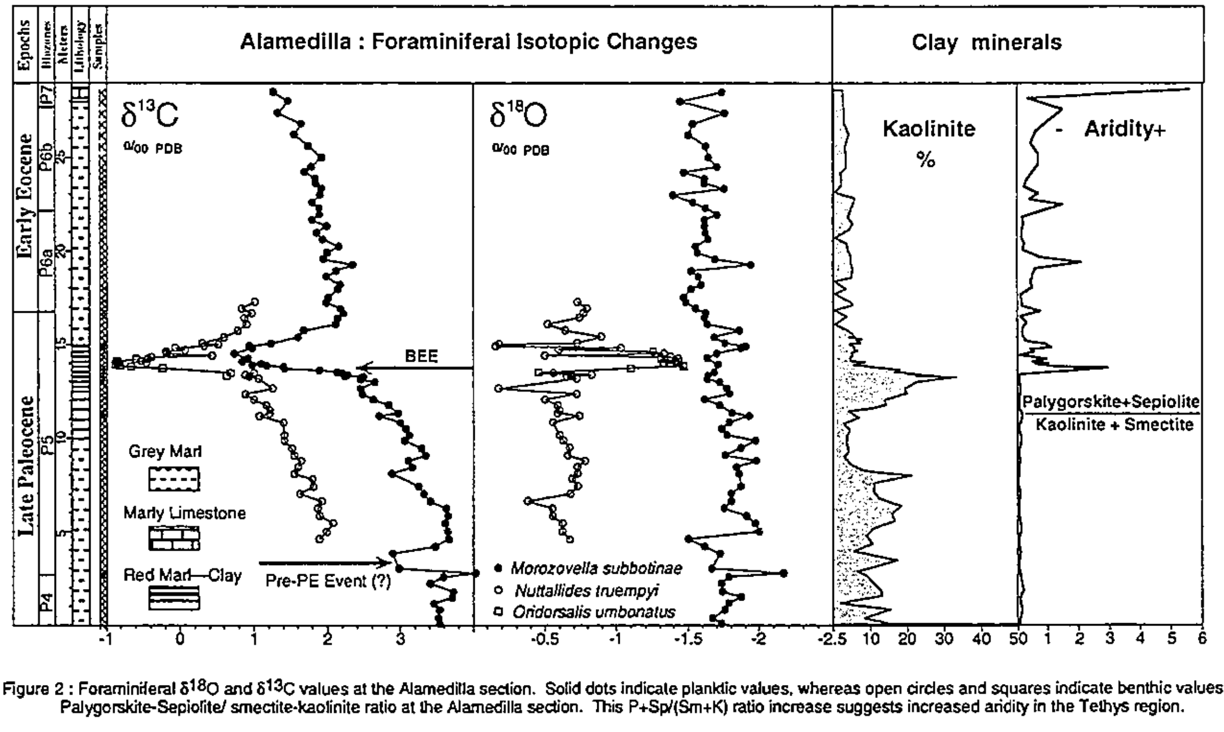

Figure 2 : Foraminiferal $\delta^{18}O$ and $\delta^{13}C$ values at the Alamedilla section. Solid dots indicate planktic values, whereas open circles and squares indicate benthic values. Palygorskite-Sepiolite/ smectite-kaolinite ratio at the Alamedilla section. This P+Sp/(Sm+K) ratio increase suggests increased aridity in the Tethys region.

THE PALAEOCENE / EOCENE BOUNDARY AT THE HIGH CHAIN. EXTERNAL BETIC ZONE (S SPAIN)

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ABSTRACT

In the Palaeocene / Eocene boundary at the High Chain Domain (Subbetic of the External Betic Zone, wester alpine belt) accounted a stratigraphic unconformity, with a gap affecting at least the Early and Middle Ilerdian. We have performed a bulk and clay minerals study of samples belonging to this domain, with the aim of evidence climatic, sedimentologic and paleogeographic changes.

INTRODUCTION

The High Chain, a mountain chain originated by a complex thrusting and located at the NE of the Málaga province (S Spain), belongs to the Subbetic (External Zone) of the Betic Cordillera (Peyre, 1974), the more occidental of the european alpine belts. The internal part of the External Zone represents the more basinal part of the sothern iberian palaeomargin between the Mesozoic and the Lower Miocene. Although this domain was mainly structured in folding and thrusting under compresive tectonic conditions during the Lower and Middle Miocene (Martín-Algarra, 1987), some tectonic should be considered during the tertiary sedimentation, as evidenced by local changes of facies and the development of unconformities.

This comunication shows the Palaeogene changes in the External Betic Zone, of great importance for the reconstruction of the yet poorly known palaeogeography of the High Chain, and with some implications for the Palaeocene / Eocene boundary. For this purpose, we have choose four sedimentary marine succession : Buitreras, Villanueva del Rosario, Sierra Prieta *s.s.* and Sierra Prieta element, respectivelly.

Total and clay mineralogy were determined using a Phillips PW-1710 diffractometer, with automatic slit, Cu-Kα radiation and 2°/min scanning rate from 2-60° 2θ. The oriented aggregates were treated with ethylene glycol and dimethyl sulphoxide and heated to 550°C. The reflecting powers of Biscaye (1965) and Babarahona-Fernández (1974) were used for the mineral quantification.

THE HIGH CHAIN DOMAIN DURING THE PALAEOCENE AND EOCENE

From a stratigraphic and sedimentologic point of view, we have divided the Tertiary of the High Chain has been divided into several depositional sequences, bounded by some recognized unconformities: Paleocene, middle Ilerdian to Lower Oligocene, Upper Oligocene to Aquitanian and Lower Burdigalian. The date presented here are refered to the first two depositional sequences.

The Palaeocene Depositional Sequence: It is boundered, at the botton and the top, by unconformities, with gaps affecting the Upper Cretaceous and the early-middle Ilerdian, respectively.

This sequence, of about 150 m in thickness, is made up of dark calcareous conglomerates (fed from the Subbetic mesozoic) at the bottom, and mainly grey Microcodium limestones with some intercalations of sligh levels of white marls and lutites. This marls and lutites are more abundant upward. The succession, dated by Peyre (1974) as early Paleocene at the bottom, has supplied the following association of planktonic foraminifera at the top: *Globorrotalia laevigata, G. velascoensis and G. pseudomenardii.* This assocciation date the early Palaeocene (Zones P4 and P5, Blow).

The samples (Fig. 1) are made up of quartz (5 to 55 %), phyllosilicates (15 to 55 %), calcite (up to 75 %), K-feldspar (up to 10 %) and traces of opal CT, dolomite and clinoptilolite. The clay minerals found are smectite (25 to 45 %), mixed-layers I/S (20 to 45 %), illite (25 to 40 %) and traces of chlorite, kaolinite and palygorskite.

The presence of mixed-layers I/S and the high amounts of smectite (more abudant upward) seem to evidence the existence of a stational climate in the neighbour continental areas. The averages of quartz and clay minerals decrease upward, meanwhile the calcite increases. These facts are interpreted as a progresive deepening in the marine conditions, in good accordance with the increase in the amount of smectite upward (Chamley *et al.*, 1990). The presence, in trace averages, of neoformed minerals such as dolomite and palygorskite indicates the erosion of cretaceous sediments rich in Mg (Martín-Martín *et al.*, 1994, Alcalá-García *et al.*, 1997).

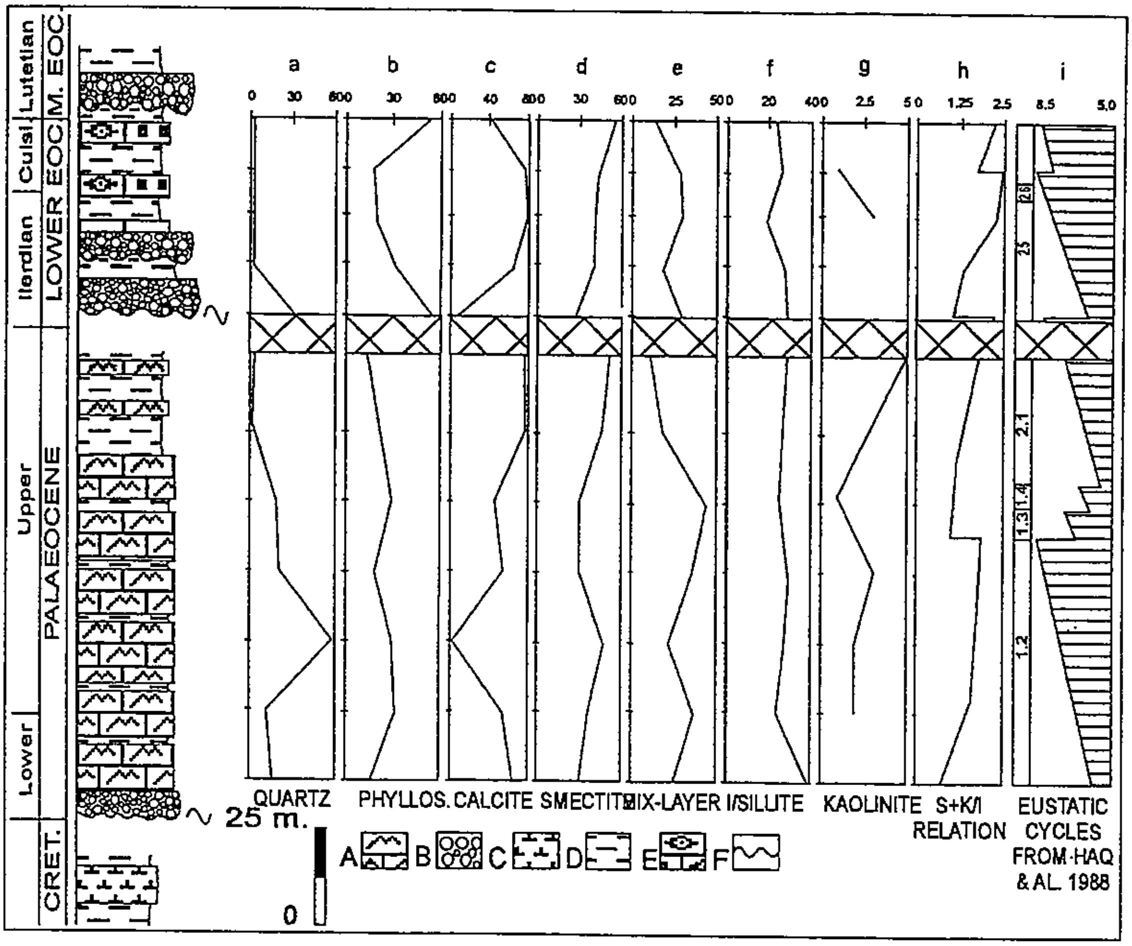

Figure 1. Synthetic stratigrafic column of the High Chain (S. Spain). Lithology: A-*Microcodium* Limestone; B-Calcareous Conglomerates; C-Marly Limestones; D-Marls; E-Nummulites rich Limestones; F-Unconformity. The columns *a* to *g* represent the minerals average. The column *h* represent the Eustatic Curve obteined from the smectite+kaolinite/illite ratio Daoudi *et al.*, (1995) for the eustatic curves. The column *i* represent the Eustatic Curve from Haq *et al.* (1988).

The middle Ilerditan to Lower Oligocene Depositional Sequence: it is boundered, at the botton and at the top, by unconformities with gaps that affect to the early Ilerdian and the Middle Eocene to Lower Oligocene, respectively. This sequence consist of a succession of about 30 m in thickness, made up of conglomerates and breccias (fed from mesozoic and palaeocene sediments), nummulite rich limestones and whitish, pinkish, and greenish marly intercalations. The presence of *Nummulites exilis* (Peyre, 1974) at the bottom, indicates a late Middle Ilerdian age (Serra-Kiel *et al.*, 1993), whereas the presence of *Assilina cf. exponens* (Peyre, 1974) at the top indicates a late middle Lutetian age (Serra-Kiel *et al.*, in press).

These samples are composed by (cf. Fig. 1) quartz (0 to 35 %), phyllosilicates (20 to 55 %), calcite (5 to 80 %) and traces of K-feldspar and opal CT. The clay minerals are smectite (25 to 55 %), mixed-layers I/S (15 to 30 %), illite (20 to 30 %), palygorskite (5 to 10 %) and traces of chlorite and kaolinite.

The low average of mixed-layes I/S and normal average of illite evidence the scarce presence of typical edaphic minerals, caused by a minor soils development. A fast decreasing, after initial high averages, in the quartz and illite content, is noticeable. Also, the absence of calcite, the low average of smectite at the botton and their progresive increasing upward is constated. These facts, together with the decreasing upward in the illite content, suggest a evolution upwards in the depth of the marine basin. The progresive increasing in phillosilicates, calcite, and neoformed minerals suggest that marly-calcareous mixed sediments were eroded (Alcalá-García *et al.*, 1997), located probably in close by jurassic terrains with a Middle Subbetic-like succession (High Chain, Martín-Algarra, 1987).

CHANGES IN THE HAIGH CHAIN IN THE PALAEOCENE / EOCENE BOUNDARY

1.- A stratigraphic gap occurs in this boundary, affecting at least the early and middle Ilerdian. The occurrence of sediments of this periods in another domain allow us to exclude a glacioeustatic change to explicate this gap. Regional changes in the plate tectonics must be proposed: during this period, a general change in the tectonic regime ocurred in the western alpine belts, trasforming the Tethysian divergent continental margins into compressive areas. Althought in the Betic External Zone this phenomenons were not well developed just to the Lower and Middle Miocene, some minor folding should be in this area that could favour the occurrence of unconformities and gaps. The observed gap in this domain is also present in the Malaguide Domain, but more developed since it affect, at least, to the entire Ilerdian (*cf.* Martín-Martín, 1996).

2.- Figure 1 shows the correlation at the High Chain between lithostratigraphy, mineralogical trends and the smectite+kaolinite/illite ratio. These variables are correlated with the eustatic curves from Haq *et al.* (1988) and Daoudi *et al.* (1995). The figure propose too a great transgression in this area located at the bottom of the late middle Ilerdian, and that probable existence of third order cycles 2.3 and 2.4.

3.- Although the presence of mixed-layers I/S and the high average of smectite (more abundant upward) during the Late Paleocene seem to evidence the development of stational soils in the neighbour continental areas caused by a mediterranean-like climate, during the Lower Eocene the lower amounts of mixed-layers (due to the minor presence the of illite) could evidence a change, caused by warning in the climate (arid to semiarid).

4.- The presence of opal CT, clinoptilolite, and a clay association of smectite+illite+mixed-layers I/S in the Palaeocene sediments reflects the erosion of cretaceous terrains, meanwhile during the Lower Eocene, the presence of smectite and lower averages of illite, mixed-layers and kaolinite, and traces of palygorskite, could indicate the erosion of mixed marly and calcareous jurassic sediments (probably the High Chain underlayer serie, *cf.* Peyre (1974). This fact evidences the subaereal exposure of cretaceous (during the Palaeocene) and later Jurassic (during the Lower Eocene) sediments.

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RESOLVING POSSIBLE PROBLEMS ASSOCIATED WITH THE MAGNETOSTRATIGRAPHY OF THE PALEOCENE/EOCENE BOUNDARY IN HOLES 549, 550 (GOBAN SPUR) AND 690B (MAUD RISE)

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Introduction

The Paleocene/Eocene transition (~55 Ma) is associated with several important events in Cenozoic history including: major turnover in the terrestrial mammalian fauna; dramatic d^{13}C decrease in the deep sea; major benthic foraminiferal extinction; long-term global ocean temperature increase; weakening of atmospheric circulation; increased precipitation and/or temperature reflected in increased rates of weathering and the production of kaolinite (Hooker, 1991, 1996; Gibson et al., 1993; Aubry et al., 1996; Berggren & Aubry, 1996; Stott et al., 1996; Charisi & Schmitz, 1996; Thomas & Shackleton, 1996; Berggren et al., 1997). Some workers (e.g. Eldholm & Thomas, 1993) have speculated that the root cause of these events was the explosive volcanism associated with the initial opening of the NE Atlantic. Attempts to elucidate the temporal relationships of these boundary events have, to a large extent, focused on (i) the stratotypes of the Thanetian and the Ypresian Stages, NW Europe; (ii) DSDP Holes 549 and 550 on the Goban Spur eastern N. Atlantic; and (iii) ODP Hole 690B on Maud Rise in the Southern Ocean.

A key element in chronostratigraphic studies of the Paleocene/Eocene boundary is magnetostratigraphy. Whilst based at Southampton, we were involved in several such studies of the Thanetian and Ypresian Stage stratotypes and their neighbouring outcrops (e.g. Townsend & Hailwood, 1985; Aubry et al. 1986; Ali et al., 1993; Ali & Hailwood, 1995; Ali & Jolley, 1996). Thus, we have followed the debate related to the establishment of a detailed, globally-applicable, chronological framework for the Paleocene/Eocene boundary interval with some interest. As a great deal of this work focused on some of the deeper ocean sequences, we began to examine the related magnetostratigraphic studies, which in almost all cases provides the primary chronostratigraphic reference frame. In summary, we feel that there are potentially serious problems associated with the magnetostratigraphy of the Paleocene/Eocene boundary at DSDP Holes 549 and 550 on the Goban Spur and ODP Hole 690B drilled on Maud Rise (we outline those concerns below). On a more positive note, we will present at the Paris meeting the results of new studies carried out on the C25n-C24r -C24n.3n interval in all three of these sections following a recent re-sampling of the cores.

Holes 549 and 550, Goban Spur, eastern N. Atlantic

The Goban Spur sites are particularly important as they record a number of the *Paleocene/Eocene events* within an expanded sequence of sediments assigned to Chron C24r (the magnetochron bracketing the broad Paleocene/Eocene transition interval, e.g., Aubry et al., 1996). Additionally, second order correlation of Ar-Ar dated tephras within Hole 550, calibrated relative to the lower and upper bounds of Chron C24r, provide a 55 Ma tie point in Cande & Kent's (1992, 1995) geomagnetic polarity time scale.

Ali & Hailwood (1995) highlighted possible problems with the normal polarity magnetozones recording Chron C25n (lower bound of Chron C24r) in Hole 549 and Chrons C25n and C24n.3n (lower/upper bounds of Chron C24r) in Hole 550. The magnetostratigraphic data from that study are presented in Fig. 1. Downhole plots of "NRM inclination" and the "shift in inclination after demagnetisation" are used in this diagram to highlight the problems of polarity determinations. The original paleomagnetic analyses were carried out by Townsend (1981), using the Glomar Challenger ship-board paleomagnetism laboratory. The demagnetisation facilities available were limited and it is now apparent that the peak alternating fields used were probably too weak to permit a reliable polarity assignment to be made for many key samples. In detail, the basic problems are as follows :

Hole 549

Chron C25n is based on four specimens from three sections within Core 18 (351.53-353.81 metres below sea floor, mbsf). Applied demagnetisation fields (AF)were very low (15-20 mT; detailed stepwise demagnetisation to 30 to 60 mT would probably have been more appropriate). The inclination shift after demagnetisation is towards shallower values and all specimens show inclinations < 20° when values of 40-60° might be expected. Thus, the magnetozone representing Chron C25n is regarded as suspect. The record of Chron C24n.3n, based on 9 specimens from 8 sections within cores 13 and 14 (306.27-317.16 mbsf), appears more reliable. Following demagnetisation, inclinations range from 19.9° to 48.1°. However, demagnetisation was typically carried out to no more than 10 mT, and the application of higher AF cleaning might have been more appropriate.

Hole 550

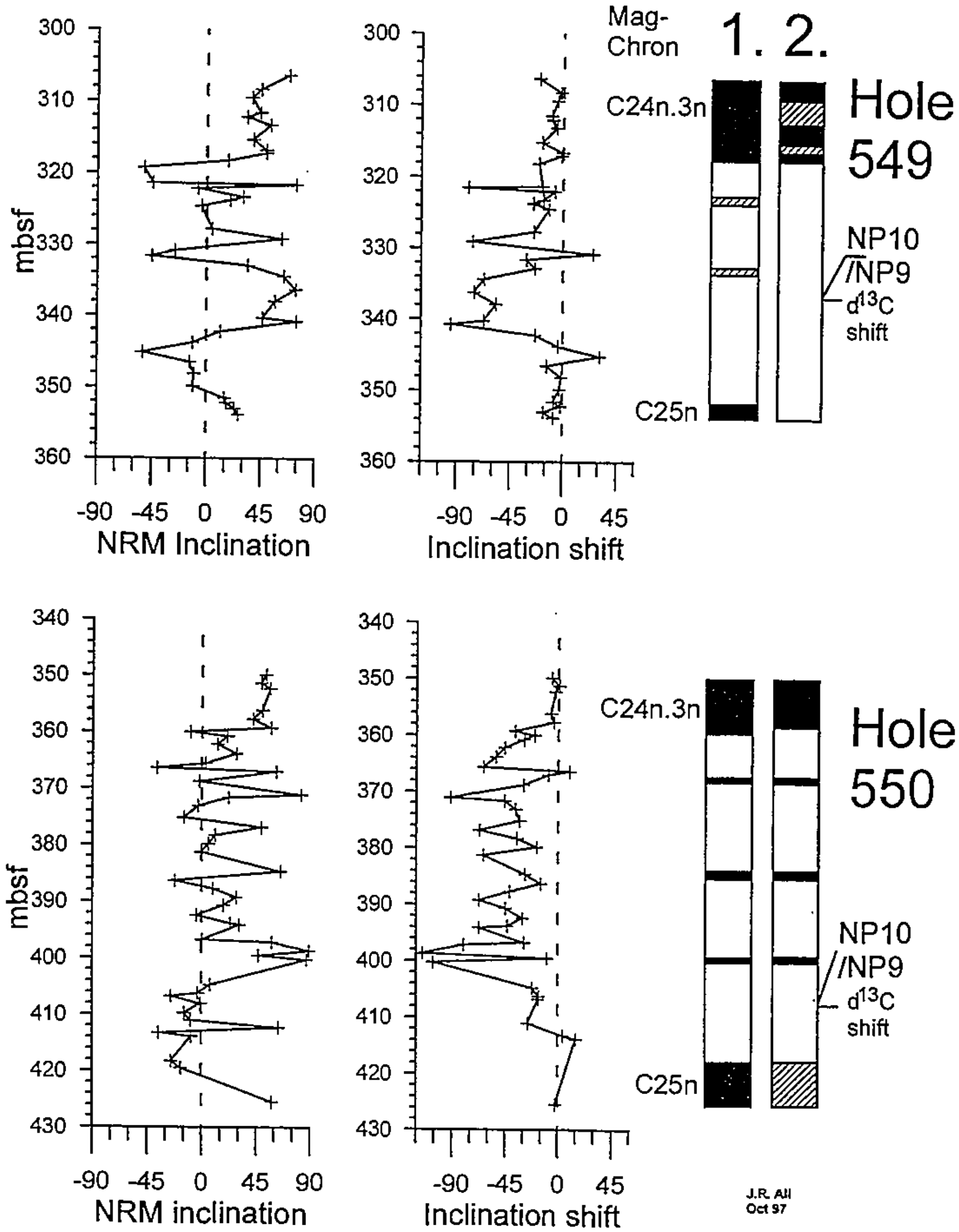
The record of Chron C25n is based on just one specimen (within section 36-2-101 at 425.01 mbsf; not 425.5 mbsf as listed in Townsend's Table 4). The three normal polarity specimens immediately below this specimen are assigned to Chron C26n (Townsend, 1981, Fig. 16). The magnetic inclination of the specimen is 55.7°, but demagnetisation was to just 5 mT. Because Chron C25n is based on a single sample, and alternating field cleaning took place only at 5 mT, the magnetozone is considered suspect. Chron C24n.3n is based on 6 specimens from the interval 359.27-349.86 mbsf. Following demagnetisation, inclinations ranged from 20-50°. However as demagnetisation was to no more than 15 mT, the polarity of the lowermost sample, which helps define the start of Chron C24n.3n, must be considered as suspect because the specimen's magnetic inclination shallowed from 36.0° to 19.9° following demagnetisation. It is possible that the lower bound of Chron C24n.3n may have to be shifted up core to below sample 550-29-2, 26-29, at a depth of 357.76 mbsf.

The aim of the Goban Spur re-investigation is to verify the magnetostratigraphic succession spanning Chrons C25n through C24n.3n in Holes 549 and 550. Closely spaced sampling should enable magnetochron boundaries to be fixed precisely. Also, it should be possible to appraise some of the short period normal polarity events identified within Chron C24r by Townsend (1981). Moreover, the application of extensive paleomagnetic analyses and experimental procedures not available to Townsend (1981) should lead to a fuller understanding of the chronostratigraphy of these key Paleocene/Eocene boundary sections.

Hole 690B, Maud Rise, Southern Ocean

The Hole 690B section is important because it too forms an expanded sequence and records several of the Palaeocene-Eocene boundary events, including the d^{13}C isotope excursion. Aubry et al. (1996, p 356 and Fig. 1) raised concerns regarding the Hole 690B Paleocene/Eocene boundary magnetostratigraphy (Spiess, 1990) due to the apparent mismatch of the polarity and biostratigraphic data. For example, there is a dominantly normal polarity magnetozone straddling the NP9-10 nannofossil zone boundary (which Spiess, 1990, correlated with Chron C24n.3n) which is essentially impossible to accommodate based on our knowledge of magneto-biostratigraphic correlations for this period. We have examined Spiess (1990) in some detail and feel that a major problem exists in that the paleomagnetic declinations from the Paleocene/Eocene boundary interval are all aligned in the direction of 240° -310° with a mean of about 270°, i.e., parallel with the split face of the core (Fig. 2). For ODP hydraulic piston [H] cores we expect that (i) *within each core* the declinations will be aligned about a randomly oriented mean with about 20° dispersion, due to geomagnetic field secular variation, either side of that value. (ii) If a geomagnetic field reversal is recorded in a core, it should be a marked by 180° change in declination, and the inclination will be inverted. (iii) For different cores, we expect the declination means to be randomly oriented with respect to each other. However, this is not the case in Hole 690B where declinations are uniform for each of the cores and polarity inversions, based on the inclination information, are not marked by significant declination shifts. There are various possibilities to explain the clustered declination data from Hole 690B. This is unlikely to be a sample "push" remanence as described by Hailwood & Clement (1991) in their work on S. Atlantic Leg 114 cores, since in core co-ordinates the declinations would tend to cluster around 0° or 180° rather than the observed 270 °. It is more likely that this remanence is related in some way to the cutting of the core along its long axis before sampling and measurement.

Therefore, the aim of our Hole 690B re-investigation is to investigate the possibility that disturbed remanences might contaminate the magnetostratigraphic record. In doing so, we aim to clarify the status of several anomalous polarity intervals recorded in Hole 690B, which are in conflict with well-established magneto-biostratigraphic correlations.


Mag-
Chron
1. 2.
Hole
549
C24n.3n
NP10
/NP9
d13C
shift
C25n
mbsf
300
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320
330
340
350
360
-90 -45 0 45 90
NRM Inclination
-90 -45 0 45
Inclination shift
Hole
550
C24n.3n
NP10
/NP9
d13C
shift
C25n
340
350
360
370
380
390
400
410
420
430
-90 -45 0 45 90
NRM inclination
-90 -45 0 45
Inclination shift
J.R. Ali
Oct 97

Fig. 1

It is worth noting also, that the ~270°±30° declination effect is pervasive throughout Hole 690B and the comments above, related to the Paleocene/Eocene boundary interval, are probably applicable to the entire cored interval in this hole.

Procedures adopted for sampling the Hole 549, 550 and 690B material

The standard DSDP-ODP procedure for the paleomagnetic sampling of soft to semi-indurated sedimentary material is to push sample boxes into the open split face of the core. Experience has shown that this can result in severe distortion of the sediment and the magnetic signal (e.g. Hailwood & Clement, 1991). Instead, sampling carried out by JRA in July 1997, involved the careful cutting of material from the core before the trimmed sediment cube was placed into its sample box (as was the case for much of the work reported in Ali & Jolley, 1996).

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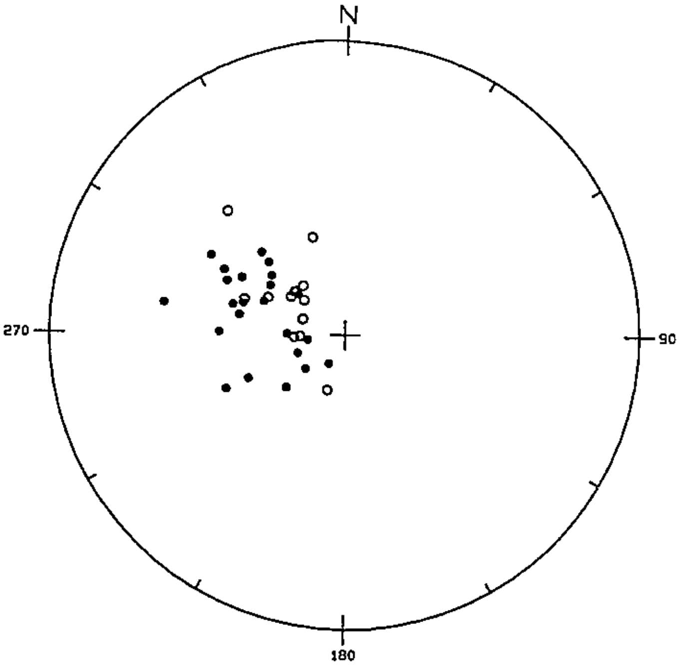
Figure Captions

1. Magnetostratigraphy of DSDP Holes 549 and 550. Black = normal polarity; white = reverse polarity, hatched =questionable polarity. "Inclination shift" is the positive or negative angular change in the inclination values after demagnetisation. 1 = magnetostratigraphic interpretation of Townsend (1981) which is used in various Paleocene/Eocene Boundary studies. 2 = a possible alternative interpretation of that same inclination data set which questions some of Townsend's magnetozone definitions (and chron correlations). Magnetochron nomenclature is based on Cande & Kent (1992, 1995). The NP9/10 boundary and d^{13}C shift are shown also.

2. Remanence data for specimens sampled ~100 cm below the top of each core section in the Paleocene-Eocene transition interval in Hole 690B (from Spiess, 1990, Appendix B). Specifically, the data are from Section 23H-3 (basal Chron C25n) up to Section 15H-5 (uppermost NP11). See text for discussion.

Fig. 2



THE BOUNDARY OF THE PALEOCENE AND EOCENE IN UKRAINE AND OTHER CIS REGION BY NANNOPLANKTON AND DINOCYSTS

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Nannoplankton and dinocysts have been studied from the Paleocene-Eocene boundary interval in the stratigraphical sequences of the North Slope of Ukrainian Carpathians and the Crimea (Ukraine), Armenia and Eastern Precaspian (Kazakhstan).

The Carpathians. The Paleocene-Eocene boundary interval of the North Slope of the Carpathians is represented with:

1) The Yamna formation (Fig.1.a 1,2; b 2) consisted with the massive sandstones with the rare interbeds of the argillites or the thin-rhythmical flysch. At the bottom of the formationYaremcha horizont (Fig.1,a 1) represented by the variegated thin-rhytmical flysch is frequently observed. The age of the formation is determined as Later Paleocen the grounds of the finds of N.solitarius de la Harpe (Nemkov, Khloponin, 1957) and Discocyclina sp. (Portniagina, Sovchik, 1968), the benthos foraminifers (Vjalov et all. 1988), nannoplankton of zones NP6-8 in the formation bottom and NP9 in the upper argillite interbeds. The association of dinocysts is improverished and represented by zones Cerodinium speciosum and Apectodinium homomorphum.

The Yamna formation (Fig. 1 a 1,2; b 2) deposits are without any break covered by the thin - rhythmical noncarbonate often flintationed flysch of the Maniava formation (a.3, river Prut) or by the Vitvitza formation thin-rhythmical noncarbonate clay flysch (b3, river Tismennitsa, t. Boryslav). The age of these formations is determined by the benthos foraminifers,the finds of N.planulatus Lam. in the upper formation part (Vjalov et al.,1988). The nannoplankton zone NP12 is ascertaine only in the marl facies of the upper part of the Maniava formation along the Opor river. Dinocysts are represented by the impoverished associations of zones Dracodinium simile and Charlesdowniella colyotrypta (Fig.2). So the boundary between the Paleocene and the Eocene in the North Carpathians is put up by the sharp lithological change. Tracing the changes in the development of the biota of nannoplankton and dinocysts is not possible yet, and the additional investigations are necessary.

The Crimea. The position of the boundary between the Paleocene and the Eocene has been studied in the parastratotypical well N1 been obrilled in the neighbourhood of the t. Bakhchisaray. The boundary interval is represented by marls of the Kachia formation in which the zones NP6 - NP9 and the lowest parts of the zone NP9 (Fig.1,C1) and dinocysts zones C.speciosum and Apectodinium homomorphum were found out. The carbonate clays of the Bakhchisaray formation with the scouring (Fig.1,C2) in the clays the zones from NP11 to NP13 are determined. The dinocysts are represented by the zones D.simisle, D.variolongituda and the lowest parts of the C.coleotrypta one. So the boundary between the Paleocene and the Eocene coincides with the break in the sedimentation which corresponds to the most part of the zone D. multiradiatus and the zone M.contortus by time.

Armenia. Nannoplankton and dinocysts have been studied from the Vedy and Shagap sequences of the Erevan basin. The boundary's deposits are represented by the upper parts of Katutska formation (Andreeva-Grigorovich et al., 1991) which are composed with the alteration of sandstones, aleurolites and argillites and also conglomerates and clays (Fig.1 d 17,18,19). The association of the nannoplankton zones NP6 and NP8 is established in the bed 17 and the upper part of the zone NP9 - in the bed 19. The dinocysts are represented by Deflandrea oebisfeldensis (akme - zone). The limestone succession with N.planulatus overcoveres these deposits. The group of aleurolite strata lies on the nummulite limestones in the Shagap sequence. In it the zone NP13 is determined by nannoplankton and the zone Charlesdowniela coleofrypta - by dinocysts. The boundary between the Paleocene and Eocene is put at the floor of the limestones with N.planulatus.

Eastern Precaspian. The rock samples have been studied from the stratotypical well SP-1. The boundary interval is represented by the Khamsaktysol's formation composed by opokas, sandstones and clays. The zone NP9 and the beds with Deflandrea oebisfeldensis (akme-zone) are found out in the upper part of clays. Above these deposits the Bailesail's formation lies with scouring. It is the noncarbonate clays with the glauconite sand in the bottom. The zones NP11, NP12 and NP13 determinations by I.P.Tabachnicova are found out in the single carbonate interbeds. And the zone D.similes,D.variolongituda and C.coleotrypta by dinocysts (Fig. 1 e; Fig. 2). So the boundary between the Paleocene and the Eocene coincides with the break which corresponds to the upper part of the zone NP9 and NP10 by time.

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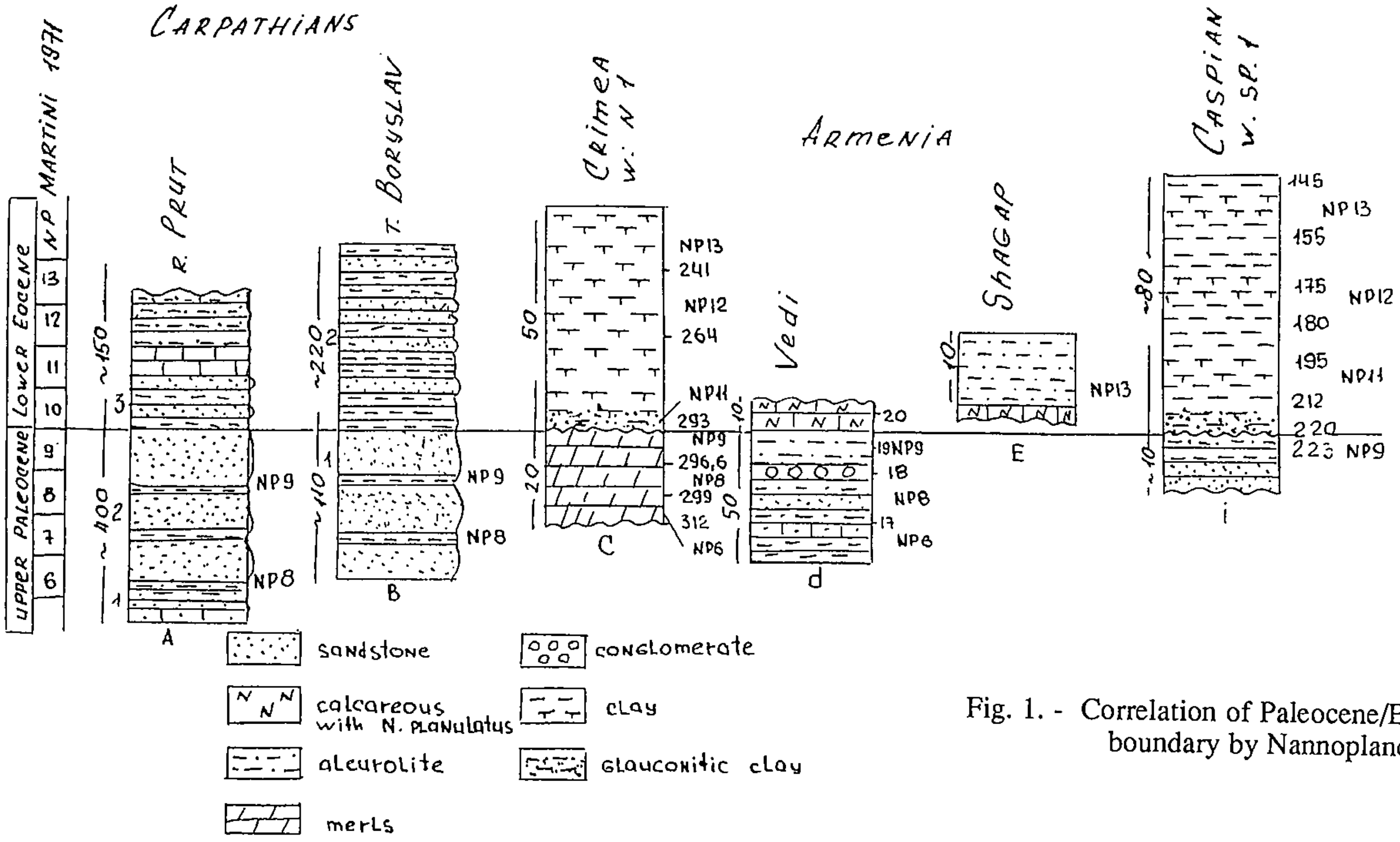

Fig. 1. - Correlation of Paleocene/Eocene boundary by Nannoplancton

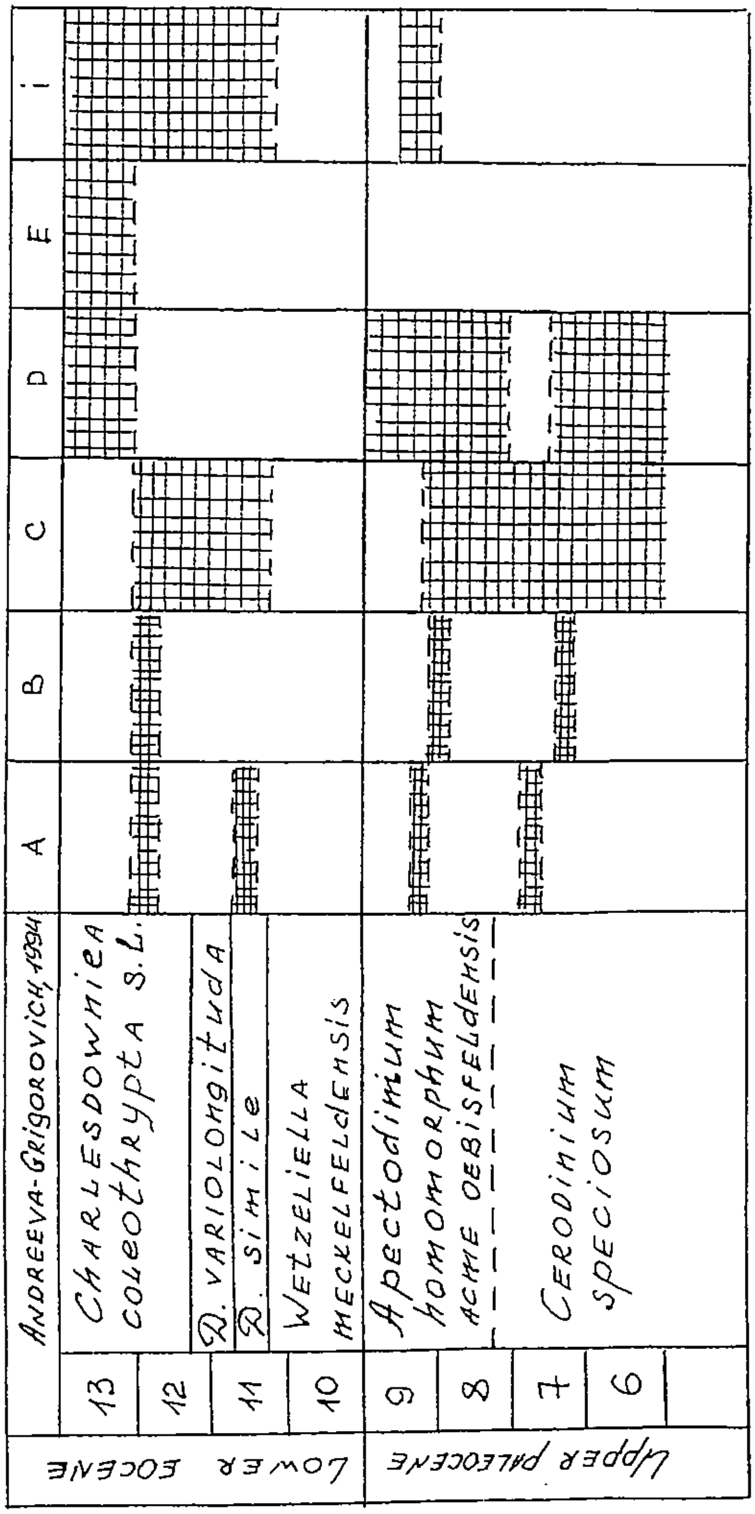

Fig. 2 - Correlation Paleocene - Eocene by dinocysts

BIOSTRATIGRAPHY AND PALEOBIOGEOGRAPHY OF ASIAN LAND MAMMALS NEAR THE PALEOCENE-EOCENE BOUNDARY : A PHYLOGENETIC APPROACH

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The fossil record of early Cenozoic mammal evolution in Asia has steadily improved over the past three decades, with some recent advances in knowledge being particularly noteworthy. This greatly enhanced set of primary data has provided the framework for development of a sequence of Asian Land Mammal Ages (Russell and Zhai, 1987; Ting, 1998), or ALMAs, that can be compared with similar biochronologies in Europe and North America. However, correlation of this Asian mammal record remains contentious because of the absence of terrestrial-marine intercalations, datable volcanics, and magnetostratigraphic and chemostratigraphic data there. For purposes of assessing faunal change across the Paleocene-Eocene boundary in Asia, two ALMAs are critical: the Gashatan ALMA, often regarded as a late Paleocene correlative, and the Bumbanian ALMA, frequently thought to be an early Eocene correlative. Here, I review previous ideas regarding correlation of the Gashatan and Bumbanian ALMAs with similar, mammal-based biochronologies established in Europe and North America. A revised correlation that takes into account the logical interplay among biostratigraphy, paleobiogeography, and phylogeny is then proposed.

Dramatic, yet similar, changes in the mammalian faunas of Europe and North America occur at or near the Paleocene-Eocene boundary, pending determination of the Global Stratigraphic Section and Point (GSSP) for this boundary. In North America, where the terrestrial record across this interval is most nearly complete, the Clarkforkian-Wasatchian boundary is marked by the synchronous first appearances of Artiodactyla, Perissodactyla, Primates, and Hyaenodontidae (e.g., Rose, 1981; Gingerich, 1989). The Clarkforkian-Wasatchian boundary occurs in the lower part of Chron 24r, coeval with a shift in carbon isotope ratios that has been recognized in both terrestrial and marine sections (Koch et al., 1992). In Europe the same four mammalian taxa first appear in the Neustrian (= Sparnacian of many workers), suggesting correlation between the basal Wasatchian and basal Neustrian. However, taxa that first appear earlier in the North American record, such as Rodentia and Coryphodon, also make their first European appearances in the Neustrian. Therefore, the magnitude of faunal turnover across the Cernaysian-Neustrian boundary in Europe is probably inflated due to the Thanetian hiatus.

The pattern of mammalian faunal turnover in Asia near the Paleocene-Eocene boundary is similar in some ways, and different in others, from that observed in Europe and North America. Earlier studies of faunal change across the Gashatan-Bumbanian boundary, based primarily on the sequence of faunas preserved in the Naran Bulak Formation, Nemegt Basin, southern Mongolia (Dashzeveg, 1982, 1988), emphasized the local first appearances of Artiodactyla, Perissodactyla, and Primates (in the form of Altanius orlovi) in the Bumbanian, although specimens attributed to cf. Sinopa (a hyaenodontid) were reported from the underlying Naran Member of the Naran Bulak Formation, of Gashatan age. Based on this pattern of faunal turnover, most workers have correlated the Gashatan-Bumbanian boundary in Asia with the Clarkforkian-Wasatchian boundary in North America and with the Cernaysian-Neustrian boundary in Europe (Dashzeveg, 1982, 1988; Russell and Zhai, 1987; Krause and Maas, 1990; Ting, 1998). If such a dramatic episode of faunal turnover on all three Holarctic continents could be shown to be synchronous, one would be forced to postulate a causal process of singular scope and exceptional magnitude.

Krause and Maas (1990) cite the early Tertiary collision between the Indian subcontinent and southern Asia as one means of explaining the rapid and widespread mammalian dispersal required by correlating the basal Bumbanian with the basal Wasatchian /Neustrian. Other workers (e.g., Gingerich, 1989) explain the same pattern as the result of rapid dispersal out of Africa.

Recent discoveries and renewed interest on the Gashatan-Bumbanian interval in Asia calls into question the degree to which the pattern of faunal change across the Gashatan-Bumbanian boundary in Asia matches that seen in Europe and North America. Specifically, two of the four taxa characterized by Neustrian/Wasatchian FADs are now documented by Gashatan records in Asia. The Gashatan record of a hyaenodontid from the Naran Member of the Naran Bulak Formation reported by Dashzeveg (1982, 1988) has been duplicated by Meng et al. (1998), who describe the new hyaenodontid species Prolimnocyon chowi from the Gashatan Bayan Ulan fauna, Erlian Basin, Nei Mongol Autonomous Region, China. Gashatan perissodactyls also occur in the Bayan Ulan fauna (Meng et al., 1998). Accordingly, faunal turnover across the Gashatan-Bumbanian boundary in Asia differs from that across the Cernaysian-Neustrian boundary in Europe and the Clarkforkian-Wasatchian boundary in North America in being characterized by the FADs of Artiodactyla and Primates alone.

Not only is the pattern of faunal change across the Gashatan-Bumbanian boundary distinct from that at the Clarkforkian-Wasatchian and Cernaysian-Neustrian boundaries, but there is also reason to doubt the previously hypothesized correlation of these boundaries. Instead, intercontinental correlation based on alagomyid rodents, known only from Asia and North America, suggests that at least some Asian Bumbanian faunas correlate with the North American Clarkforkian. Only two genera of alagomyids have been described: the primitive Gashatan genus Tribosphenomys, known only from the Bayan Ulan fauna of Nei Mongol (Meng et al., 1994, 1998), and the dentally derived Bumbanian genus Alagomys, documented in Asia from the Bumban Member of the Naran Bulak Formation in Mongolia (Dashzeveg, 1990) and the Wutu fauna of Shandong Province, China (Tong and Dawson, 1995). The sole North American record of alagomyids was recently described from Big Multi Quarry, Washakie Basin, Wyoming, where Alagomys is associated with a diverse early (but not earliest) Clarkforkian mammal fauna (Dawson and Beard, 1996). Correlation between certain Bumbanian faunas and the North American Clarkforkian is further supported by the presence of the carpolestid plesiadapoids Chronolestes simul and Carpocristes oriens in the Wutu fauna from Shandong Province, China (Beard and Wang, 1995). In North America, carpolestids apparently become extinct by the end of the Clarkforkian (Rose, 1981).

At the intercontinental scale, biostratigraphic correlation of continental rocks based on land mammals must be integrated with paleobiogeographic and phylogenetic reconstructions for the relevant taxa. Both biostratigraphy and paleobiogeography are logically dependent on phylogeny, because the continuity of phylogenetic descent requires that sister taxa originate in the same place (a constraint imposed by phylogeny on biogeographic reconstruction) and at the same time (a constraint imposed by phylogeny on biostratigraphic correlation). Hence, knowledge of the phylogeny of a clade allows one to reconstruct the biogeographical history of that clade, which can then be used to assess the likelihood of synchroneity versus diachroneity in intercontinental biostratigraphic correlation.

Phylogenetically-derived biogeographic reconstructions for mammalian taxa that immigrate into North America during the interval spanning the late Paleocene (late Tiffanian zone Ti5 = early part of Chron 25n; ca. 56.4 Ma) to early Eocene (basal Wasatchian zone Wa0 = early part of Chron 24r; ca. 55.5 Ma) reveal a repeating pattern whereby most or all immigrant taxa appear to have dispersed from Asia. This "East of Eden" biogeographic pattern now appears to be the dominant mode of modernization of the North American mammal fauna throughout the Cenozoic Era (Beard, 1998). Two major clades of mammals, the order Dinocerata (Probathyopsis) and the family Arctostylopidae (Arctostylops) exhibit late Tiffanian FADs in North America; both are interpreted as immigrants from Asia based on phylogenetic and biostratigraphic datasets. Three major clades of mammals, the order Rodentia (Paramys, Alagomys), the order Tillodontia (Azygonyx), and the family Coryphodontidae (Coryphodon), first appear in North America during the basal Clarkforkian (Cf1). Phylogenetic and biostratigraphic data show unambiguously that rodents and tillodonts originated in Asia and dispersed to North America at this time. Coryphodontids are interpreted as also conforming to the East of Eden pattern, although phylogenetic and biostratigraphic data are less clear in this case. Finally, during the basal Wasatchian (Wa0) four major mammalian clades immigrated into North America: Perissodactyla

(Hyracotherium), Artiodactyla (Diacodexis), Primates (Teilhardina, Cantius), and Hyaenodontidae (Prolimnocyon, Arfia, Prototomus, and Acarictis). Phylogenetic data unambiguously support an Asian origin for Perissodactyla and Primates; phylogenetic evidence equivocally supports an Asian origin for Artiodactyla. Biostratigraphic data support an Asian origin for Perissodactyla and Hyaenodontidae. All four Wasatchian immigrant taxa are interpreted here as having originated in Asia. The East of Eden model is consistent with seemingly anachronistic distributions of Gashatan and Bumbanian mammals in Asia because taxa showing otherwise inexplicably early Asian records (such as Gashatan perissodactyls and hyaenodontids) merely reflect an Asian origin for these clades, and are therefore to be expected rather than explained away.

Paleobotanical and isotopic data clearly indicate that the late Paleocene-early Eocene was an interval of sustained climatic warming (Zachos et al., 1994; Wing et al., 1995). The temporally staggered immigration into North America of three consecutive waves of Asian endemic mammals during this interval suggests that ameliorating climatic conditions, or reduction of latitudinal climatic gradients, allowed successively more cold-intolerant mammalian taxa to disperse across the high latitude filter of Beringia as global warming progressed. Once dispersal across Beringia was achieved, access to western Europe would only have been denied by sea-level highstands (such as that which occurred during Chron 25r, about 56.5 Ma). Barring this type of geographic segregation of western Europe from eastern North America, it is unlikely that the Cernay mammal fauna from the Paris Basin is as young as the North American Clarkforkian, because it lacks rodents, tillodonts, and Coryphodon.

Biostratigraphically, the East of Eden model suggests that both the Gashatan and Bumbanian ALMAs are likely to be older than traditional correlations have implied. The Gashatan probably correlates with roughly the latter half of the Tiffanian and possibly the earliest part of the Clarkforkian (approximately Chron 26n, Chron 25r, and the early part of Chron 25n). The Bumbanian probably correlates with most of the Clarkforkian and the early part of the Wastachian (most of Chron 25n and the first half or so of Chron 24r). These correlations based on mammalian biostratigraphy are sorely in need of testing on the basis of magnetostratigraphy, chemostratigraphy, and other methods.

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DYNAMICS OF DEVELOPMENT OF BENTHIC PALEOCENE - EARLY EOCENE FORAMINIFERA FAUNAS OF THE EUROPEAN PALEOBIOGEOGRAPHIC AREA (EPA)

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In the Paleocene-Early Eocene, the EPA (=Northern Peri-Tethys) was a very extended (ca. 7000 km), epicontinental seaway between the Fennoscandian craton, northern Russian craton, and the Tethys. The EPA encompassed the following basins: North Sea (NS), Southeastern England (SEE), Belgium (B), Denmark (Dk), southern Sweden (SS), Poland (PO), Lithuania (L), northern Ukraine (NU), lower Volga (LV), Crimea (C), Scythian Plate (SP), Northern Caucasus (NC), Peri-Caspian (PC), Mangyshlak (M), and Turkmenia (T). The West Siberian Sea (WSS) communicated with the EPA through the Turgai Strait (TS).

The Paleocene-Early Eocene EPA strata feature ten types of benthic foram faunas, each reflecting a distinct environment.

Type **1**: *"Ratalia saxorum"* foram fauna with *Vacuvalvulina keijzeri, Rotalia marginata R. saxorum, Cibicidoides cuvillieri, Schlosserina astericus, Rotorbinella marie, Rotalia lithotamnia, Boldia sp*. According W. Berggren and J.Aubert (1975) this complex reflects the tropical upper sublittoral zone , to 50 m. Type **2**: "*Anomalinoides danicus*" foram assemblage with *Osangularia lens*, *Pulsiphonina prima*, *Discorbis binkhorsti,Gavelinella pertusa, Anomalinoides. danicus, Brotzenella acuta, Heterolepa proprius, H. hemicompressus, Cibicidoides commatus, C. succedens, Karreria fallax, Tappanina selmensis*.

Type **3**: *"Bulimina trigonalis-Cibicidoides howelli (= C. lectus)-Polymorphinidae"* assemblage with *Guttulina hantkeni, G. lactea, Purulina fusiformis, Pseudopolymorphina paleocenica, Sigmomorphina brotzeni, Globulina lacrima, G. ipatovzevi, Gyroidina pontoni, Valvulineria ravni,Stensioeina parvula, Eponides lunatus, Cibicidoides howelli (= C. .lectus), Bulimina trigonalis, B. paleocenica, B. rozenkrantzi, B. constrictula, Angulogerina wilcoxensis, A. europea.* Type **4**: "*Stensioeina beccariiformis* (= *S. caucasica"*) assemblage with *Lenticulina platypleura*, *Vaginulinopsis midwayensis, V. echinata, Neoflabellina delicatissima, Dentalina eocenica, D. colei, Stylostomella paleocenica, S. midwayensis, Gyroidinoides subangulata, Stensioeina beccariiformis, Eponides plummerae, Ceratobulimina perplexa, Cibicidoides howelli, C. alleni, Bulimina midwayensis, B. quadrata, Pleurostomella paleocenica, Bolivina midwayensis*. Foraminiferal assemblages **2-4** (specially **4** type of foram fauna) are similar to Tethyan "Midway-type faunas" (Berggren & Aubert,1975). According W. Berggren and J.Aubert, this type of benthonic foraminiferal assemblages (middle-lower sublittoral zone, 50-200 m water depth) developed in a shale-marl environment of Atlantic-Tethyan regions and is characterized by textulariids, nodosariids, polymorphinids and anomalinids. According to F. Morkhven, W. Berggren & A. Edwards (1986). “Midway fauna” is the upper cosmopolitan deep-water generalized faunal assemblage and settled in neritic bathymetric zone. Type **5**: *"Nuttallides truempyei"* foram fauna with *Spiroplectammina spectabilis, Tritaxia paleocenica, T. trilatera, Gyroidinoides globosus, Stensioeina beccariiformis, Nutalinella florialis, Nuttallides truempyei, Gavelinella rubiginosa,Cibicidoides dayi,Gavelinella velascoensis, Pullenia coryelli*. This type of foraminiferal fauna ocorresponds to the Atlantic-Tethyan "Velasco-type fauna". According W. Berggren & J. Aubert(1975), and F. Morkhven, W. Berggren & A. Edwards (1986), “Velasco fauna” is cosmopolitan deep-water benthic foraminiferal assemblages (>100 m water depth, lower sublittoral-bathyal-abyssal) which developed in pseudoabyssal, continental slope, continental rise and abyssal plain and is characterized by robust anomalinids, lagenids (including nodasariids and dentalinids), small agglutinated forms (*Dorothia, Gaudryina*, and *Tritaxia*) and various gyroidinids and buliminids.

Type **6**: “*Rhabdammina"* foram fauna with *Bathysiphon dubia, B. nodosariiformis, Dendrophrya maxima, Rhizammina indivisa, Saccammina complanata, Hormosina carpenteri, Nodellum velascoensis, Rhabdammina abyssorum(=R.cylindrica), Ammodiscus glabratus, Glomospira charoides, , Rzehakina epigona,Cribrostomoides obesus, Recurvoides varius, Ammomarginulina aubertae, Adercotryma spp., Reticulophragmoides jarvisi, Spiroplectammina navarroana, Dorothia trochoides, Marssonella identata, Tritaxia trilatera, Karreriella horrida*. This type of foraminiferal fauna is named "flysch-type" or “*Rhabdammina”* fauna. This assemblage shows almost only agglutinated forams (astrorizids, hormosinids, ammodiscids, rzehakinids, litualids, spiroplectamminids, textulariids). Independent geological evidence and deep-sea drilling data indicate that this type of foraminiferal fauna has an

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extensive paleobathymetric distribution (~200 m to over 4 km). Besides the bathymetry are other factors which determine the distribution of this assemblage: relatively rapid deposition of fine-grained, carbonate-poor, dark-colored, organic and phosphorite rich sediments with fine detritus. This complex is widely distributed in flysch compartmented basins under somewhat restricted bottom water circulation and in epicontinental basins under upwelling with 0_2-poor (suboxic, dysoxic, and anoxic), PO_4-rich water (Heckel, 1977; Gradstein & Berggren, 1981). Type **7**: "Asselina-Nummulites" foram fauna. The appearance of large foraminifers: *Asselina pustulosa, A. leymeri, Operculina semiinvoluta, Nummulites crimensis*, and various *Discocyclina* forms represents distinctive feature of this assemblage. It developed in warm (tropical- south subtropical), shallow water (< 30-50 m). Type **8**: "*Clavulina anglica-Turrilina brevispina-Uvigerina batjesi*"foram fauna with species-index, and *Spiroplectammina adamsi, Gaudryina hiltermanni, Vaginulinopsis decorata, Gavelinella grosserugosa, Anomalinoides ypresiensis, Asterigerina bartoniana.* Type **9**: " *Pseudogaudryina externa- Vaginulinopsis eofragaria- Ammophilla subbotinae*" foram fauna with *Spiroplectammina tenera, P. externa- Robulus vialovi, Vaginulinopsis ex gr. mexicana, V. asperuliformis, V. eofragaria, Valvulineria intenta, Gyroidina depressaeformis, Siphonina wilcoxensis, Brotzenella pseudoacuta, Cibicidoides faraonis, Cibicidoides limarius, C. felix, Heterolepa eocaenus, A. subbotinae, Bifarina ex gr. millepunctata, Euuvigerina praecompacta* .The **8** and **9** benthonic foraminiferid assemblages corresponds approximately to the "Midway type" assemblage of Berggren and Aubert (1975), and indicates water depths of approximately 50-200 m. Type **10**: *"Aragonia"* foram fauna with *Valvulina spinosa, Nuttallides truempyei, Oridorsalis umbonatus, Bulimina tuxpamensis, B. trinitatensis, Aragonia aragonensis*, *Pullenia quinqueloba*. It is characterized by the associaton of pleurostomellids, stylostomellids, small gyroidinids, thin nodasariids, and dentalinids. This corresponds approximately to "Velasco-type" assemblage of Berggren and Aubert (1975) and suggest depths of > 200 m. It corresponds to bathyal-abyssal benthic foraminifera from the Atlantic Ocean (Tjalsma & Lochman, 1983).

These faunal types show the following turnover through time and space:

Early Danian - beginning of eustatic transgression (Haq et al., 1987). Type **1** faunas *(" Rotalia saxorum")* became widespread in the B (Mons basin) (Moorkens, 1982), and Type **2** ("*Anomalinoides danicus*") in the Dk, PO, SS, C, SP, NC, PC, M, and T (Subbotina,1947; Pozaryska,1965; Morozova et al., 1967; Pozaryska,Szczechura, 1968a; Jarzeva,1973; Maslakova & Nguen van Ngok,1989) and Type **4** ("*Stensioeina beccariiformis* ") and 5 ("*Nuttalides truempyei*") in the NC and T basins (Schutskaja, 1956, 1970). Late Danian (Montian Substage) - culmination of transgression and eastward migration of Type **1** fauna from the B through PO to the C basin (Schutskaja,1958; Szczechura & Pozaryska, 1971; Naidin & Beniamovskii,1988, 1989, 1994; Maslakova, Nguen van Ngok, 1989). Eustatic regression in the terminal Montian. Hiatuses and unconformits in sections. Disappearance of Type **1** and **2** foram faunas. Selandian stage: eustatic transgression. Two major rearrangments in fauna distribution: a) Type **6** ("*Rhabdammina*") appears in NC, corresponding there to Type **4** and **5** faunas (Subbotina,1951), b) Type **3** ("*Bulimina trigonalis-Cibicdoides howelli (= C. .lectus)-Polymorphinidae")* fauna migrates westward from the TP to PR, LV, SP, C,PO, SS, B and northward from TP to WSS (Brotzen,1948; Vassilenko,1950; Nikitina, 1962; Pozaryska,Szczechura, 1968a,1968b; Kaplan et al.,1977; Podobina, 1990; Naidin et al., 1994). In the terminal Selandian, a eustatic regressian took place, giving rise to hiatuses and unconformits (Pomerol, 1989; Naidin & Beniamovskii, 1994). Early Thanetian: eustatic transgression, expansion of marine basine (SEE basin appearans; El-Naggar, 1967), and further changes in faunal distribution. The continuous habitat of Type **3** foram fauna ("*Bulimina trigonalis-Cibicdoides howelli (= C. lectus)-Polymorphinidae")* splits into two regions: northwestern (B, Dk, SS; Brotzen, 1948; De Coninck et al.,1981) and eastern (TP and WSS; Beniamovskii at al., 1989). This separation was due to the land barrier that arose on the site of the PO, L and NU basins (Moroz, 1965; Kaplan et al., 1977;Vinken, 1988). Type 4 faunas ("*Stensioeina beccariiformis = S. caucasica*") became widespread in the C , PC, and M basins (Schutskaja, 1960; Nikitina,1962; Vinken, 1988; Naidin & Beniamovskii, 1994,1996). Type **6** foram fauna (*"Rhabdammina "*) became widespread in the NC, SP, and T basins, where it neighbors with Types 4 and 5 (Subbotina, 1951; Nikitina, 1962; Schutskaja, 1970). This event was due to upwelling from deep-water Caucasus basin to the epicontinental NC, SP, and T basins. The middle Thanetian (beginning of the *Morozovella velascoensis* Zone) is a peak of eustatic transgression, further development of upwelling, and northward spreading of Type **6** fauna into the PR basin (Naidin et al., 1994). Late Thanetian (*Morozovella velacoensis* Zone). Short-lived and areally limited transgression and beginning of large eustatic regression. Terrigenous shallow-water sections. Paleocene types of benthic forams disappear, and radiolarian-diatom assemblages became widespread in the NC, PR, TP and WSS basins (Naidin et al., 1994). Terminal Thanetian-beginning of Ypresian (top of NP9 and most of NP 10 Zones). Break in marine sedimentation almost everywhere in the EPA (Muzylev,1980; Pomerol, 1989; Steurbaut & Nolf, 1986). Early Ypresian (NP11-NP12 Zones): eustatic transgression and distribution of the next four types of foram faunas: Type 7 ("Asselina-Nummulites") in C basin, Type **8** (*"Clavulina anglica-Turrilina brevispina-Uvigerina batjesi"*) in NS, SEE, B basins (Kaasschieter, 1961; King, 1981;Vinken 1988); Type **9** (*"Pseudogaudryina externa- Vaginulinopsis eofragaria-Ammophilla subbotinae"*) in NC, SP, PR, M, TS, and T basins (Naidin et al.,1994; Schutskaja, 1956, 1970), and Type **10** (*"Aragonia"*) in NC, T where it neighbors on Type **9** (Schutskaja, 1956, 1970).

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DECOUVERTE DE BYKOVIELLA (FORAMINIFERES TROCHAMMINIDAE) DANS LES FALUNS SPARNACIENS A CYRENES DU BASSIN DE PARIS

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A la différence des Ostracodes et des fossiles palynologiques, les Foraminifères ont été rarement signalés dans les faciès sparnaciens. En quelques gisements, les niveaux les plus marins, dits "Fausses Glaises" ou argiles à *Cyrena* et *Tympanotonus*, sont cités dans la littérature pour avoir livré des Foraminifères peu nombreux, la plupart à test hyalin.

L'examen minutieux de résidus de lavage de plusieurs échantillons de Fausses Glaises à *Cyrena* et *Tympanotonus*, provenant de la carrière de Limay et d'affleurements près du Cap d'Ailly, a révélé la présence de Foraminifères agglutinés à ciment chitinoïde. Leurs tests d'abord subplanispiralés, puis déroulés, et leur ouverture aréale les font attribuer à une espèce (probablement nouvelle) du genre *Bykoviella* Korchagin, 1964. A cause de leur mauvais état de conservation et de l'écrasement de leurs tests, ces Foraminifères n'avaient pas été identifiés jusqu'à présent, ou rapportés à des *Trochammina* et/ou des *Haplophragmoides*. Les *Bykoviella* sont accompagnés de quelques *Verneuilinoides subeoceanus* (Wick, 1943) et de rares hyalins: *Rosalina* cf. *elegans* (Hansen, 1970), *Neocorbina* (n?) sp. et autres, différents dans chaque gisement.

Ces *Bykoviella* sont également présents dans les Woolwich Beds anglais, à Charlton (east of London) et à Swanscombe (Curry *in litteris*), et peut-être aussi, sous le nom d'*Haplophragmoides* et/ou de *Trochammina*, dans l'Untereozän 1 d'Allemagne du N.W.

Par l'absence de Miliolidés, sa très faible diversité et la prédominance d'une espèce (environ 95% des individus sont des *Bykoviella*), cette association témoigne d'un milieu "extrême", défavorable aux Foraminifères à la fois à cause d'une salinité diminuée oscillant autour de 8 à 10‰, et d'un déficit en oxygène.

Cette association confirme et précise l'hypothèse d'un golfe sparnacien, très profondément engagé dans les terres émergées couvertes de forêts, et ne communiquant avec l'Océan que par l'isthme de la Mer du Nord. En quelques deux Ma, ce golfe passe par plusieurs étapes, diachrones selon les endroits. C'est d'abord un lac calcaire localement occupé par des hauts fonds émergés à *Microcodium*, puis une vaste tourbière anoxique et acide. Par la suite, des arrivées successives d'eaux marines augmentent peu à peu la salinité et la teneur en oxygène de la pellicule d'eau recouvrant le golfe. C'est dans de tels environnements entourés de mangroves que s'installent les *Bykoviella* et les autres Foraminifères "extrêmophiles". L'évolution s'achève avec la transgression cuisienne, amorcée avec le dépôt de la Formation de Varengeville et celui des Sables de Laon.

Les sédiments spamaciens, intercalés dans le Bassin Parisien entre le Thanétien et le Cuisien, sont généralement considérés comme étant synchrones des biozones P5-6 et NP9-10 et du paroxysme volcanique hébridéen. Ils se situent précisément - et c'est leur intérêt majeur - dans l'intervalle de temps où doit être définie la limite Paléocène-Eocène.

CLIMATIC EVOLUTION AND ISOTOPIC EXCURSION ACROSS THE PALEOCENE-EOCENE TRANSITION AT THE SOUTHERN MARGIN OF THE TETHYS (TUNISIA)

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An integrated multidisciplinary approach to the paleo-environmental changes occurred at the Paleocene/Eocene transition, is proposed here on two sections located in Tunisia, combining micropaleontology, sedimentology, mineralogy and geochemistry (stable isotopes). Besides their economic importance (enriched phosphatic layers), these two sections because of their paleogeographical location, at the southern margin of the Tethys, constitute a clue to understand the climatic and environmental changes observed globally across the P/E transition.

The Foum Selja section is located in the restricted shallow Gafsa Basin (Fig.1) which is connected to the west-south to the Saharan Platform, but isolated from the Tethyan Realm by the Kasserine Island. The absence of direct connections with the open sea was probably due to several small uplifted areas and gulfs which may have active as bars.

The Elles section is located in the El Kef Basin, between the Kasserine Island to the South and the Algerian Promontory to the northwest. This basin is characterized by a continous deeper water sedimentation during Paleocene-Eocene time. Thus, the nature of the sedimentation at Foum Selja is dominated by alteration products coming from the Saharan Platform whereas the sedimentation at Elles is influenced by.the Tethyan region and the Kasserine Island.

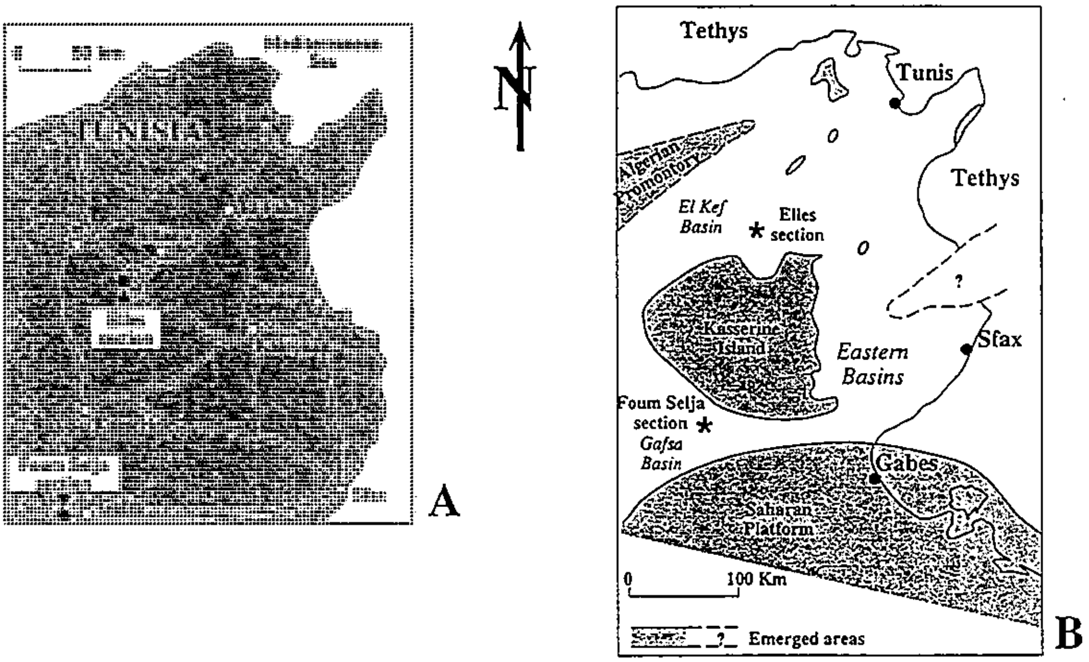

Fig.1. A. Geographical location of the Foum Selja and the Elles sections, Tunisia. B. Paleogeographic setting during the Paleocene and Eocene time. This figure is modified after Burollet, 1956; Sassi, 1974; Beji-Sassi, 1985; and Sassi et al., 1987. Note the shallow water location of Foum Selja between the Saharan Platform and the Kasserine Island and the more open location of Elles which is directly connected with the Tethyan Realm.

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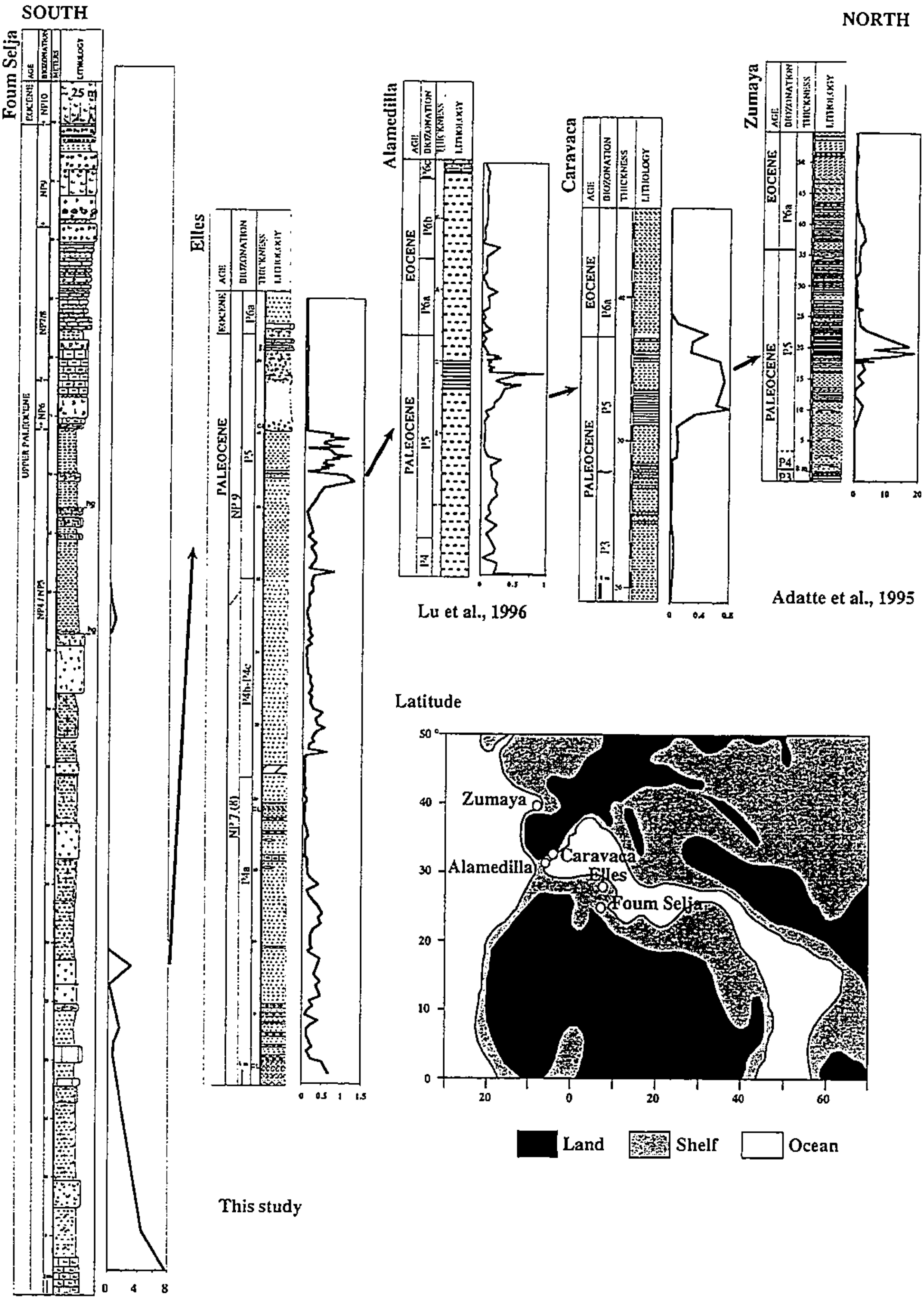

Fig.2. Kaolinite/smectite ratio distributions in the Foum Selja, Elles (Tunisia), Alamedilla, Caravaca (Southern Spain) and Zumaya (Northern Spain) sections with their location on a paleogeographical map done by Lu et al., 1996. Note the disappearance of kaolinite with a shift in time which suggests a latitudinal move of the climatic zones, from lower latitude to higher latitude.

Our data set reveal major changes across the Paleocene-Eocene transition. The Elles section exhibits a gradual decrease of 1.6 ‰ in $\delta^{13}C$ values in P4b/c and in the lower part of Zone P5 followed by a more rapid 1 permil negative excursion. This negative excursion coincides with a drastic decrease in calcite down to 7%, an important increase of phyllosilicates (70 to 85%) and the occurrence of potassic feldspar. The restricted distribution of K-feldspar and the very high detritus/carbonate ratio suggest an increase in terrigenous input across the P/E transition. These mineralogical changes probably indicate decreased primary productivity of carbonates and/or dissolution and increased erosion in the Tethyan Region at this epoch. The occurrence of phosphate deposits in the Late Paleocene, above the negative isotopic excursion, reflect the first invasion of nutrient- rich oceanic waters in the El Kef Basin during the major transgression which extends into the Eocene in all major basins.

Detrital clay-mineral compositions in marine sediments record qualitative signals of regional climatic changes. For instance kaolinite forms abundantly under warm and humid climate with intensive rainfall. Palygorskite and sepiolite indicate opposite climatic conditions. These fibrous minerals developp typically under warm but very arid conditions with enhanced evaporation. At least, smectite indicate seasonal intermediate climatic conditions or sea-level transgression. The studied sections located in two different paleogeographic area suggest two different climatic history. At Foum Selja, clay-mineral distributions reflect climatic variations on the adjacent continental area, the Saharan Platform. The abundance of kaolinite indicates for the lower part of NP4/5 warm and humid climate. The disappearance of this mineral and the high amount of smectite indicate a warm climate with enhanced evaporation on the adjacent area at the end of NP5, as indicated by the presence of evaporites. Very arid climatic conditions developp on the Saharan Platform in NP6 to reach a maximum of aridity in NP10 as indicated by the maximum of palygorskite and sepiolite. At Elles, the source of the sediments is located in the the Tethyan Region and the clay-minerals record the climatic evolution in this area. The low content of kaolinite and the abundance of smectite indicate for the NP7/8 (P4) zones seasonal conditions in the Tethyan Region whereas increased kaolinite across the Paleocene-Eocene transition implicate warm and humid climatic conditions. The absence of kaolinite, the abundance of smectite and the increase of mica indicate for the upper part of the section the development of seasonal to arid climatic conditions as at Foum Selja.

At Elles, as at Alamedilla and Caravaca (southern Spain), clay-minerals composition across the P/E transition indicate warm and humid climatic conditions coinciding with the negative carbon isotopic excursion. Kaolinite disappears abruptly in NP9. The abundance of smectite indicate increased aridity in the Tethyan Region just above the P/E transition. At Foum Selja, section located southern and influenced by the Saharan platform, kaolinite disappears already in NP4/5. The increase of smectite and the appearance of palygorskite reflect the development of arid conditions from the upper part of NP5 to NP10. From the South to the North, kaolinite disappears with a shift in time, in NP5 at Foum Selja, in NP9 at Elles (Tunisia) and Alamedilla (Southern Spain) and only in NP10 at Zumaya (Northern Spain, Atlantic Realm). This shift in time suggest thus a latitudinal shift of the source of the production of kaolinite (subtropical zone) and a consequently latitudinal move of the climatic zones, from lower latitudes to higher latitudes (Fig. 2).

Despite several uncertaincies as the exact location and the origin of the kaolinite or the particular distribution of palygorskite and sepiolite in the Tethyan Realm, the post-P/E Event is characterized by an increase of aridity in the southern Tethyan Region reflecting a shift of climatic zones, from lower latitudes to higher latitudes.

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CARBON ISOTOPE- AND MAGNETO- STRATIGRAPHY OF THE LOWER PALEOGENE CONTINENTAL SERIES OF PROVENCE (FRANCE)

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The Lower Paleogene series of the Aix-en-Provence basin consist largely of mudstones and siltstones. These fluvial deposits are interfingered by some lacustrine, palustrine and pond carbonates. The paucity of fossils in these type of sediments prevents to establish a detailed biostratigraphic framework and to propose correlations with the marine realm. A magnetostratigraphy and a carbon isotope-stratigraphy have been developed in order to overcome this lack of detailed biostratigraphic information.

Depositional setting :

The studied formation consist of 300 m of interbedded red mudstones, siltstone to fine-grained sandstones and lacustrine to palustrine carbonates. These series are well exposed throughout the present day Aix-en-Provence syncline. A Lower Paleogene section was sampled along the road D9 from Griffon area to Petit arbois.

The fluvial deposits represent in volume more than half of the sediment that outcrop in the region. Due to the abundant bioturbations and to the numerous paleosols, no original stratifications are preserved in this material. The paleosols constitute widespread horizons that could be latterally traced. On a macroscopic scale, these paleosols were identified on the base of the mottling, the presence of carbonate glaebules, slickenside features. The mineralogical composition of the floddplain alluvium is fairly stable along the studied interval. Samples devoid of pedogenic imprint show an average content of 20% in quartz, 40 to 60% in carbonate and up to 40% in clay. On this carbonate-rich parent material, the paleosols which display carbonate accumulation horizons are widely distributed. These carbonate-rich horizons represented obstacle to erosion and had great chances to be preserved.

The lacustrine facies are composed of nearly pure micritic limestone containing gastropods, ostracods and charophytes. These subaquatic facies often display dessication features and root traces, indicating that these lakes were of variable shallow depths and that the associated deposits have been periodically emerged, probably in relation to seasonal water fluctuations. The **palustrine carbonates** correspond to these shallow lake deposits that have been exposed to subaerial conditions. Identification of pedogenic imprint on these facies is based on the presence of root traces, most often surrounded by vertical structures interpreted as pedogenic stacked nodules. The mottling is faint. The upper part of the lacustrine carbonate mud is sometimes mixed with terrigenous sediments when there was a substantial detrital input. Then, the calcic soils that developed on the palustrine sediment show scattered nodules in a reddish shaly sediment and do resemble these on the fluvial alluvium.

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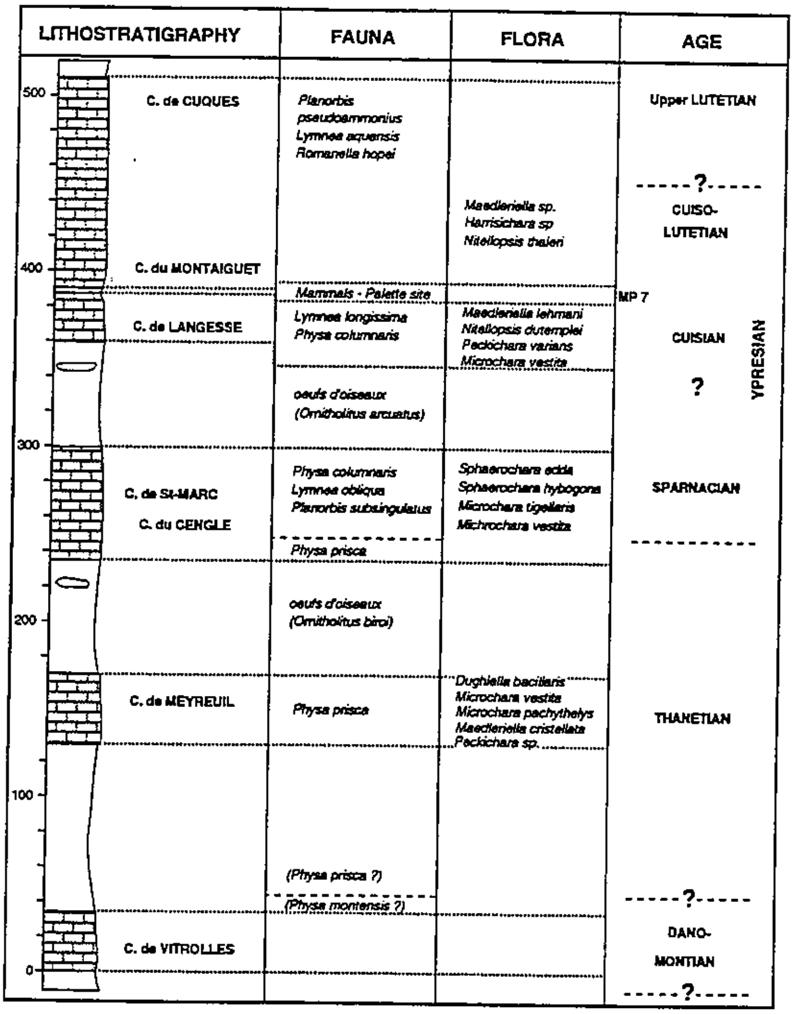

Figure 1 : Composite stratigraphic section based on the lithostratigraphy of borehole Meyreuil 2 (from Cavelier, 1984; Godinot *et al.*,1987)

The biostratigraphy

The biostratigraphic framework that has been established for the Paleogene series of the Aix-en-Provence basin is presented on a composite section (Fig. 1) (Cabvelier, 1984). The distribution of fossils is closely related to the depositional facies and no single section could provide a complete stratigraphic information.

The lacustrine facies are the more favourable for fossil preservation. The abundant gastropods, charophytes are grouped in several assemblages that are characteristics of the Thanetian up to the Upper Lutetian. In the red siltstones, the only fossils that were found are birds eggs. The first type spans most the Thanetian while the second one corresponds to the upper Thanetian - Spamacian.

One site contains fossil bones of mammals. It is located in a marly layer that is intercalated between the Calcaire de Langesse and the Calcaire du Montaiguet (Palette site). The faunal forms suggest for Palette an age close to that of the Dormal reference level (MP 7 - Cuisian) and probably even older (Godinot *et al.*, 1987).

The calibration of this stratigraphic scale with the marine record is difficult due to the fact that most of the beds do not contain stratigraphic elements to allow correlations.

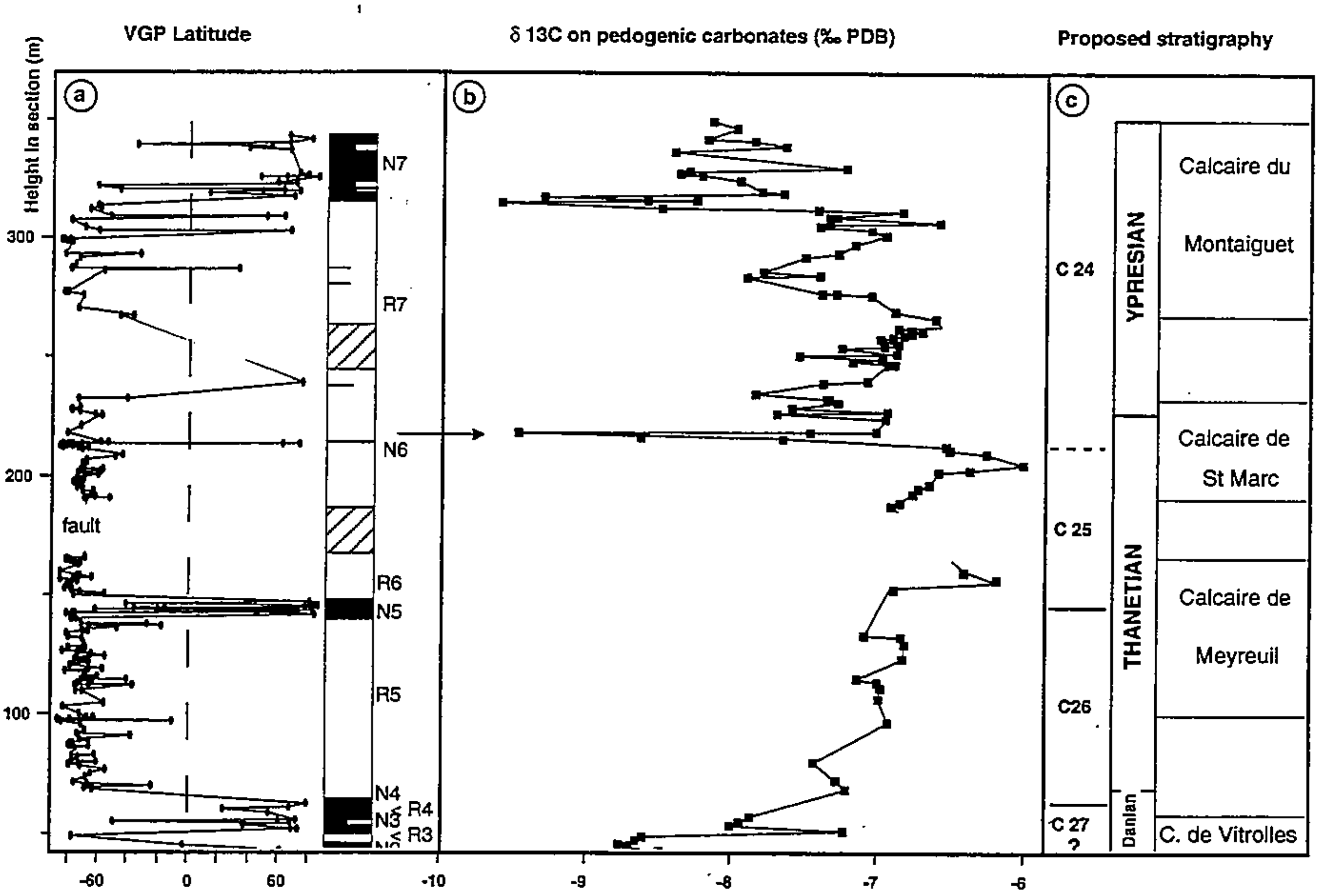

Figure 2 : The studied Lower Paleogene section in the western part of the Provence basin
a) magnetostratigraphic results plotted as VGP latitudes versus height and interpreted polarity log,
b) carbone isotope data plotted versus height,
c) Proposed stratigraphy as constrained by rhe bio-, chemo- and magnetostratigraphy.

The magnetostratigraphy

The lower Paleogene section on the western part of the basin has been sampled, by coring 2,5 cm diameter plugs, regurlarly spaced by 1m. Two to four samples were collected from each of the over 200 sampling levels. for all sections, magnetic carrier with high coercivity are found. All samples were thus subjected to a stepwise thermal demagnetization.

The paleomagnetic behaviour ofdifferent sampled facies ranges from very good to poor. Despite the weak intensity of magnetization, the palustrine facies proved to give the best signalbecause they do not bear any diagenetic overprint. On the contrary, for many samples in the red siltstones, the natural remanent magnetization contains several components probably due to pedogenic alterationsor later fluid circulations. It is not possible to define their primary component.

A virtual geomagnetic pole is calculated from the charactéristic directions. The stratigraphic plot of VGP latitudes shows a pattern of magnetozones (Fig. 2a). Correlation of our data to the geomagnetic reversal scale is constrained by the given biostratigraphy :

- sequence N2 to N4 is difficult to interpret due to the lack of biostratigraphic control
- sequence R5/N5/R6/N6 is attributed to Chron 26 and to the base of Chron 25 from the Thanetian faunas. The normal interval of chron 25 has been identified by two samples

- sequence R7/N7 is attributed to Chron 24 on the base of the Ypresian fauna.

The carbon isotopic stratigraphy

The δ13C has been measured on the pedogenic carbonates. The sampling interval is in average 2m but was constraint by the vertical distribution of the paleosols. The pedogenic nodules were collected from both facies that contain carbonate-rich paleosols : the floodplain alluvium and the palustrine sediments.

The nodules were cleaned to remove all the particles of clayey material that could have remained before being crushed for the isotopic analysis of the carbone and oxygen. Samples from close-by horizons showed that the carbone isotpic signature of the pedogenic nodules from palustrine facies or the floodplain alluvium are very similar and that they are clearly different from these of the lacustrine carbonate mud. The carbone isotopic stratigraphy has been constructed with data from both types of pedogenic nodules (Fig 2b).

Results and discussion

The Early Cenozoic period is very suitable for this type of integrated stratigraphy because 1) the paleomagnetic scale shows several reversals and 2) the marine carbon isotopic record is marked by a long-term shift from low values in the early Paleocene to higher values in the late Paleocene and then by another shift towards values that are 2 to 3‰ lower in the Early Eocene. Superposed on these long-term shifts are some distinct negative excursions that are also well recorded on a global scale (Kennet and Stott, 1990).

The carbone isotope- and magneto-stratigraphy built from the continental series of Provence are well in accordance with the marine data.The long term evolutions are very comparable in term of duration and of amplitude :

- The Early Paleocene carbone isotope data show a long term shift toward more positive values that are reached in the Late paleocene (from chron 26 to 25 with an amplitude around 2,3 ‰).
- the Late Paleocene carbone isotope data show fairly negative values (δ13C≈8, chron 24) with a peak towards more negative values by the base of 24N (amplitude of the peak is around 1‰).

The C13 spike associated with the late Paleocene thermal maximum has been identified in the lower part of chron 24, around 220m. The amplitude of the spike is around 3‰ and very comparable to the marine data.

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TIMING, MECHANISM AND LOCATION OF MASSIVE CARBON DIOXIDE INPUT DURING THE LATEST PALAEOCENE : ANATOMY OF A RAPID GLOBAL CARBON ISOTOPE EXCURSION

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A brief time interval at (or near) the end of the Palaeocene Epoch ca. 55 Ma known as the late Palaeocene thermal maximum (LPTM) was marked by a 4 to 8 °C increase in deep ocean and high-latitude temperatures, and major turnovers in terrestrial and marine flora and fauna. Upheaval of the global carbon cycle also occurred during the LPTM: carbon isotope records of carbonate and organic matter display a prominent excursion in $d^{13}C$ of at least -2.5 ‰, and deep sea sediment sequences are characterized by an interval of pronounced carbonate dissolution and oxygen deficient facies. Observed carbon cycle perturbations are best explained by massive input of CO_2 enriched in ^{12}C to the ocean or atmosphere with concurrent removal of O_2 in the ocean.

Recent manuscripts by Bralower et al. (1997), Pardo et al. (1997) and Schmitz et al. (1997) highlight a fascinating aspect of global change and CO_2 input during the LPTM. These authors state, in accord with previous suggestions (Kennett and Stott, 1991; Canudo et al., 1995), that onset of the $d^{13}C$ excursion occurred within 5 to 25 x 10^3 years. This interpretation is of great interest. First, the interpretation makes the $d^{13}C$ excursion an ideal marker for the P/E Boundary. Second, the interpretation makes the LPTM a past analog for future changes in the global carbon cycle. An excursion of at least -2.5 ‰ in all major carbon reservoirs within 5 to 25 x 10^3 years necessitates "catastrophic" addition of CO_2 at rates approaching anthropogenic inputs of fossil fuel.

There is, however, a major dilemma. The given interpretation of the $d^{13}C$ excursion assumes that some natural process actually can add massive quantities of carbon enriched in ^{12}C to the global carbon cycle over very short time intervals. Such a process does not exist in conventional models of the carbon cycle (Sundquist, 1986). Topics discussed here are: (1) a potential process for substantial input of CO_2 in the past, and (2) evidence for this process during the LPTM.

Thermal dissociation of gas hydrate provides a mechanism for massive input of CO_2 during abrupt ocean warming (Figure 1; Dickens et al., 1997). Gas hydrates are crystalline substances composed of gas and water that are stable at high pressure and low temperature. Gas hydrates with a composition approximating $CH_4 \bullet 6H_20$ and carbon isotopes greatly enriched in ^{12}C ($d^{13}C$ of -60 to -70‰) exist in uppermost sediment in many regions of the ocean (Kvenvolden, 1993). The 4 to 8 °C rise in bottom water temperature during the LPTM should have shifted sediment geotherms and increased subbottom temperatures. This heating would have caused dissociation (melting) of gas hydrate. Presumably, free CH_4 gas released during dissociation could escape from sediment sequences to the ocean through sediment failure. Methane then would be oxidized to CO_2. The net effect of this process would be input of CO_2 greatly enriched in ^{12}C to the ocean (and atmosphere), deep sea carbonate dissolution, and removal of dissolved O_2 in the ocean.

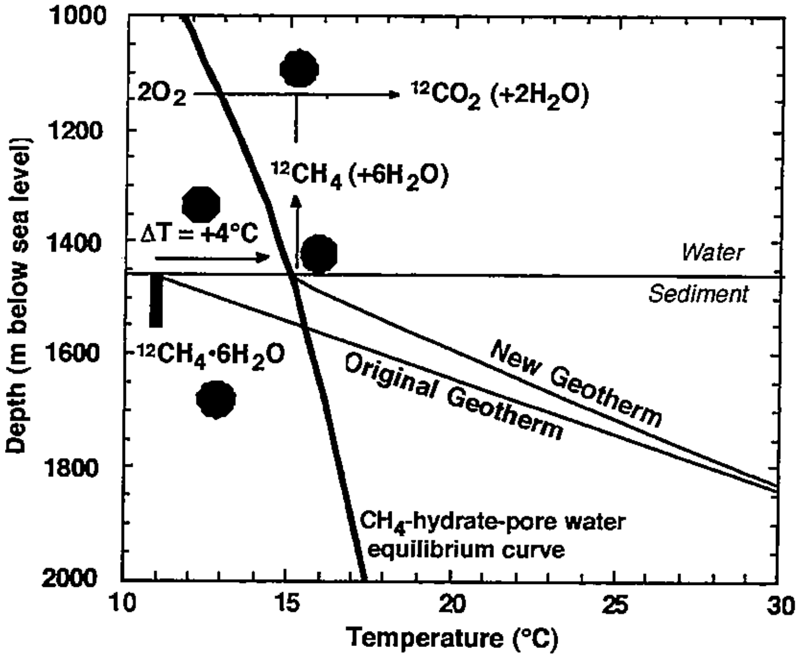

Figure 1. Methane hydrate stability conditions for a hypothetical sediment sequence at intermediate water depth in the late Palaeocene and fate of CH_4 in this sediment sequence after a rapid warming of bottom water temperature by 4 °C during the LPTM. (A) Prior to the LPTM, solid CH_4 hydrate ($CH_4 \cdot 6H_2O$) greatly enriched in ^{12}C is restricted to a shallow sediment zone between the sediment water interface and the intersection of the geotherm and a CH_4-hydrate-water equilibrium curve. (B) Rapid warming during the LPTM induces a shift in sediment geotherms. (C) Warming of sediment causes dissociation of solid CH_4 hydrate to free CH_4 gas greatly enriched in ^{12}C (and H_2O) which then escapes to overlying water by sediment failure. (D) Oxidation of CH_4 in the water column removes dissolved O_2 and produces CO_2 greatly enriched in ^{12}C. Depth axis on this figure assumes a hydrostatic pressure gradient. The equilibrium curve on this diagram is for pore water with an activity similar to that of seawater.

Mass balance considerations, simple heat propagation calculations, and numerical model results indicate that a scenario involving about 10% of the estimated present-day gas hydrate reservoir (1 to 2 x 10^{18} g CH_4 with a $d^{13}C$ of -60‰) is a plausible explanation for interpreted carbon cycle perturbations during the LPTM (Dickens et al., 1997). Moreover, there is a good argument that massive release of CH_4 to the deep ocean occurred during the LPTM. ODP Site 690 in the South Atlantic at intermediate water depth has $d^{18}O$ and $d^{13}C$ records of benthic and planktic foraminifera that span the LPTM (Kennett and Stott, 1991). Two important observations in these records are: (1) a negative $d^{18}O$ excursion in benthic foraminifera precedes a negative $d^{13}C$ excursion in all foraminifera, and (2) the negative $d^{13}C$ excursion in thermocline dwelling foraminifera precedes the negative $d^{13}C$ excursion in surface dwelling foraminifera. Thus, isotope records at Site 690 indicate deep sea warming during the LPTM preceded massive input of ^{12}C enriched CO_2, and that input of CO_2 occurred first in deep and intermediate marine environments rather than in the atmosphere or shallow marine environments. Isotope records at Site 690 (assuming they are not a post-depositional artifact) are difficult to explain by mechanisms other than thermal dissociation of gas hydrate and release and oxidation of CH_4 in deep and intermediate waters. No process exists in conventional models of the carbon cycle to rapidly add massive quantities of CO_2 into the ocean after deep sea warming and into the deep ocean prior to the shallow ocean.

Benthic and planktic $d^{13}C$ records at the same location provide an important independent cross-check on interpretations of rapid CO_2 release during the LPTM. Input of ^{12}C enriched CO_2 (or CH_4) into the deep ocean will result in a temporal offset between $d^{13}C$ records of deep marine and shallow marine carbon reservoirs. The reason is simple: exchange of carbon between deep and shallow ocean reservoirs involves thermohaline circulation and significant

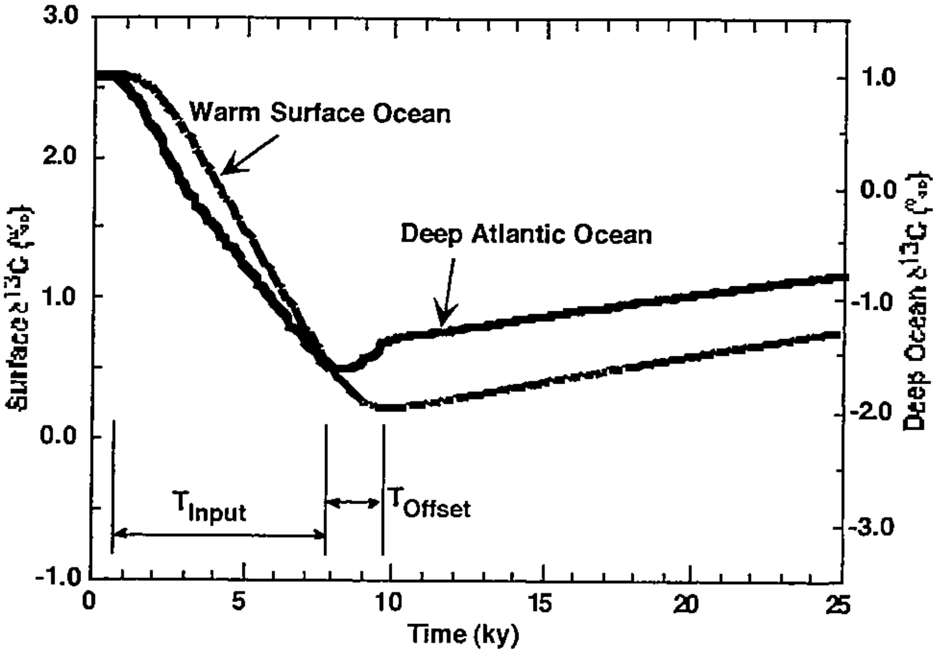

Figure 2. Expected effect on d^{13}C of SCO_2 after release and oxidation of 1.1 x 10^{18} g CH_4 with a d^{13}C of -60‰ in the present-day deep Atlantic Ocean over 7.5 x 10^3 years. A temporal offset of approximately 2 x 10^3 years should occur between negative d^{13}C excursions in the deep Atlantic Ocean and warm surface water. Moreover, the d^{13}C excursion will occur in the deep Atlantic Ocean prior to warm surface water because this is the specified location of CH_4 oxidation. Note that the time offset between the two d^{13}C records (2 x 10^3 years) is a significant fraction of the time represented by the onset of the isotope excursion (7.5 x 10^3 years). Results were obtained using a slightly modified carbon cycle model by Walker and Kasting (1992).

time frames (Sundquist, 1986; Walker and Kasting, 1992). For example, addition and oxidation of 1.1 x 10^{18} g CH_4 with a $d^{13}C$ of -60‰ in the present-day deep Atlantic Ocean over 7.5 x 10^3 years should result in a negative $d^{13}C$ excursion in the deep Atlantic Ocean of approximately 2.4‰ that precedes a similar magnitude negative $d^{13}C$ excursion in the shallow Atlantic Ocean by approximately 2 x 10^3 years (Figure 2). Excursions in $d^{13}C$ records of benthic and planktic foraminfera at Site 690 occur over a distance (10 to 20 cm) about two to three times the distance of the offset between $d^{13}C$ records of thermocline dwelling and surface dwelling foraminifera (6 to 7 cm). Assuming that exchange fluxes between various carbon reservoirs were similar in the late Palaeocene and present-day, and assuming that isotope records at Site 690 have not been extensively bioturbated, dissolved or winnowed, the observed offset between the two $d^{13}C$ records is consistent with massive input of CO_2 into a deep or intermediate water carbon reservoir over a relatively short time interval (within 8 x 10^3 years).

Massive release and oxidation of CH_4 from submarine gas hydrates satisfactorily explains fundamental conceptual problems with current interpretations of carbon cycling during the LPTM. However, such a scenario must be regarded as speculative. Amounts of CH_4 and potential carbon transfer mechanisms in present-day gas hydrate reservoirs are poorly understood. Quantitative estimates regarding Palaeocene carbon exchange fluxes, carbon reservoir masses, lysocline depths, and dissolved oxygen contents have not been made.

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THE PALEOCENE-EOCENE BOUNDARY INTERVAL OF THE CHANNEL COAST (NW PARIS BASIN) MAIN EVENTS AND STRATIGRAPHICAL INTERPRETATION.

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Near its connexion with the Dieppe-Hampshire basin, along the Channel coast, the NW part of the Paris basin exhibits several Tertiary outliers resting on the Upper Cretaceous chalks. They yield Upper Paleocene and Lower Eocene deposits encompassing the P/E boundary interval (NP8 - NP11) in which the continental, brackish and marine "Sparnacian" facies are especially well exposed.

We present here a SW-NE cross section concerning the most important of these outliers from the Varengeville-Dieppe region to the Montreuil surroundings. To have a link with the more northern P/E facies, the cross section was extended up to the St Omer outcrops.

All these sites are known since the 18th century (see Leriche, 1939 for reference of ancient works) and often studied after. In the past few years observations were possible in scarce quarries now abandoned. More recently boreholes and road-cuttings related to the A16 highway construction brought new data. But the most interesting source of information still remains the Varengeville, Dieppe and Criel cliffs (South western part of the cross section). Paleocene-Lower Eocene deposits are largely exposed along several hundred meters sometimes kilometers. They are also more or less regularly cleaned and renewed by landslides favouring detailed and occurate studies.

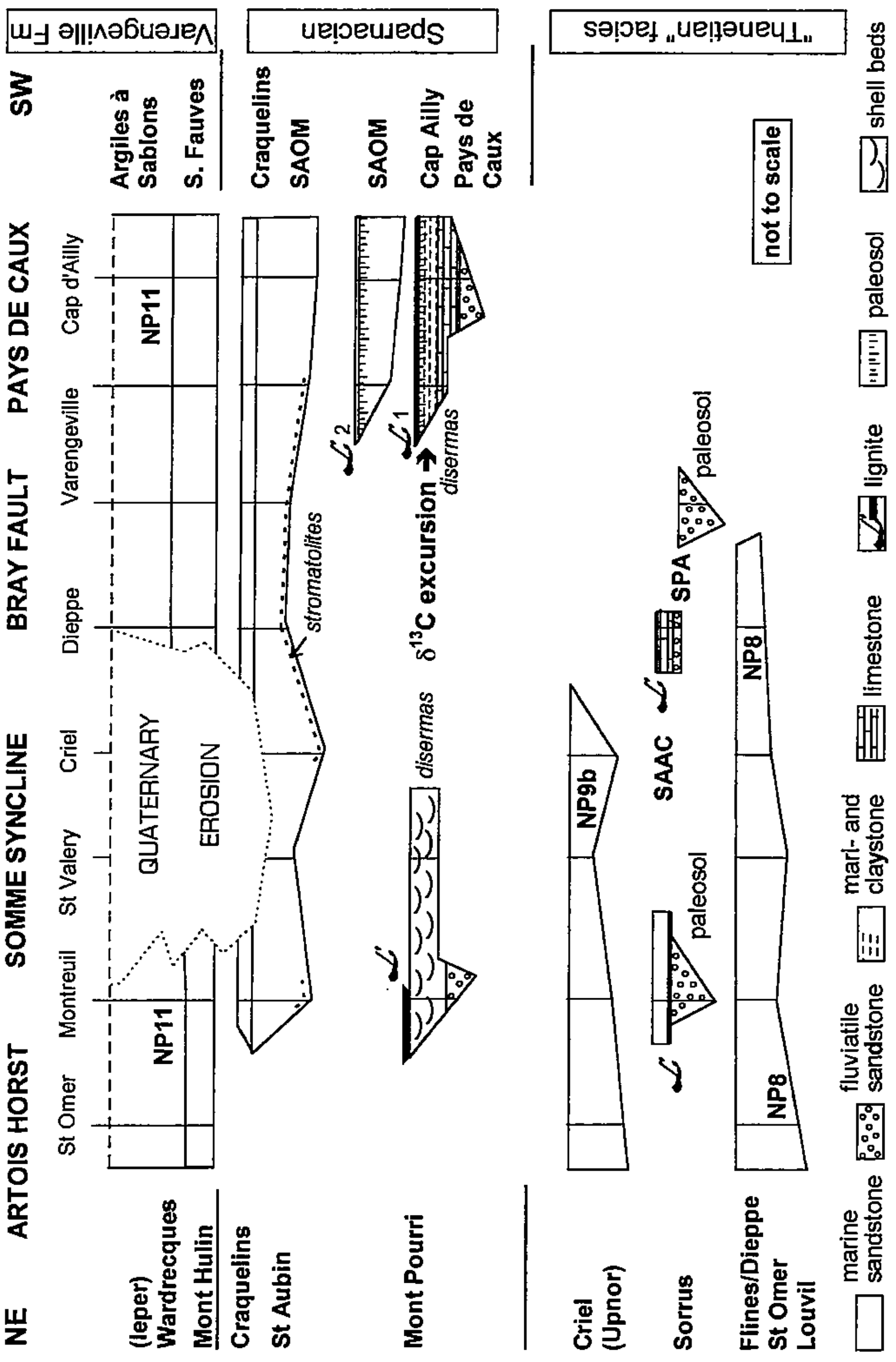
The cross section and the detailed succession of the Varengeville outlier (which will be shown during the field trip) offer to describe the main events of the P/E interval, especially the lateral and vertical facies variations linked to the relative sea level changes. Results of mineralogical studies, ^{13}C geochemistry and significant biostratigraphic information (calcareous nannoplancton, Dinoflagellates, Charophyta) allow to put these events in a global perspective.

Three main sets of facies can be distinguished. In ascending order : 1-the Thanetian facies, 2-the "Sparnacian" and 3-the Formation de Varengeville (Leriche, 1939).

1. The Thanetian facies comprise two transgressive sheets of glauconiferous sands separated by a drastic relative sea level fall (type 1 sequence boundary) inducing the erosion of large channels which were infilled by a transgressive aggradation of continental-brackish deposits (Sorrus, Sables et Argiles à Annélides de Caudecote, S.A.A.C., Sables du Petit Ailly, S.P.A.). This sea level fall seems linked to a widespread tectonic event locally strong enough to induce an angular unconformity.

The marine units are respectively attributed to the NP8 nannozone (Louvil, St Omer, Flines, Dieppe) and to the NP9 nannozone (Upnor, Criel) (Dupuis and Steurbaut, 1987). Both show complicated sea level changes, especially the uppermost one (Steurbaut *et al.*, in progress).

2. The "Sparnacian" began with another relative sea level fall promoting the erosion of a second generation of large channels subsequently infilled with transgressive continental facies (Mont-Pourri : silts à *Unio* and *Polymesoda cordata*, Cap d'Ailly : calcaires lacustres, marnes et lignite L_1,...). Diversified associations of Charophyta, belonging to the middle part of the *Peckichara disermas* charozone were collected in these deposits. The $\delta^{13}C$ negative excursion (~ - 27‰ on plant organic matter) was found in the lignite L_1 and in the top of the underlying paleosol (Sinha, 1997).



Stratigraphic scheme of the P/E boundary interval of the Channel coast (NW Paris Basin). Three main sets of deposits are distinguished: The Thanetian facies, the Sparnacian and the Varangeville Formation. The two first ones associate marine to continental deposits. The third one is almost marine.

Smaller scale units are constructed on a lithologic basis with local biostratigraphic control. The Cap d'Ailly and Mont Pourri units are correlated according their Charophyta content belonging to the middle part of the *Peckichara discermas* zone.

The interpretation of the relationships of these units allow to trace and suggest the main hiatuses wich are especially long in some tectonic areas: Artois horst, Bray fault and Pays de Caux high.

Above, two sequences of brackish-marine sediments rest with minor erosion on the previous units (type 2 sequence boundary). They are made from an alternation of clays, silts, sands and coquinas (SAOM : Sables et Argiles à Ostracodes et Mollusques, Bignot, 1965; Dupuis and Steurbaut, 1987) and contain abundant fauna indicating brackish to marine paleoenvironments (*Corbicula* spp., *Tympanotonos*, *Ostrea*, etc).

The first unit of SAOM ends within a thin lignitic layer resting on a paleosol interpreted as the expression of a highstand in continental environment.

The second unit of SAOM is capped by the Argile glauconieuse des Craquelins which indicates a fully marine environment yielding for instance a diversified association of Dinoflagellates (in progress). A lag of small flint gravels and sparse shark teeth underlines the maximum fooding surface. The top of this unit become progressively sandy and deeply bioturbated and is interpreted as a highstand. This upper part of the Sparnacian contains a form of the Dinoflagellate *Nematosphaeropsis* which can be compared with ? *Nematosphaeropsis* sp. by Jolley and Spinner (1989) found in a very short interval in the Harwich Member near Ipswich probably correlatable with the NP10 biozone (Deconinck, in preparation).

The widespread extension of the continental-brackish Sparnacian facies may be related to the regional uplift associated with the onset of the Island plume and with the Groenland-Rockall plateau separation (Knox, 1997).

3. The Formation de Varengeville. We will only describe here the lower part of the Formation de Varengeville and its relationships with the underlying Sparnacian. The Sables Fauves form the well defined base of the Formation de Varengeville. They are made of several meters of fine glauconitic sands with large cross-stratifications characteristic of a shoreface paleoenvironment. The contact with the underlying Sparnacian is sharp and shows minor erosion below local small channels located at the base of the unit in which rare accumulations of well rounded black flint pebbles known as "galets avellannaires" were observed locally. They are correlated since a long time with the Oldhaven sands (Whitaker, 1871). We interprete this unit as a shelf margin wedge.

The overlying units of the Formation de Varengeville, the Argiles à Sablons, the Argile brune à concrétions and the Argile Glauconieuse du Phare (NP11, Lezaud, 1967; Aubry, 1983) correspond to three sequences with an increasing transgressive tendency towards the top.

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EARLY EOCENE BENTONITES FROM TWO SECTIONS OF THE EASTERN ALPS (AUSTRIA)

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In the area of Salzburg montmorillonite-layers (bentonites) of Paleogene age occur in two different tectonic units. In the north the Anthering section is part of the Rhenodanubian flysch whereas in the south the Untersberg section is located within the Northern Calcareous Alps. These were overthrusted on the flysch in the early Oligocene. This north verging thrusting shortened the distance between the two sections to a range of 15 km. Originally this distance must have been much longer.

Within the Anthering-section the dominating rocks are turbiditic marls which alternate with hemipelagic claystones. The latter prove a depositon of this flysch below of the local CCD. In a 40 m thick part of the section around 30 bentonite layers were found which are seen as tuffs of air-fall derivation.The bentonite layers display thicknesses from 2 mm to 3 cm.They are totally converted to smectitic clay. Immobile element concentrations point to a alkali basaltic composition of the original magma. A few of the thicker layers, however, show concentrations typical for highly evolved alkaline ashes. The ash-bearing sequence is restricted to the lower part of nannoplankton-zone NP 10. It ends immediately before the first appearance of Tribrachiatus contortus. The age of the bentonites and their chemical compositions point to a correlation with bentonites of the North Sea region. According to HEILMANN-CLAUSEN at Anthering the dinoflagellate cyst Apectodinium augustum has its last appearance (LAD) approx. 25m beneath of the oldest bentonite layer. In the North Sea region the LAD of this species is within the "negative" ash series. Therefore the ash-bearing sequence of Anthering may be synchronous with the upper part of the "negative" ash series and/or the "positive" ash series in the North Sea region.

At the base of the Untersberg-section 1.5m of red claystone occur. This claystone might represent the time of the benthic extinction event (BEE) within the nannoplankton-zone NP9. At the top of the claystone 0.5m of clayey marls form a transition zone to nonturbiditic marls which are dominating in the rest of the section. In the clayey marls coccoliths are rare (10% or less) and the nannoplankton assemblages mainly consist of Rhomboaster cuspis (up to 49%), Discoaster multiradiatus (up to 35%) and Discoaster falcatus (up to 17%). This composition of the nannoplankton assemblages is interpreted as an effect of carbonate solution which led to an secondary enrichment of species less prone to dissolution. Approx. 25m above of the top of the basal claystone the first bentonites appear already within NP10. Unfortunately only the base of the ash bearing sequence is exposed there. Within 5m of marl 15 bentonite layers were observed. These layers display the same immobile element concentration as the coeval layers at Anthering. So there is little doubt that the bentonites of both sections belong to the same eruptive volcanic events. Usually the layers of this southern section are thicker than those at Anthering which is a discrepancy to the assumed origin of the ashes from the north. However, these greater thicknesses could be explained by higher contents of calcareous plankton which was deposited together with the ashes of the Untersberg-section.

INTEGRATED BIOSTRATIGRAPHY AND ISOTOPE GEOCHEMISTRY ACROSS THE PALEOCENE/EOCENE BOUNDARY IN A CLASSICAL TETHYAN SETTING: THE CONTESSA-ROAD SECTION

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In the Umbria-Marche Basin the Paleocene/Eocene (P/E) Boundary is comprised in the pelagic carbonate rock succession of the Scaglia Rossa Formation which is suitable for precisely correlating the Paleogene biostratigraphy to geomagnetic polarity history. In particular, magnetostratigraphic, lithostratigraphic and biostratigraphic investigations in the Gubbio Area have confirmed and dated the geomagnetic reversal for most of the Paleogene interval (Lowrie et al., 1982; Napoleone et al., 1982) while Corfield et al. (1991) reported a good carbon isotope record including a marked negative shift of $\delta^{13}C$ across the P/E boundary from the Contessa Highway section.

However, small tectonic disturbance and/or the presence of covered intervals have hampered a detailed study of the P/E transition in this classical Tethyan setting. In the Contessa Highway section a 1 m-thick covered interval prevents a continuous sampling across the P/E Boundary which is, in turn, involved in a slump in the Bottaccione section.

To overcome these problems the Contessa-Road (CR) section (Fig. 1) has been re-sampled across the P/E Boundary.

Although a ca. 1.7-m thick duplicated interval is present below the P/E the section provides a continuous record across it.

Morevover, this section provides a more complete magnetostratigraphic record across the P/E Boundary than Contessa Highway and Bottaccione (see Lowrie et al., 1982).

An integrated bio- and chemo-stratigraphic analysis has been carried out including foraminiferal and calcareous nannofossil distribution, and carbon and oxygen isotope record across the P/E boundary at the CR section.

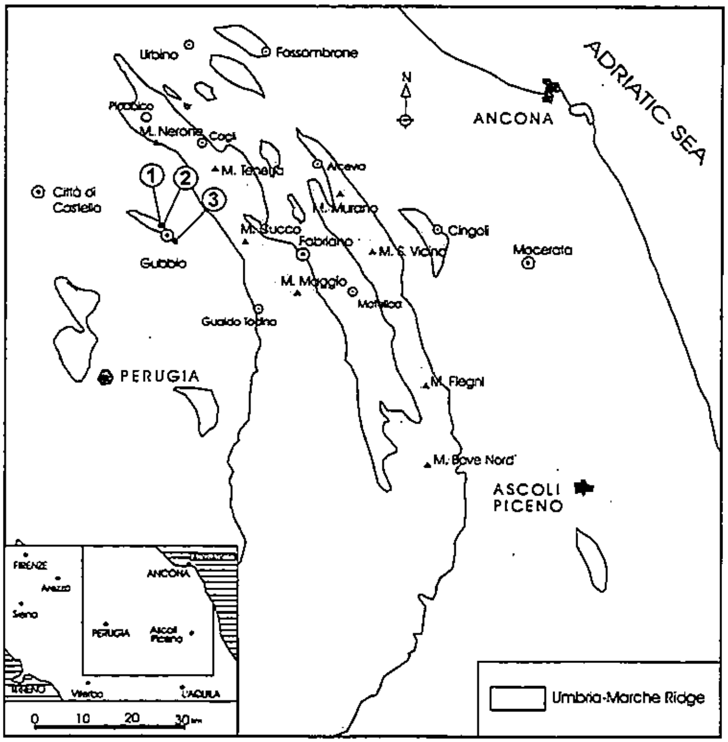

Fig. 1 Location map of the Contessa Highway (1), Contessa-Road (2) and Bottaccione (3) sections

Taxononomic composition of benthic foraminiferal assemblages suggests that the Scaglia Rossa limestones of the CR section deposited in a lower bathyal environment at a paleodepth of 1500-2000 m.

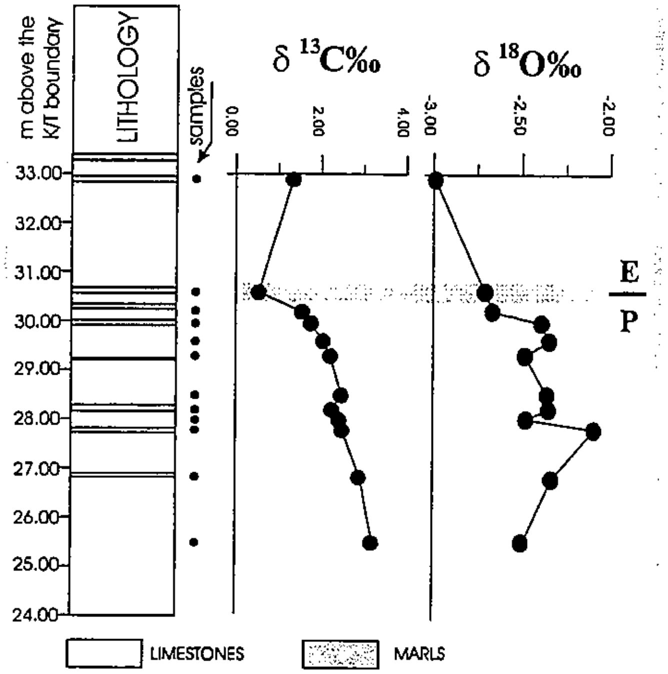

Fig. 2 - $\delta^{13}C$ and $\delta^{18}O$ record acros the P/E boundary at the Contessa-Road section

The lowest value in $\delta^{13}C$ (0.549‰) is recorded at 30.60 m (Fig. 2). This negative peak is preceded by a gradual decrease from a maximum of 3.189‰ recorded at 24.50 m and is followed by a recovery up to 1.336‰ at 33.90 m. The lowest value recorded at 30.60 m is here intepreted to correspond to the well known negative shift of the carbon isotopic values occuring globally close to the end of the Paleocene. Accordingly, we place the P/E boundary at 30.60 m. In the same inteval the $\delta^{18}O$ curve shows a trend to lighter values, which however, is strongly affected by diagenesis.

Distribution of calcareous nannofossil well fits with the position of the P/E boundary at 30.60 m. The First Occurrence (FO) of *Discoaster multiradiatus* marks the NP8/NP9 and CP7/CP8 zonal boundaries at 27.40 m (Fig. 3). A strong decrease of *Fasciculithus* is recorded 1 m below the P/E boundary. The NP9/NP10 zonal boundary is placed at 29.80 m at appearance level of *Rhomboaster bramlettei* whereas the FO of *Discoaster diastypus* marks the CP8/CP9 zonal boundary at 32.00 m.

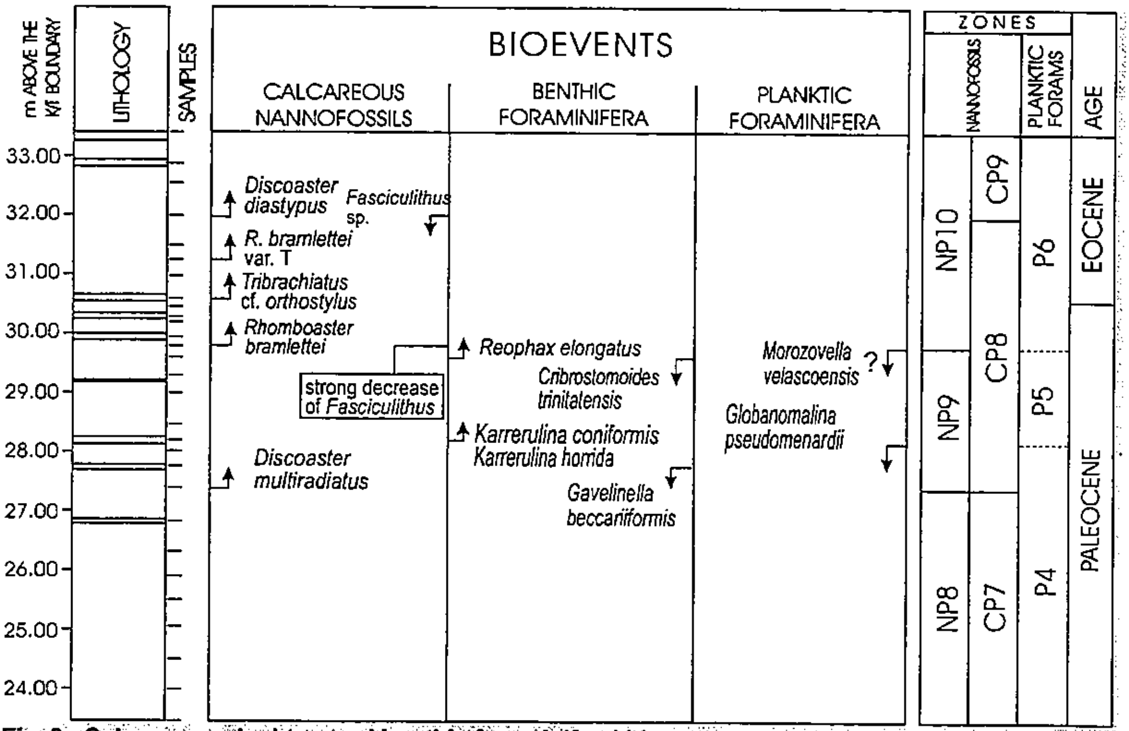

Fig. 3 - Calcareous plankton and benthic foraminiferal bioevents across the P/E boundary at CR

Concerning benthic foraminifera a biostratigraphic control on the P/E boundary position has been carried out on the marly layers where it was possible to check the presence of calcareous-hyaline taxa including the markers of the P/E transition (see Berggren & Miller, 1989). The last occurrence of *Gavelinella beccariiformis* which marks the BB1-BB2 zonal boundary of these authors, is recorded at

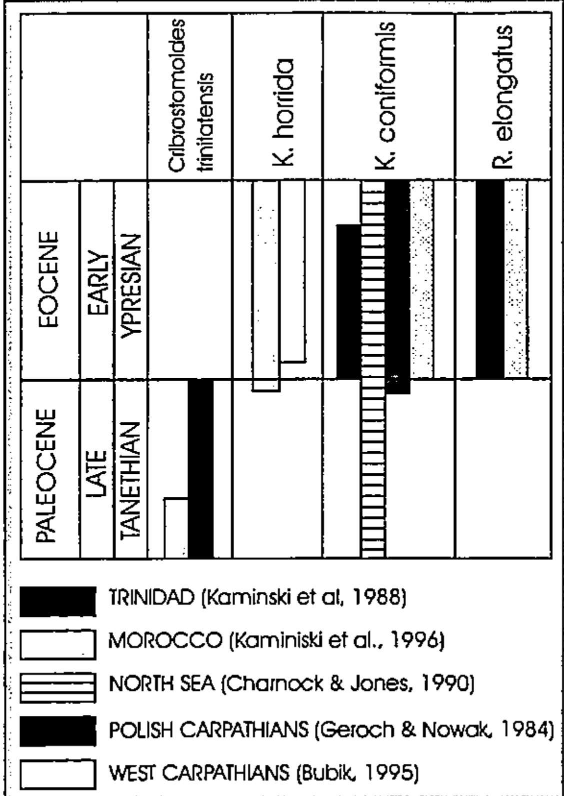

Fig. 4: Distribution of selected DWAF across the P/E boundary from different sites

27.45 m (fig.3). However, this cannot be considered the true Last Occurrence (LO) of this taxon due to its scattered distribution in the CR section.

The LO of *Morozovella velascoensis* at 29.60 m is also doubtful occurring 1 m below the δ^{13}C negative shift. However, planktic foraminiferal distribution analysis is still in course.

Across the P/E boundary a succession of discrete events within DWAF assemblages is recorded. *Karrerulina coniformis* and *K. horrida* first occur at 28.20 m whereas *Reophax elongatus* at 29.60 m where *Cribrostomoides trinitatensis* last occurs (Fig. 3). As indicated in Fig. 4 these taxa show stratigraphically comparable FOs and LOs from other sites.

According to the distribution of DWAF from the CR section an environmental perturbance affected the sea-floor across the P/E boundary as testified by the faunal parameter record (Fig. 5).

The late Paleocene is characterised by a bloom of tapered/flattened morphotype mainly due to increasing abundance of the infaunal opportunistic *Spiroplectammina spectabilis* in the middle part of zone NP9. At the same time the incoming of shallower water taxa belonging to the family *Textulariacea* such as *Clavulinoides asperus* occurs. Interestingly these taxa have been shown to disappear temporarily from the shallower water setting of the Trabakua Pass section (Coccioni et al., 1994) at the same stratigraphic level.

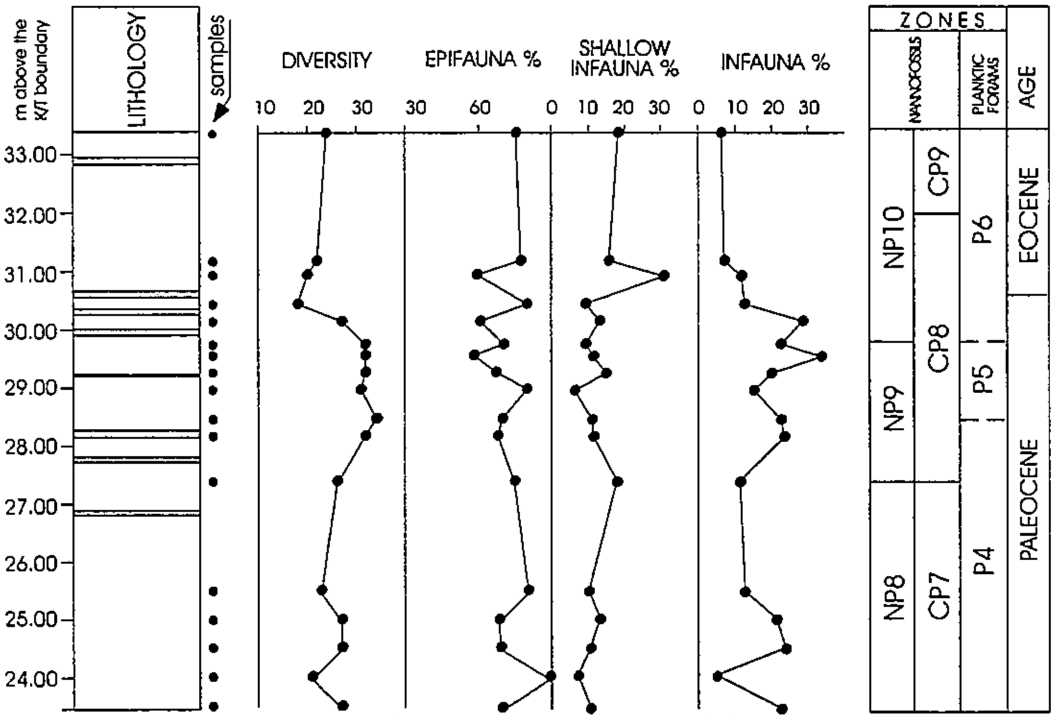

Fig. 5 - DWAF faunal parameter record across the P/E at the CR section

The possible downward migration of benthic foraminferal taxa from middle bathyal to lower bathyal environment is here interpreted as the response to a late Paleocene warming of bottom waters which in the CR section resulted in an increase of diversity values among DWAF assemblages prior to the P/E boundary (Fig. 5).

According to a sedimentation rate of 7 m/Ma estimated for the upper Paleocene/lower Eocene limestones of the Scaglia Rossa Formation this trend started some 150 ka prior to the P/E boundary.

Above the P/E boundary a bloom of *Glomospira* is recorded in the CR section that is comparable to the onset of assemblages with common ammodiscids reported from the lower Eocene in other sites (e.g. Alpine Carpathian region, Moroccan flysch, and Alpine Flysch Zone fo Switzerland)..

Some modern species of *Glomospira* appear to be environmentally tolerant, thriving under low oxygen conditions. The onset of a *Glomospira* Biofacies in the lower Eocene may, therefore, be the response to decreased oxygenation of the sea-floor that possibly caused a drop of species diversity as recorded in the CR section just above the P/E boundary.

However, despite the decrease of species diversity and the well-documented extinction event among calcareous hyaline taxa, there is little taxonomic turnover in the DWAF assemblages of the CR section across the P/E boundary

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THE PLESIADAPIDAE ACROSS THE PALEOCENE-EOCENE BOUNDARY IN THE PARIS BASIN

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Genera of the family Plesiadapidae (Mammalia, Plesiadapiformes) have been used in biostratigraphy of Paleocene-Eocene continental strata in North America and Europe. Russell (1964) considered *Plesiadapis tricuspidens* as one of the eight Cernaysian species having close relatives in the North American Tiffanian, and one of the three among them to show some affinity with Clark Fork species. Gingerich (1976) proposed a formal biozonation of North American Land Mammal Ages based on plesiadapids (2 zones in the Torrejonian, 5 in the Tiffanian, 1 or 2 in the Clarkforkian; each of these zones being named after a *Pronothodectes* [Torrejonian] or a *Plesiadapis* [other ages] species). Gingerich worked out a phylogeny of the European species different from that of Russell, and a European plesiadapid-based zonation: a *P. tricuspidens* zone, an unsampled zone for the end of the Thanetian, and in the Sparnacian three zones named after *P. russelli*, *Platychoerops daubrei* and *Plat. richardsonii.*

Gingerich also proposed a tentative correlation of plesiadapid zones of North America and Europe, with *P. cookei* (Clarkforkian) being equivalent to *P. russelli*, implying that the Clarkforkian was largely of Sparnacian age, i.e. Eocene. Later, Rose (1981) introduced a *P. gingerichi* zone covering the end of the Tiffanian and the beginning of the Clarkforkian, restricted the *P. cookei* zone to the middle Clarkforkian, and distinguished a third Clarkorkian zone, the *Phenacodus-Ectocion* zone (during which the long lived *P. dubius* survives, entering the base of the Wasatchian). Rose maintained in 1981 the time equivalence of Clarkforkian and early Sparnacian, but this correlation was later abandoned by most North American colleagues in favor of a return to Wasatchian-Sparnacian (Gingerich, 1989; Krause & Maas, 1990). Hooker (1994) described a new species of *Platychoerops* which will be discussed below.

We report here new plesiadapid material which complements our knowledge of this group in Europe and has biostratigraphical implications. More material of Meudon allows a better description of *P. russelli* (Russell et al., 1988) and good material has recently been found in the new Creil locality (Ploëg et al., this volume), which appears to be a new species. These two species will provisionally be referred to *Platychoerops*, a choice which could be discussed.

The new Creil *Platychoerops* species shows the following characters. I1/ with a well formed laterocone, a mediocone reduced in ventral view but salient in medial view, no centroconule and crest. Upper molars increasing in size posteriorly, having a strong and salient mesostyle; paraconule decreasing in size from M1/ to M3/, metaconule reduced and postprotocrista lowering from M1/ to M3/; small hypocone on M1-2/, strong continuous lingual cingulum on M3/. The P4/ have a variable paraconule on the preprotocrista, and vary markedly in their degree of crenulation. On the lower molars, the trigonid is broad on M/3, with a subrectilinear protocristid, less so on M/2, and narrow on M/1. There is a large anterolingual opening of the talonid basin on M/3, narrower on M/1-2; there is a postmetacristid with incipient metastylid. The P/4 have a metaconid well separated from the protoconid, and the paraconid varies from a low cuspule at the base of the metaconid to a small cusp well separated from it, linked or not with the protoconid. Many of these characters show that this species is closer to *P. tricuspidens* than *Plat. russelli* and *daubrei* are.

The new Meudon material allows a more complete description of *Plat. russelli*, with the following characters distinguishing it from the Creil species : I1/ with laterocone and mediocone vestigial or lacking, posterocone smaller; slightly larger size for all jugal teeth; on M1-2/, the protocone is higher and the postprotocrista steeper; on M1/ paracone and metacone are higher, with steeper labial walls; P4/ with metacone slightly better isolated from the paracone; P3/ broader in its lingual part; M/3 with a stronger lingual cusp on the third lobe, isolated from the hypoconulid by a notch, and with a more vertical postmetacristid; M/1 and M/2 with a broader trigonid, and labial slope of the protoconid more vertical; M/2 with postcristid more rectilinear.

Added to its larger size, and higher and steeper cusps, *Plat. russelli* from Meudon is clearly advanced over the Creil assemblage in its degree of reduction of I1/ tip cusps. The new Creil species appears to partly fill the gap between *P. tricuspidens* and *Plat. russelli*, giving additional evidence in favor of the lineage hypothesized by Gingerich (1976). We will scrutinize if no character precludes this *Platychoerops* lineage to be rooted in *P. tricuspidens*, and try to evaluate what the morphological distances between these species could suggest in terms of time intervals.

Compared with the preceding hypothetical lineage, *Plat. georgei* from Try appears to have a peculiar I1/, anteroposteriorly shorter and with broad posterocone. Hooker (1994) wondered if it was a side-branch or could fit in a *P. tricuspidens - Plat. russelli* lineage, also comparing its peculiar incisor characters with those of *Chiromyoides*. The morphology of the Creil species, being intermediate between *P. tricuspidens* and *Plat. russelli* with the same type of elongated incisor having a narrow posterocone, strongly suggests that *Plat. georgei* is a side-branch. To what species belong the Dormaal M2/ and the Erquelinnes I/1 referred by Hooker to *Plat. georgei* is not clear. Also, *Plat. georgei* being a side-branch does not clarify if this species belongs to the Paleocene-like or to the Eocene-like part of the Try assemblage (if the assemblage is mixed; Louis, 1996), or if it comes from a yet unsampled late Thanetian or early Sparnacian level, probably older than Creil.

The biostratigraphical consequences of our analysis are :

1) A confirmation through the *Platychoerops* lineage that Meudon is by far not an ealiest Sparnacian fauna. Hooker (1994) already drew a similar conclusion from the presence in Meudon of MP 8-9 marker taxa.

2) The formal proposition of a succession of fossil localities in the Paris Basin across the Paleocene-Eocene boundary : Cernay, Berru, Creil, Meudon, Mutigny.

3) The Creil locality clearly documents for the first time a level close to that of Dormaal in the Paris Basin. It includes such MP 7 taxa as *Teilhardina* and *Landenodon* (Ploëg et al., this volume). *T. belgica* had been listed in Meudon but a more detailed study showed that another genus was present.

4) Further work is needed to estimate time intervals, however the morphological distinctness of the known plesiadapid species suggests that more intermediate levels would be recognized if they were sampled. The preceding succession of levels/localities is incomplete.

Concerning the environment, it is interesting to note that the new Creil *Platychoerops* is dominant in the Creil fauna, as *P. tricuspidens* is in Cernay and Berru. Other dominant species are condylarthrans in Cernay and Berru, and a perissodactyl in Creil, showing that the major faunal turnover had occurred. The dominance of a medium-sized plesiadapid in these localities shows that Paleocene- Eocene boundary events did not immediately displace all faunal elements.

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LES OSTRACODES ET LA LIMITE PALÉOCENE-ÉOCENE

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La fin du Paléocène est marquée, au moins en Europe du Nord, par une régression marine à la fin du Thanétien et l'instauration d'un régime lagunaire au Sparnacien (argiles à lignites, fausses glaises, Woolwich beds, Landénien supérieur). Il est intéressant de considérer les conséquences éventuelles de ces événements sur les ostracodes dont certains sont inféodés à des milieux marins épinéritiques détruits par la ou les régressions de la fin du Paléocène et d'autres à des milieux non **directement** affectés, marins plus ou moins profonds.

Remarque: sur les deux tableaux présentés, les ostracodes sont rangés **selon leur ordre de disparition** et secondairement selon leur ordre d'apparition, ce qui facilite la mise en évidence d'éventuelles "crises" et permet de mieux apprécier leur importance. Notons, néanmoins, que la précision des datations , sur le tableau 1, n'est que de l'ordre de l'étage.

I Les Ostracodes des mers épicontinentales au passage Paléocène-Eocène

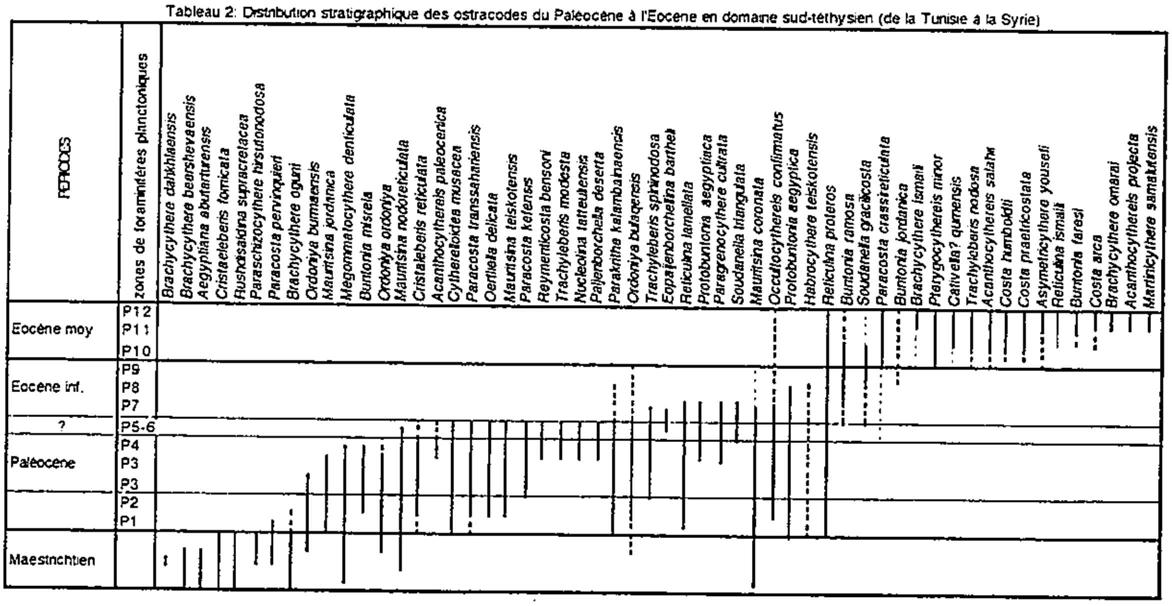
Deux exemples sont envisagés: celui des bassins du NW de l'Europe qui émergent à la fin du Thanétien et celui de la partie sud de la Téthys occidentale.

1 L'exemple du nord-ouest de l'Europe (tableau 1)

Etablis pour les bassins de Bruxelles et de Londres, les tableaux seraient sensiblement identiques, avec, majoritairement les mêmes espèces, du Thanétien au Lutétien. Le fait marquant est que la **quasitotalité des espèces marines paléocènes disparaissent définitivement à la fin du Thanétien**, les deux ou trois espèces qui semblent survivre étant d'ailleurs mal définies et donc relativement peu significatives. Les espèces qui apparaissent au Cuisien - considéré ici synonyme d'Yprésien - sont, pour la plupart, encore présentes dans le Lutétien lui-même caractérisé par un accroissement de la biodiversité. Le Sparnacien, par ses ostracodes lagunaires, se rattache au Paléocène plutôt qu'à l'Eocène inférieur.

2 L'exemple du sud de la Téthys (tableau 2)

Tableau 2: Distribution stratigraphique des ostracodes du Paléocène à l'Eocène en domaine sud-téthysien (de la Tunisie à la Syrie)



d'après Barsotti, 1963, Salahi, 1966, Bassiouni, 1969, 1970, Said, 1978, Boukhary et al., 1982, Donze et al., 1982, Mohamed, 1982, Bassiouni & Luger, 1990, Foster et al., 1983, Boukhary et al., 1993, Keen et al., 1 Andreu, 1990

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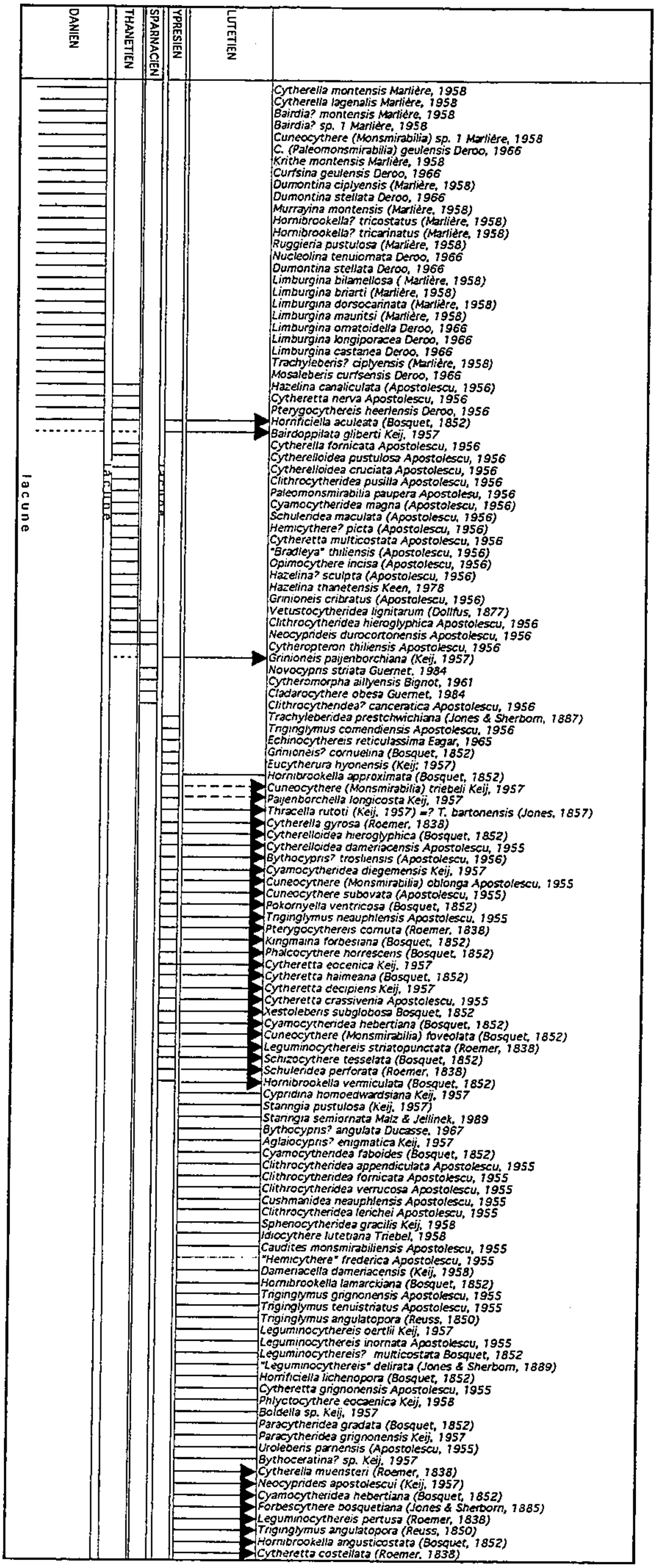
Tableau 1: Distribution stratigraphique des ostracodes du Paléocène à l'Eocène moyen dans le Bassin de Paris



d'après Apostolescu, 1954, 1955, 1956, Keij, 1957, 1958, Marlière, 1958, Deroo, 1966, Keen, 1972, 1978, Guernet, 1984, Ngatse, 1985, Malz & Jellinek, 1989, Liebau, 1991 et observations personnelles

Comme dans le NW de l'Europe, le renouvellement des espèces d'ostracodes est important au passage Paléocène-Eocène mais il semble moins brutal, peut-être en partie par ce que les biozones P5 et P6 ne sont pas distinguées par les auteurs.

II Les Ostracodes des mers profondes au passage Paléocène-Eocène

Là aussi, 2 exemples sont présentés, l'un dans les Cordillères Bétiques concernent des milieux modérément profonds, l'autre correspond aux abysses ou à l'infrabathyal de l'Océan Atlantique.

1 Les ostracodes circalittoraux ou épibathyaux au passage Paléocène-Eocène dans les Cordillères Bétiques (Espagne). A Caravaca, près d'Alicante, alternent des marnes à foraminifères planctoniques et des calcaires ou calcarénites à Nummulites et autres grands foraminifères. La limiteThanétien-Yprésien est marquée par un hiatus de la P5. Parmi les 14 espèces d'ostracodes recueillis au milieu ou au sommet de la P4, 11 montent dans la P6 et 7 de celles-ci jusque dans la P7. Il est donc possible que, dans des milieux profonds de quelques dizaines de métres ou plus, les ostracodes soient peu affectés par les événements fatals à ceux qui vivaient à faible profondeur (Guernet & Molina, 1997).

2 Les ostracodes des milieux océaniques infrabathyaux ou abyssaux au passage Paléocène-Eocène. Il existe peu de données sur les ostracodes crétacés ou paléocènes provenant de sondage océanique profond mais, compte-tenu des ostracodes provenant de séries sédimentaires actuellement émergées et supposées d'origine profonde, il semble que la plupart des genres et espèces psychrosphériques apparaissent à l'Eocène inférieur ou moyen (Whatley & Coles, 1991, Guernet, 1991, Whatley, 1993). Les ostracodes paléocènes semblent peu diversifiés et hérités du Crétacé; leur disparition, très progressive, s'échelonne jusqu'à la fin de l'Oligocène (Whatley & Coles, 1991). La limite Paléocène-Eocène est donc marquée principalement par des apparitions, principalement au cours des biozones P5 et P6.

Conclusion générale

Les **mers épicontinentales** du Paléogène étaient peuplées d'une faune adaptée à des conditions très particulières et notamment à des fonds à la fois bien éclairés et peu agités. En phases régressives, les milieux néritiques se réduisent aux plateaux continentaux contigus aux mers profondes où ces conditions favorables ne régnent que très localement. Les **régressions marines de la fin du Paléocène** sont donc une explication suffisante à l'**hécatombe** d'organismes néritiques observée dans le nord-ouest de l'Europe, notamment à celle des ostracodes. Cette hécatombe est d'autant plus spectaculaire qu'à l'épisode marin du Thanétien font suite dans ces régions une régression et un épisode continental à lagunaire, celui du Sparnacien. Les faunes téthysiennes épicontinentales semblent moins affectées, peut-être parce que, dans une région en tectogénèse, ont localement été ménagées des zones de faible profondeur protégées des vagues. Enfin, en domaine marin profond où le passage Paléocène-Eocène semble marqué plus par des apparitions que par des disparitions, **l'eustatisme positif éocène et l'ouverture progressive de l'Atlantique vers l'Océan Arctique ont facilité ou permis l'arrivée d'eaux froides polaires bien oxygénées au fond de l'Atantique - la mise en place de la psychrosphère** (Guernet, 1993) - **et donc la colonisation des fonds océaniques.**

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(pour une bibliographie compléte, se reporter à celles des travaux répertoriés ci-dessus)

THE EXPLOSIVE RADIATION OF THE GLIRES (MAMMALIA) AT THE PALAEOCENE/ EOCENE BOUNDARY. FORTUNE OF THE RODENTIA

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In recent years the discovery in the early Palaeogene of Central Asia of the fossil orders Mixodontia and Mimotonida has given a paleontological basis to the Glires concept, which embraces also the Recent orders Rodentia and Lagomorpha.The fossil forms of the superorder Glires include small to medium sized animals (mouse to squirrel) possessing a peculiar masticatory apparatus : the evergrowing chisel-like incisors are separated by a large diastema from grinding jugal teeth ; orbits are situated vertically to the jugal tooth row. For contributing to a better understanding of the beginning of the early Glires radiation, and particularly that of the early Rodentia, two aspects of this new knowledge are emphasized: the geographical and chronological distribution of early Glires as given by the fossil record, and the diversity of the chewing function as observed through the study of the wear facets of their molars. It appears that the Glires originate in Central Asia and the oldest representative remains the unique Late Cretaceous so-called "MgMidae". In the Middle and Late Palaeocene of Mongolia one more family of Mimotonida and 3 families of Mixodontia have been recognized. These animals do not spread out of Asia. In the Early Eocene first representatives of Lagomorpha and Rodentia can be identified. If Lagomopha do not spread out of Asia until the Late Eocene, from the earliest Eocene Rodentia are known not only in Asia but also in North America (Clarkforkian), Europe (Spamacian) and North Africa (Early Eocene) ; concerning their diversity, 4 families can be recognized in the earliest Eocene and 11 in the late Early Eocene.

The study of the wear facets of molars in the 12 species of Glires (Mimotonida, Mixodontia, Rodentia) found in the Bumbanian (early Eocene) of Tsagan Kushu locality (Mongolia) illustrates the diversity of the chewing function of early Glires. Four types of chewing can be observed: 1) the sectorial type is found in the Rodentia genus *Alagomys* and in primitive Mixodontia with developed lateral BB6 facet. The upper molars of *Alagomys* have very sharp cusps arranged in a triangular pattern, and probably this early rodent had an insectivorous diet. 2) the "sciurid" type is found in Rodentia Cocomyidae. 3) The ecdental type, i.e. lagomorph type, with developped BB2 facet, is found in some derived Mixodontia and Mimotonida. 4) The propalineal type, i.e. murid type, is realized in the minute Rodentia *Ivanantonia* .

In conclusion it appears that: 1) Central Asia is the probable craddle for primitive Glires; 2) very early in their adaptative history, Glires have realized different types of chewing; 3) in the early Eocene the murid sized and omnivorous early Rodentia are diversifying and spreading in all continents (with the noteworthy exception of South America); 4) on the opposite, the squirrel-sized and herbivorous Mixodontia, Mimotonida and Lagomorpha do not spread out of Asia; 5) the time "window" for the early dispersion of Rodentia is a few millions years wide: between the early to late Ypresian.; 6) most of the subsequent diversification of the Rodentia at the suborder and family level seems a consequence of vicariant conditions which prevail in the succeeding epochs; 7) with an average rate of 2.8 families/Ma and 7.3 genera/Ma during the early Ypresian to early Lutetian period (-55 to -49 Ma), the early radiation of Rodentia can be considered as explosive.

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THE PALEOCENE/EOCENE TRANSITION IN DENMARK

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During the Thanetian and Ypresian Denmark was located in an offshore, relatively deep- water position within the North Sea Basin. The stratigraphical record is, therefore, very similar to that of the central North Sea.

The talk will give an overview of these strata which include the magnificently exposed Fur Formation with nearly 200 ash layers preserved in a diatomite and representing the main phases of the North Atlantic volcanism. Also successions very similar to the upper Lista Formation, the Sele and Balder Formations, and the Ypresian red clays of the North Sea can be seen at outcrop in Denmark.

The stratigraphical column is more complete and yet lithologically simpler than the shallower, more coastal records known from Belgium, England and Northern France. The completeness of the strata is evidenced by the full succession of North Sea dinoflagellate zones.

Major lithological changes that can be observed in Denmark reveal profound shifts especially in basinal circulation. Such changes occurred twice: at the base of the laminated clay unit of the Ølst Formation (equiv. to laminated shales of the Sele Formation) and at the base of the Ypresian red clays of the Røsnæs Clay (both shifts appear to have been stepwise). The changes coincide with considerable turnover in the dinoflagellate assemblages and are associated with widespread submarine unconformities which, however, appear to have been of short duration.

A detailed correlation between Denmark and the shallow marine to continental succession of Belgium will be presented. The correlation is based on dinoflagellates and suggests that the changes in Denmark coincide with major sequence boundaries in Belgium.

The complete absence of calcareous micro- and nannofossils in NP10 and parts of the NP9 and NP11 chronozones in NW Europe has allowed dinoflagellates to become the primary biostratigraphical tool at the Paleocene/Eocene transition. They have proved their value in the offshore hydrocarbon exploration, but are also useful for correlations between offshore and coastal settings, as exemplified by the correlation between Denmark and Belgium.

Prominent dinoflagellate events in the Paleocene/Eocene transitional interval of NW Europe include the acme of *Apectodinium* spp., the LAD of *Apectodinium augustum* near ash layer no. - 19b (i.e., below the main volcanic phase) and the FAD and rapid radiation of the *Wetzeliella- Charlesdowniea* group (with the distinct species *W. meckelfeldensis* and *C. crassiramosa*) in shallow as well as offshore settings in the early phase of the 'classical' Ypresian transgression (basal Ieper Clay and basal Røsnæs Clay).

The LAD of the morphologically extreme species *Apectodinium augustum* has recently been recorded near the NP9/NP10 boundary in Austria (Heilmann-Clausen & Egger, in prep.), allowing for the first direct biostratigraphical correlation of the Paleocene/Eocene boundary strata of the North Sea Basin with the Tethys Ocean. The correlation seems to agree with the previous identification of ash layer no. -17 near the base of NP10 in the North Atlantic.

It appears, perhaps contrary to their reputation, that dinoflagellates may be useful also for interbasinal correlations at the Paleocene/Eocene transition.

SHALLOW BENTHIC FORAMINIFERA AT THE PALEOCENE-EOCENE BOUNDARY

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The activities in the frame of the IGCP 286 project have produced a revision of the correlations between biozones based on benthic, mainly larger foraminifera, magnetostratigraphy and the planktonic standard zonations (Serra et al. 1997 in press, Bull. Soc. Geol. Fr.). The shallow-benthic biozones are called now by numbers with the prefix SBZ (Shallow Benthic Zonation) to stabilize their denomination in view of the necessary changes in the Linnean nomenclature of numerous index fossils. The transtethyan correlation reaches from the shores of the gradually opening Atlantic Ocean in the Pyrenean basin to the so-called Lockhartia Sea in the Indian Ocean as exposed in the Salt-Range and Kohat-Basins (Pakistan) and the Himalayans. The transatlantic correlation between the tethyan and the caribbean realm continues to be difficult.

According to these correlations, the biotic turnover of shallow benthic communities seems not to correlate with any possible Paleocene-Eocene boundaries based on planktic or deeper benthic organisms as proposed so-far. The closest in time shallow biotic turnover is interpreted as the result of evolutionary processes according to the laws of population genetics in the communities involved, favouring in a first phase the diversification of moderate K-strategies at a generic level during the Thanetian stage and, in a second step, the diversification of species belonging to a restricted number of successful genera at the beginning of the Ilerdian stage. The two events are dated as having taken place between SBZ 2 and 3, during P4a and between SBZ 4 and 5 during Early P5a respectively (Fig. 1).

The diversification of the species was linked to a considerable increase of shell size and adult dimorphism which is interpreted as an adaptation to seasonality in oligotrophic warm (tropical to subtropical) environments (Fig. 2). Thus, the shallow benthos hints at a stability of the shallow environments during the Paleocene-Eocene boundary epoch admitting evolutionary change by adaptation to predictable environmental conditions, as they are necessary for K-strategists.

In recent tropical environmental change, there must be a decoupling of shallow from deeper benthic environmental conditions because the biogeographic patterns of photic and dysphotic benthic communities do not match. Therefore, an ecological stability during the Paleocene-Eocene boundary period in shallow water does not mean that the tethyan oceans of that time had stable conditions in all their depth zones.

Strata Série 1., vol. 9 p. 61-64, 2 fig. 1998

From a practical point of view (for mapping for instance), the collaborators in IGCP 286 recommend to fix the Paleocene-Eocene boundary in shallow carbonate deposits at the level of the second turnover event which can be easily recognized by the microfacies documenting the rise of larger-sized and distinctly dimorph foraminifera in the field with a hand lens: All the well-known limestones with larger porcelaneous forms, i.e. orbitolitids, alveolinids, lacazinids and/or the lamellar-perforate, involute nummulitids (s.str.) would fall within the Eocene. This second turnover event coincides with a major, worldwide transgression corresponding to the TA 2.1 - TA 2.2 sequence boundary (Pignatti, 1991).

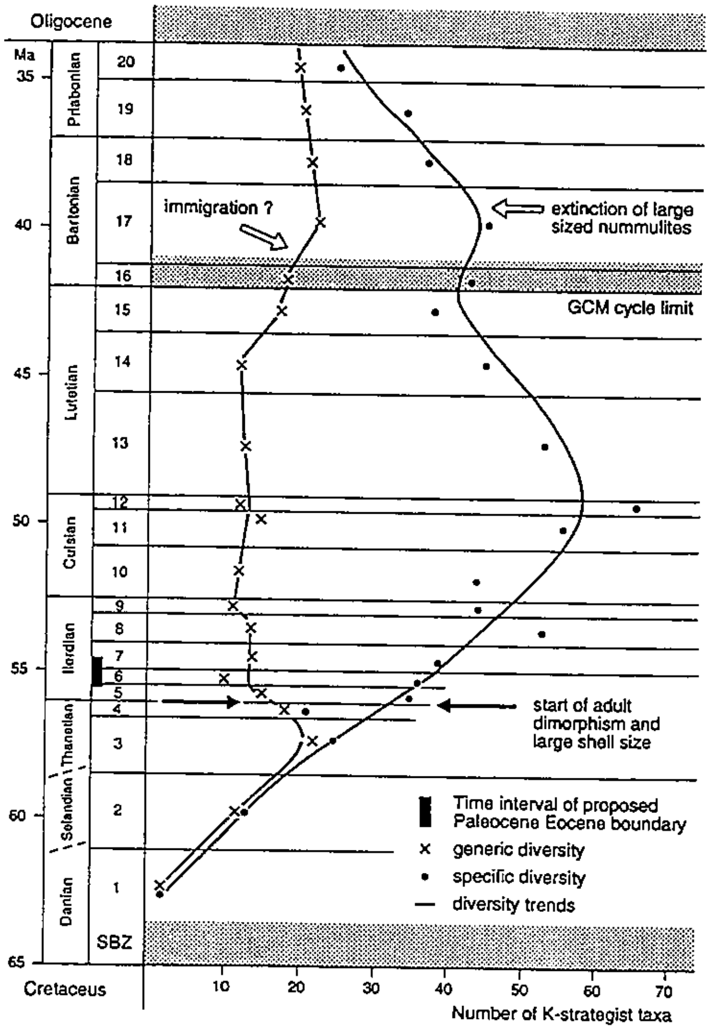

Fig. 1. Generic and specific diversity of larger foraminiferal K-strategists during Paleocene and Eocene.

Time calibration according to Serra et al. 1997. Number of taxa exclusively according to explicit data base given by the same authors where 61 genera and 411 species are documented. Similar diversity curves constructed by Hottinger (1990) for the two genera *Assilina* and *Nummulites* and their phyla according to Schaub (1981) show comparable trends. Therefore, the shape of the curves given here is not expected to change significantly when the present data base on all major groups will be corrected, completed or refined in the near future. Note the different starting conditions for the two Global Community Maturation (GCM) cycles (Hottinger, 1997) during Paleocene-Eocene and the truncation of the second one by the Eocene-Oligocene boundary events.

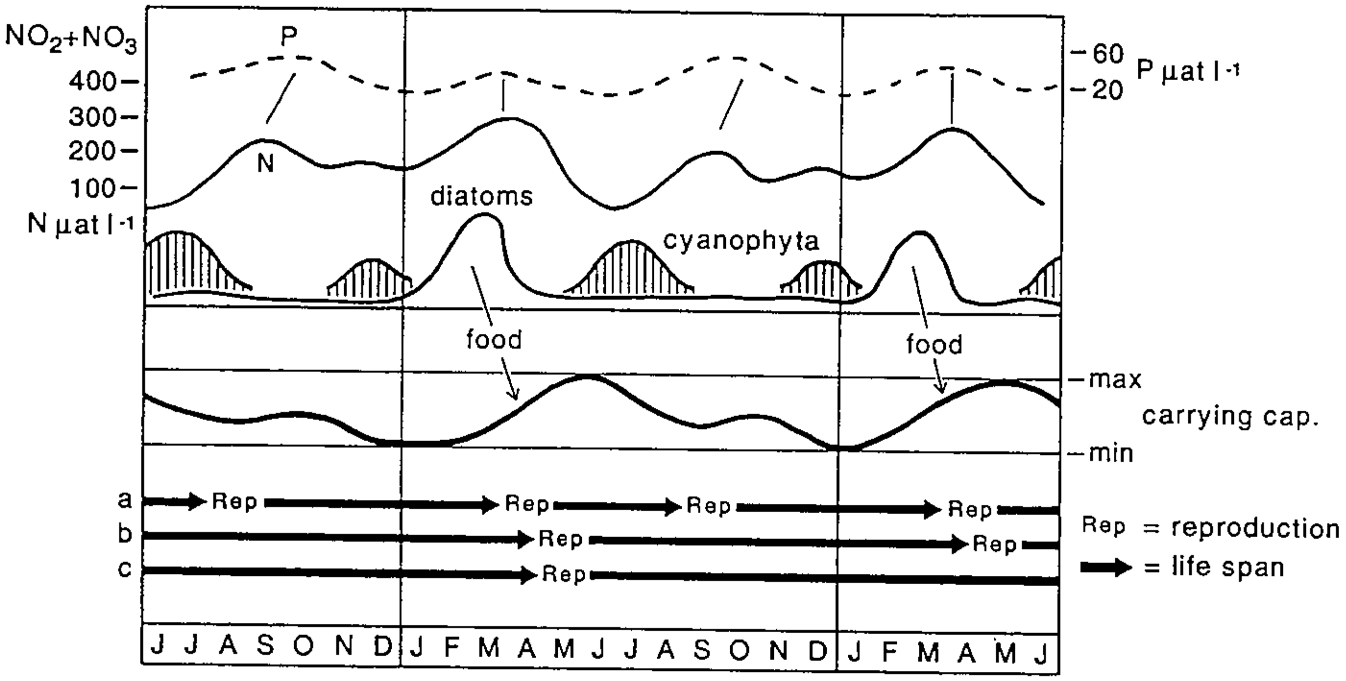

Fig. 2. Environmental conditions corresponding to larger foraminifera which adapt to K-strategies of life by shell size increase and adult dimorphism in order to exploit the seasonal change of nutrient input limiting the carrying capacity. NO_2-NO_3 and P concentrations after measures carried out in the Gulf of Aqaba, summarized by Reiss and Hottinger (1984). Note asymmetry of spring and autumn input. The cyanobacterial populations grow during the "dead" seasons because they have direct access to atmospheric N_2 dissolved in the seawater. They profit in the same time from maximum light penetration during the periods of low phytoplankton density in the surface water. Diatoms are food for foraminifera and regulate their growth. By a combination of reproductive periods, in the yearly cycle, of generations with different life-times, a species can develop a two-fold strategy of life: the long-lived agamontic (B-)generation exploits the low minimum carrying capacity throughout the yearly cycle (b) or during several years (c). The shorter-lived gamonts and/or schizonts, produced by asexual reproduction, exploit by their accelerated initial growth the maxima of nutrient input, forming spring and autumn blooms (a). Suppression of these particular, oligotrophic, environmental conditions is immediately fatal for populations of dimorphic larger foraminifera, as observed under conditions of anthropogenous eutrophication in extant tropical seas.

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SEDIMENTATION HISTORY OF THE UPPER PALAEOCENE AND THE LOWER EOCENE IN THE WESTERN BELGIAN BASIN (SOUTHERN BIGHT NORTH SEA)

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Introduction

In Palaeogene times, the Southern Bight of the North Sea functioned as an intracratonic, shallow-marine, siliciclastic basin with a ramp-type margin shelf setting. This so-called 'Belgian Basin' developed on top of the relatively stable Lower Palaeozoic London-Brabant Massif that sheltered the area from strong subsidence throughout the Tertiary. It accumulated a few hundreds of meters of gently dipping sediment packages that in general consist of alternating sand and clay layers. The shallow shelf environment persisted throughout the Palaeogene and the area was periodically flooded during periods of high relative sealevel. During Thanetian and Ypresian times the shallow North Sea extended westwards into the English Channel, where the rising Weald-Artois High started to form a barrier closing the connection from Lutetian times onwards (or possibly even earlier).

In the Belgian Basin, the Palaeogene strata onlap the Late Cretaceous chalk and dip less than 0.5 % to the NNE. Offshore, they crop out locally on the seabed between the discontinuous sediment cover of the Quaternary 'Flemish Banks'. Onshore in northern Belgium, numerous well-known outcrops in the strike-parallel WSW-ESE oriented strata exemplify the classic Lower Cenozoic geology of the Belgian Basin.

Lithostratigraphy onshore

Dumont (1849) defined the Ypresian as a twofold system, consisting of a 110 m thick transgressive stiff clay (Yc on the old geological map), covered by a 40 m regressive, fine sand (Yd on the geological map). Unfortunately, Dumont only indicated a type-area, without designating a type-section. In view of a revision of the stratigraphical scale, the Ypresian in the Belgian Basin has been revised, both from a lithostratigraphical as from a biostratigraphical point of view. The most important findings (King, 1988; Steurbaut, 1987, 1988; Willems & Moorkens, 1988) have been collected in the special 'Ypresian' volume of the Belgian Society of Geology (1988). Vandenberghe et al. (1988) will geophysically characterise the Ypresian on basis of natural gamma ray logs.

Maréchal & Laga (1988) proposed a new lithostratigraphical subdivision of the Palaeogene, that served for the revision of the geological map, commissioned by the Flemish Government in co-operation with the Belgian Geological Survey. For the time being, the essentially marine, more than 100 m thick, clayey Kortrijk Formation (after the name of the town Kortrijk in the Province of West-Flanders), is subdivided into 4 Members (from top to bottom) :

- Aalbeke Member: a homogeneous, very fine silty clay, practically without any sand fraction, with an average thickness of about 10 m (maximum recorded thickness 20 m); probably corresponding with the 'Argile de Roncq' of French authors; this clay has been deposited during a rather restricted and discontinuous transgressive phase; its top is buried, phosphatic and it has therefore been interpreted as a maximum flooding surface ;
- Moen Member: a rather heterogeneous silt with *Nummulites planulatus*, in the southern part of the basin (area around Kortrijk) consisting of a clayey, coarse to medium silt with thin silt lenses containing up to 10 % fine to very fine sand; to the north the heterogeneity decreases towards a clayey coarse silt; locally, homogeneous clay layers of a few m thickness can occur; it reaches an average thickness of about 40 to 50 m, increasing towards the north to more than 90 m (Kallo well); sedimentation characteristics indicate slightly regressive conditions; corresponds probably with the 'Argile de Roubaix' of French authors ;
- Saint-Maur Member: a homogeneous, very fine silty clay with few thin intercalations of coarse silty clay or clayey, very fine silt; in the southern part of the basin, its thickness reaches up to 15 m, increasing towards the north up to 24 m; it is exceptionally well developed in the Knokke well (more than 100 m) as a result of a major transgressive phase; probably corresponding with the 'Argile d'Orchies' of French authors ;

- Mont-Héribu Member: a shallow marine deposit consisting of an alternation of horizontally laminated glauconitic clayey sands or sandy clays with compact, stiff silty clays or clayey silts; locally bioturbated; its base is characterised by oxidised and compacted clayey sand with lenses pure sand; its thickness increases towards the north, from a few m in the south and the east towards 10 to 15 m in the central part of the basin.

The Kortrijk Formation (basal part of the Ieper Group, Eocene) covers the Landen Group of Late Palaeocene age.

Well log facies of onshore deposits

The revision of the geological map, commissioned by the Flemish Government (Administration of Natural Resources and Energy) and the Belgian Geological Survey, suggested the subdivision of the clayey Kortrijk Formation on the basis of a qualitative interpretation of (long normal and short normal) resistivity and natural gamma well logs, into 7 distinct sediment packages that seemed correlatable over long distances (De Ceukelaire & Jacobs, in press). Their lower boundaries have been colour-coded.

The 'brown' boundary characterises a sharp resistivity drop and a sharp gamma increase, delineating a homogeneous clay layer of restricted thickness (3 to 7 m). The 'grey' boundary marks an increase only traceable in resistivity logs, thus indicating a 7 to 10 m thick sediment package with a more sandy but glauconitic character. Resistivities decrease and show a more homogeneous character above the 'red' boundary, suggesting the presence of a stiff and compact clay layer of about 12 m thickness. The 'green' boundary marks the first of three consecutive resistivity increases, indicating an overall coarsening upwards grain size trend. Although of minor importance, this first increase can be traced over several wells, delineating a 16 m thick sediment package. The second resistivity increase above the 'orange' boundary is much more pronounced and shows a more heterogeneous character, mostly underlined by 2 distinct but broad peaks that fade to the west. It is almost accompanied by a gamma decrease. This package is about 15 m thick. The third resistivity increase above the 'blue' boundary characterises a heterogeneous sediment package of 25 to 30 m thickness. It consists in itself of 4 peaks, the lowermost of which is a broad but easy recognisable one, traceable over long distances. As these 4 maxima decrease from bottom to top, they indicate a general but slight increase in clay or (more probably) glauconite content as reflected by the slight gamma increase. The last 'yellow' boundary separates this heterogeneous (more sandy) sediment package with high resistivities from the top package with low resistivities, indicating the presence of a clay layer of about 10 m thickness.

The 'brown' boundary evidences the superposition of the Kortrijk Formation on the Landen Group of Thanetian age. It represents the basal boundary of a thick clay layer on a marked sand package. Their strong lithological difference is reflected by a sharp resistivity drop. In the Ieper area, the 'red' boundary has tentatively been used to separate the Mont-Héribu Member from the overlying Saint-Maur Member, although this interpretation is not conform with the initial definition of the Mont-Héribu Member as a thin basal sandy deposit (De Coninck et al., 1983). For the time being, the 'brown-red' sediment package is therefore considered as a clayey equivalent of the Mont-Héribu Member. For the geological mapping, the 'orange' boundary has been considered as the boundary between the Saint-Maur Member and the overlying Moen Member, arbitrarily chosen at the midst of the 3 resistivity increases easily recognisable by grain size increases in samples of carefully executed well drillings. Detailed correct correlation though necessitates the presence of all 3 resistivity increases in all wells. Finally, the 'yellow' boundary marks the base of the Aalbeke Member.

Detailed analysis of the geophysical well log character of the Kortrijk Formation sediments allows further subdivision of these 7 packages in to a series of 23 that can be correlated throughout the southern part of the Belgian Basin.

Seismic stratigraphy offshore

A fine-scale seismic stratigraphic model for the Palaeogene offshore on the Belgian continental platform was formulated on basis of a dense, high-resolution reflection seismic grid (De Batist 1989; De Batist & Henriet 1995; Jacobs & De Batist 1996). Geometry and seismic facies characteristics defined 13 major seismic-stratigraphical units. Thanetian unit T2 could be correlated with the continental to lagoonal Tienen Formation (Knokke Member) of NW Belgium and the marine to lagoonal Woolwich/Reading Beds plus the Oldhaven Member of SE England. Distinct channel cut-and-fill structures can be observed on the seismic sections at the base and at some metres above the base of unit T2, defining 2 erosional surfaces. The incisions are most likely related to Thanetian eustatic sealevel falls, but a tectonic influence related to a first phase of updoming of the Weald-Artois High, cannot be ruled out. The basal channels, up to 20 m deep, seem to be part of a fluvial drainage system that is oriented towards the east and north-east. Unit T2 probably contains 2 sequences, each represented by backfilling channel deposits (late lowstand or transgressive systems tract) and overlying parallel reflections (highstand systems tract). The seismic facies could be interpreted as resulting from tidally influenced sedimentation.

Five seismic units, Y1 to Y5, have been designated an Ypresian age. Unit Y1 reaches a thickness of more than 150 m, and its subcrop covers a large part of the Belgian continental platform. It correlates with the 'leper clay' of Belgium, which includes the Kortrijk Formation and the Kortemark Member of the Tielt Formation, and its equivalent in southern England, the London Clay Formation. A seismic marker about 4 m above the conformable base of unit Y1 can be correlated with the Harwich Stone Band in the Thames Estuary, a volcanic ash layer deposited during the intense volcanism associated with the onset of seafloor spreading north of Scotland. Unit Y1 is characterised by a homogeneous seismic facies, suggesting a relatively long period of stable, low-energy depositional environments. Onshore, the 'leper clay', interpreted to be deposited in a mudshelf environment (Jacobs & Sevens 1993), is subdivided into 5 members (Maréchal & Laga 1988) and into 5 depositional sequences (Vandenberghe et al., in press), that cannot be identified on the seismic sections because the internal reflection pattern is strongly disturbed by a wide range of intraformational 'sediment tectonic' deformations.

Lithostratigraphy offshore

This seismic stratigraphy has been integrated with the results of 4 boreholes near the Belgian coast, that contained a composite, nearly continuous, 200 m thick sediment succession of Eocene age (Jacobs & De Batist 1996). A major part of this Eocene sediment series is formed by the Ypresian clays, that were studied on macroscopical and submicroscopical scale. Sediment facies analysis of the cored wells suggest that the upper part of the Ypresian clays was deposited partly on a muddy shelf and partly in a pro-delta environment. Submicroscopical analysis of the clay core samples documented clay-particle orientations that indicate transgressive/regressive sedimentation trends (Van Bavinchove 1993; Jacobs et al., in press). In combination with the occurrence of thin siderite laminae and of organic-rich black clay layers, these sediment characteristics can be used to decode sequence stratigraphic signals and unravel burial history.

Lower Eocene sedimentation history

Combination and integration of large-scale (such as seismic and facies analysis) with small-scale techniques (such as submicroscopical and even micropalaeontological analysis) enabled reconstruction of the evolution of the depositional environment during Early Ypresian times (Jacobs et al., in press). In a first phase, deposition takes place below storm wave base in palaeowater-depths of 20 to 40 m on a constantly but slowly subsiding mudshelf with a low-gradient relief. Sedimentation with a southern sediment supply and low sedimentation rates is characterised by fine clay decantation on the central to inner shelf. Relative sealevel rose constantly but stepwise, resulting in the deposition of a stacked pattern of 4 dip-stretched TST-parasequences of decametric thickness, that can be correlated over long distances as has been testified by detailed grain-size analysis of onshore cored wells (Geets 1988). A dark, organic-rich horizon with preferential clay-particle orientations indicate anoxic bottomwater conditions. This anoxic event represents a condensed section and might correlate with one of the thin glauconitic horizons or concentrations, known from onshore outcrops and boreholes. It might stand as a candidate for the Lower Ypresian maximum flooding surface as younger organic-rich horizons show sediment reworking by submarine erosion.

In a second phase, the mudshelf gradually comes under storm influence due to rapid relative sealevel fall at the end of Ypresian clay deposition times. In more proximal positions characterised by shallower water depths, bacterial sulphate reduction produces early diagenetic framboidal and botryoidal pyrite and siderite during early burial phases at shallow sediment depths before any compaction occurs. Storm activities erode these sideritic sediments that are transported basinwards over short distances, concentrating the siderite into distinct storm deposits. Reinforced or resumed bottomwater circulation (probably induced by the ongoing uplift of the Weald-Artois High) breaks up the anoxic conditions, depositing regressive HST-parasequences of reduced thickness. Meanwhile, Ypresian bottom clay layers are sealed through compaction and consequent water discharge into the underlying Landenian (or Thanetian) sands.

In a third and final phase, during and after Ypresian sands deposition, Ypresian top clay layers are compacted and sealed, building up a Rayleigh-Taylor density instability in the water-logged middle portion of the clay (Henriet et al., 1988). Consequent water escape and breakthrough generate intraformational, postdepositional, (tectonic or slope instability induced?) sediment deformations, visible onshore as e.g. listric fault patterns.

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MAGNETOBIOSTRATIGRAPHIC CORRELATIONS IN THE UPPER PALEOCENE-LOWER EOCENE ALAMEDILLA SECTION, SPAIN

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Based on published biostratigraphic and chemostratigraphic data (Lu *et al.*, 1996), the Alamedilla section, located in the Betic Cordillera, appears as a valuable section for determining the position of the late Paleocene carbon isotope excursion in Chron C24r. The $d^{13}C$ excursion is reported to be lie in mid Zone P5, a position that is predicted from a composite stratigraphic section representing Chron C24r (Berggren and Aubry, 1996). The samples upon which the calcareous nannofossil stratigraphy is established are those collected by E. Molina and reported on by Lu *et al.* (1996) and Arenillas and Molina (1996). They were taken at 20 to 50 cm intervals except in the vicinity of the excursion where they were taken at 10 cm intervals. We sampled the approximately 30 m-thick Alamedilla section for magnetic polarity stratigraphy in an attempt to construct an integrated magnetobiostratigraphic framework. The oriented block samples used for magnetostratigraphic investigation were taken at approximately 1 m interval depending on suitability of lithology. The sampling was done with the help of E. Molina and is thus precisely tied to his own sampling. Our results thus can be directly integrated with published data. Smear slides were prepared for calcareous nannofossil analysis.

For the paleomagnetic analysis, subsamples were cut from the oriented block samples and natural remanant magnetization (NRM) was measured with a cryogenic magnetometer. Stability of NRM was investigated by progressive thermal demagnetization in 10 to 15 steps typically to 580° C (for dominant magnetite carrier) but extended to 680° C in a few samples that evidently contained hematite as a carrier of the natural remanent magnetization. Unfortunately, only 6 of the 30 samples showed evidence of a characteristic component, identified on the basis of a linear demagnetization trajectory. The magnetization directions appear to be tectonically rotated in the steep to overturned beds but samples at 7.0 m, 7.8 m, 14.0 m, and 29.0 m can be interpreted to record reverse polarity, whereas samples at 0.8 m and 1.8 m apparently record normal polarity; the polarity in the remainder of the samples cannot be determined with any confidence due mainly to unstable magnetic behaviour. There are too few samples with reliable data from the Alamedilla section to construct a meaningful magnetostratigraphic polarity pattern for correlation to the geomagnetic polarity time scale (GPTS).

The Paleogene part of the Alamedilla section extends from Zone NP8 to Zone NP11 (?NP12), but because preservation is extremely poor at most levels, with both dissolution and overgrowth, (sub)zonal boundaries are difficult to position. Likewise, fine delineation of nannofossil datums is not possible. A large contrast occurs in the qualitative and quantitative compositions of assemblages between levels 13.30-13.35 m and 13.71 m (samples taken at levels 13.50-13.55 and 13.60 m were barren). The lowest occurrence (LO) of species of *Rhomboaster* is located at 13.71 m. In addition, we note the LO of *Discoaster araneus*, an unusual asymmetrical discoaster which we regard as equivalent to the excursion foraminiferal species (El Naggar, 1966; Kelly *et al.*, 1996), and a drop in abundance of fasciculiths. According to Lu *et al.* (1996) the benthic foraminiferal extinction occurs at 13.45 m and the carbon isotope excursion is spread over a 50 cm interval, with minimal values beginning at 13.71 m. The coincidence of the LO of *Rhomboaster* spp., *D. araneus* and the drop in abundance of *Fasciculithus* spp. with the excursion has been observed elsewhere, and in particular in the Bass River section, New Jersey. However, because of the occurrence of a 20 cm-thick barren interval between 13.70 m and 13.50 m these calcareous nannofossil events and the highest occurrence of the *Stensioina beccariiformis* fauna at 13.45 m (no foraminifera are preserved in the barren interval, Arenillas and Molina, 1996) should not be regarded as real events, but pseudoevents. The Alamedilla section thus does not contain a satisfactory record of the succession of paleontologic events in the vicinity of the carbon isotope excursion.

Poor calcareous nannofossil biostratigraphy combined with a weak magnetic record compromise the utility of the Alamedilla section for studying late Paleocene-early Eocene long-term changes. Only poor chronologic resolution can be achieved for this section, and it cannot be precisely determined how much time is represented by the carbonate-barren interval immediately above the HO of the *S. beccariiformis* fauna and immediately below the LO of *Rhomboaster* species and other nannofossil taxa. Only a few levels yield a reliable polarity. Based on the delineation of the NP8/NP9 zonal boundary between levels 1.80 and 1.85, we are confident that the interval between 0.8 m and 1.8 m represents part of Chron C25n. Clear reversed polarity at 7 m, 7.8 m, and 14 m are indicative of Chron C24r, correlative with Subzones NP9a and b.

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TEMPORAL SEPARATION OF NORTH SEA BASIN RESTRICTION AND THE LATEST PALAEOCENE WARM WATER PULSE: THE HIGH RESOLUTION PALYNOFACIES VIEW.

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UK Well 22/10a-4 is situated in the Everest Field, on the flanks of the Jaeren High in the UK North Sea. This well yielded over 122m of continuous, well preserved core, representing an expanded section through the late Palaeocene Maureen, Lista and Sele Formations. The intervals of interest in Well 22/10a-4 are the Lista/Sele boundary, marking the onset of North Sea basin restriction, and the Sele Formation, which was found to contain the well documented late Palaeocene warming event.

Palynofacies analyses carried out on samples from Well 22/10a-4 indicate the onset of the tectonic restriction of the North Sea Basin extremely well. Dramatic changes occur in the palynofacies assemblage at the level of the Lista/Sele boundary, the green to grey-green, bioturbated claystones of the Lista Formation containing a palynofacies dominated by black (oxidised) wood, and very few other components, this formation representing the open marine conditions present prior to the tectonic restriction of the basin. The Sele Formation is characterised by dark grey to black, well laminated mudstones, and has a much more diverse palynofacies assemblage, dominated by brown (unoxidised) wood, and moderate abundances of sporomorphs. Many of the component groups show significant changes in abundance at the Lista/Sele boundary.

The tectonic event causing the restriction of the North Sea Basin is clearly separated from the late Palaeocene climatic warming event, which in Well 22/10a-4 is indicated by an influx of the warm water wetzelielloid dinocyst *Apectodinium*. At this locality, this dinocyst is present in the Sele Formation from a level 15m above the Lista/Sele boundary. *Apectodinium augustum* is present in the highest samples taken from well 22/10a-4 indicating that the core ends just below the level currently taken as the Palaeocene/Eocene boundary in the North Sea, which is taken at the last appearance of *Apectodinium augustum*.

The preliminary data collected illustrate that the onset of North Sea basin restriction and late Palaeocene climatic warming can be clearly temporally differentiated at this locality.

LATE PALEOCENE MICROBIOTIC CHANGES IN THE NORTHEASTHERN PERI-TETHYS : COMPARATIVE STUDY OF THE SOKOLOVSKY QUARRY (TURGAY DEPRESSION) AND SENGILEY (MIDDLE REACHES OF THE VOLGA) SECTIONS

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The well-understood sections in the middle reaches of the Volga R. (Sengiley) and in the northern Turgay depression (Sokolovsky quarry) display two transgressive–regressive sedimentary cycles spanning the upper Thanetian–lowermost Ypresian (?) (figure). The lower cycle, roughly corresponding to the nannoplankton Zone CP7, features a major role of organic deposits—carbonates (bioherm limestones) in the Sokolovsky quarry or silicious sediment (diatomites) in the Sengiley section. The synchroneity of these deposits is demonstrated by the radiolarian and foram coappearances in the transitional sections of the northern Peri-Caspian area. In both sections, the upper cycle comprises terrigenous–siliceous deposits and spans the terminal Paleocene–base of the Ypresian (?).

Sokolovsky quarry. The lower cycle includes the Sokolovsky Sequence, composed of calcareous sandstones with bioherms and superposed by carbonate-free sands. The calcareous sandstones yielded a nannoplankton assemblage of the CP7 **Discoaster mohleri** zone. The middle part of this sequence bears an intercalation of bioherm limestone with *Briozoa, Cirripedia, Rugosae, Ostracodes, Bivalvia*, and planktonic forams of the **Acarinina subsphaerica** zone (Shutskaya, 1970). Based on dynocysts, the carbonate portion of the section is assigned to the **Cerodinium speciosum** zone.

The upper cycle of the section—the "Banded Sequence" of (Amon, 1992)—begins and terminates with members of alternating gaizes and loose sandstones. The middle part of the section is largely siliceous and consists of diatomaceous clays.

The "Banded Sequence" bears sporadic nannoplankton, equally assignable to the upper Thanetian and lower Ypresian (CP8 **Discoaster multiradiatus**–CP9 **Discoaster diastypus** zones). The diatoms and silicoflagellates abound in the lower member of alternating gaizes and siliceous sandstones and in the middle, clay member. In the lower member, the diatom–silicoflagellate assemblage, albeit not diverse, contains index species of the **Coscinodiscus uralensis** zone of the diatom provincial scheme (Strelnikova, 1992) and of the **Naviculopsis constricta** zone of the standard scale on silicoflagellates. The middle part of the sequence hosts a diverse diatom–silicoflagellate assemblage of the same zonal units. The diatomaceous clays of the middle part of the sequence yielded radiolarians of the **Petalospyris foveolata** provincial zone (Kozlova, 1984). Further upsection, all three groups of siliceous nannoplankton acquire elements characteristic of Eocene assemblages. On dynocysts, the "Banded Sequence" corresponds to the **Apectodinium homomorphum** zone. Beside the index species, *A. quinquelatum* appears here. Both species are characteristic of the upper Thanetian.

Sengiley. The base of the lower cycle hosts a thin member of alternating sandstone and gaize, and the overlying part of the cycle is composed by a 20-m diatomite member. The upper cycle consists of alternating gaize and siliceous clay. On radiolarians, the deposits of the lower cycle correlate with the **Buryella tetradica** and **Tripodiscinus sengilensis** zones (Kozlova, 1984), and the deposits of the upper cycle correspond to the lower part of the **Petalospyris foveolata** zone. The deposits of the lower cycle contain assemblages of the **Trinacria ventriculosa** and **Hemiaulus proteus** diatom zones, and in the deposits of the upper cycle species appear that became widespread in the Eocene—*Coscinodiscus uralensis (= Coscinodiscus josephinus).* On silicoflagellates, both cycles

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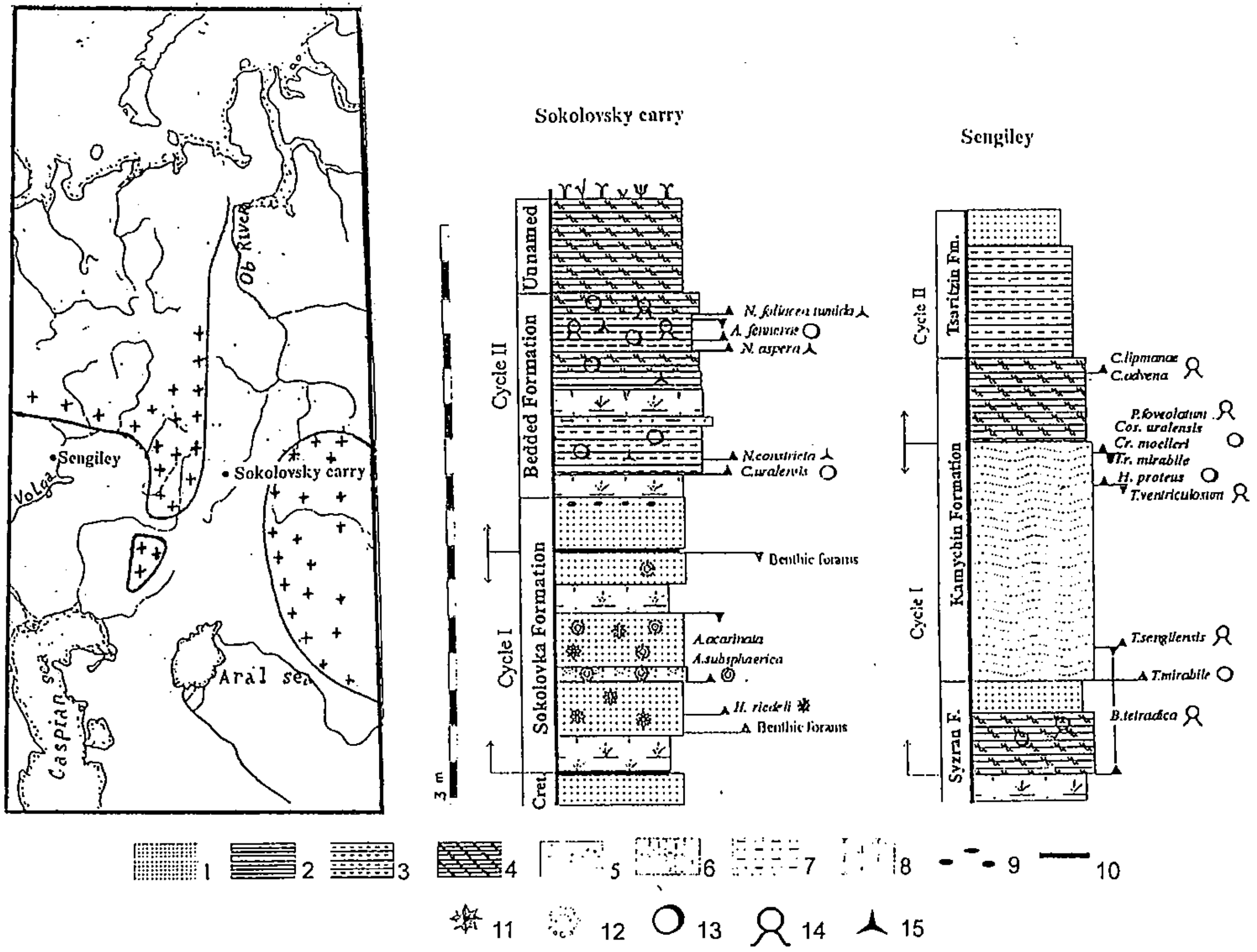
Ob River
Sengiley
Sokolovsky carry
Volga
Caspian sea
Aral sea
Sokolovsky carry
3 m
Cycle II
Cycle I
Unnamed
Bedded Formation
Sokolovka Formation
Cret.
N. foliacea tumida
A. feminae
N. aspera
N.constricta
C.uralensis
Benthic forams
A.ocarinata
A.subsphaerica
H. riedeli
Benthic forams
Sengiley
Cycle II
Cycle I
Tsaritzin Fm.
Kamychin Formation
Syzran F.
C.lipmanae
C.advena
P.foveolatum
Cos. uralensis
Cr. moelleri
Tr. mirabile
H. proteus
T.ventriculosum
T.sengilensis
T.mirabile
B.tetradica
1
2
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correlate with the **Naviculopsis constricta** zone. This sequence is overlapped with clear erosion by sediments containing continental flora assemblages.

The studied sections contain abundant siliceous and organic-walled plankton illustrating the character of changes in radiolarian, diatom, silicoflagellate, and dynocyst assemblages at the Paleocene/Eocene transition.

The **B. tetradica** zone is dominated by well-known tropical oceanic radiolarian species suggestive of a short-term invasion of a Tethyan assemblage into a boreal basin. A dramatic change in the species composition occurs at the boundary of the **B. tetradica** and **T. sengilensis** zones. Tethyan species disappear, and the assemblage becomes drastically reduced to as many as 10–12 species. The newly appearing species are chiefly boreal and are also characteristic of coeval West Siberian strata. The **P. foveolata** zonal assemblage becomes increasingly abundant in new species. Distinguished among them are boreal, high-latitude and Tethyan bipolar as well as widespread tropical species, the last being dominant. The Sokolovsky quarry displays a younger assemblage of the **P. foveolata** zone. Species characteristic of the Eocene make their first appearance here: *Eusyringium striata striata* (Brandt), *Lychnocanium tripodium* Ehr., *Triceraspyris palmipodiscus* Petrushevskaya, *Tripodiscinus tumulosus* (Kozlova), and, possibly, *Petalospyris fiscella* (Kozlova).

The diatom assemblage assigned to the **T. ventriculosa** zone consists of neritic species. Further upsection, starting from the base of the **T. proteus** zone, oceanic species of the *Hemiaulus, Ceratium,* and *Soleum* genera appear in the assemblage. *H. proteus, H.incurvus, H.incisus,* and *Pyxidicula moelleri* make their first appearance. From the beginning of the upper cycle, the assemblage becomes essentially more diverse. Along with the oceanic species *Triceratium heibergii* and *Craspedodiscus moelleri,* the assemblage yields new elements of neritic flora: *C. uralensis* Jousé (= *Aulacodidcus suspectus* Schmidt), presently classed as the new genus *Veniaminia* (Strelnikova, 1997). In the terminal Paleocene, several monogenus taxa successively replace one another: at the bottom, *Cylindrospira simsi*, and at the top, *Annaulus fennerae*. .

The silicoflagellate assemblage, beginning from the base of the lower cycle, already shows all of the species characteristic of the Late Paleocene–Early Eocene: *N. constricta, N.danica, N.robusta, Dictyocha elongata, Corbisema disimmetrica.* Within the upper cycle, in the **C. uralensis** zone, *N. aff. punctilia–N. aspera*–replace one another in succession. *N. foliacea tumida* and *N. foliacea* appear at the very top of the section.

The organic-walled plankton assemblage characteristic of the carbonate sequence in the Sokolovsky section is dominated by chorate cysts evidencing the existence of a normal marine basin.

In summary, the assemblages of all of the siliceous plankton groups show rapid changes owing to alternating open-sea and semiclosed-shelf environments and to the appearance of short-lived taxa. The terminal Paleocene features a nearly total absence of carbonate plankters; species characteristic of the Eocene, including those that became widespread in the Middle and Late Eocene, appear in all of the microfossil groups.

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Legend: 1-Sand; 2- Clay: 3 -Sandy gaize; 4-Siliceous siltstone; 5- Diatomite; 6-Bioherm limestone; 7- Siltstone; 8- No exposure; 9- Phosphatic pebbles; 10- Hiatus; 11- Nannofossils; 12- Foraminifers; 13- Diatoms; 14- Radiolaria; 15- Silicoflagellates. Crosses on the map show the Paleocene land .

THE PALAEOCENE/EOCENE TRANSITION IN THE NORTH SEA AREA, REFLECTED IN FORAMINIFERAL FAUNAS

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Correlating the foraminiferal events of the North Sea basin with the standard planktonic zonation is extremely difficult as the sediments are virtually non-calcareous, but through correlation with dinoflagellates a chronostratigraphic framework may be established.

The benthic extinction event at the Palaeocene/Eocene boundary is believed to be reflected in the agglutinated foraminiferal faunas of the North Sea region. This is illustrated by an example from onshore Denmark in the Bovlstrup well (Laursen & Andersen, 1997) (Fig. 1).

Three zones comprising exclusively agglutinated faunas have been established in the Bovlstrup well (zones 2-4), including the first record of foraminifera from the Ølst Formation close to the Palaeocene-Eocene boundary (Fig. 1).

Relatively high concentrations of agglutinated foraminifera are found in the uppermost part of the Palaeocene. *Spiroplectammina spectabilis* dominates the fauna (Zone 2) of the Holmehus Formation (Lista Formation equivalent). Then the diversity of the fauna decreases and two species characterise the fauna of Zone 3: *Verneuilinoides subeocaenus*, which is restricted to this zone, and *Haplophragmoides* sp. 1. Zone 3 is a transition zone from the rich Zone 2 to the poor Zone 4 (See Fig. 1). In Bovlstrup Zone 4, the Ølst Formation (Sele and Balder Formation equivalent) has a fauna characterised by a very low number of species in rather fluctuating numbers. The foraminiferal fauna of Zone 4 consist mainly of *Evolutinella* sp. 2 and *Verneuilinoides* sp. 1 accompanied by an influx of the diatoms *Coscinodiscus* sp. 1 and *Coscinodiscus* sp. 2.

Through correlation with dinoflagellates a rough chronostratigraphy of the Bovlstrup well is established. A sample 3 metres (at 209m below surface) into the lowermost agglutinated zone (Zone 2) contained a Viborg Zone 3 dinoflagellate flora, and two samples from the uppermost part of Zone 2 posessed Viborg Zone 4 floras (Heilmann-Clausen, 1985; pers. comm., 1988). This implies that Zone 2 corresponds to both a part of dinoflagellate Zone 3 and probably most of Zone 4 of the Viborg section (Heilmann-Clausen, 1985). Therefore, Zone 2 of Bovlstrup can probably be referred to nannoplankton Zones NP 6 - NP 8 of Martini (1971), indicating a possible Late Selandian to Thanetian age. This is also in agreement with the correlation of Zone 2 with the *Trochammina ruthvenmurrayi-Reticulophragmium paupera* Zone of Gradstein *et al.* (1994) and with Zone NSA 1b (King, 1989).

The top of Bovlstrup Zone 3 probably corresponds to bioevent M5 of Mudge & Copestake (1992) and O'Conner & Walker (1993). These authors state that the event is isochronous over large parts of the North Sea basin as it probably reflects a change in the bottom water conditions. Unfortunately no dinoflagellate analyses were carried out in Bovlstrup Zone 3 and the chronostratigraphic allocation of the zone is thus uncertain. Bovlstrup Zone 3 falls between samples dated by dinoflagellates to be Early Thanetian (Viborg Zone 4, Heilmann-Clausen, pers. comm.) and samples dated to Late Thanetian to Early Ypresian. Bovlstrup Zone 3 is therefore suggested to be of Thanetian age.

Dinoflagellates were studied in three samples from Bovlstrup Zone 4 (156, 154, and 152 m's below surface). The flora placed the samples in Zone 7 of Viborg and thus with the upper part of the Ølst Formation (Heilmann-Clausen, 1985, p. 30) indicating a Late Thanetian to Early Ypresian age. The dinoflagellates suggest a tentative correlation of Bovlstrup Zone 4 with nannoplankton Zones NP 9-10 of Martini (1971). The formal boundary between the Palaeocene and the Eocene is not yet internationally agreed upon, but it is normally placed at, or close to the NP 9-10 boundary, and therefore the Bovlstrup well gives a detailed picture of the foraminiferal events at the Palaeocene/Eocene transition.

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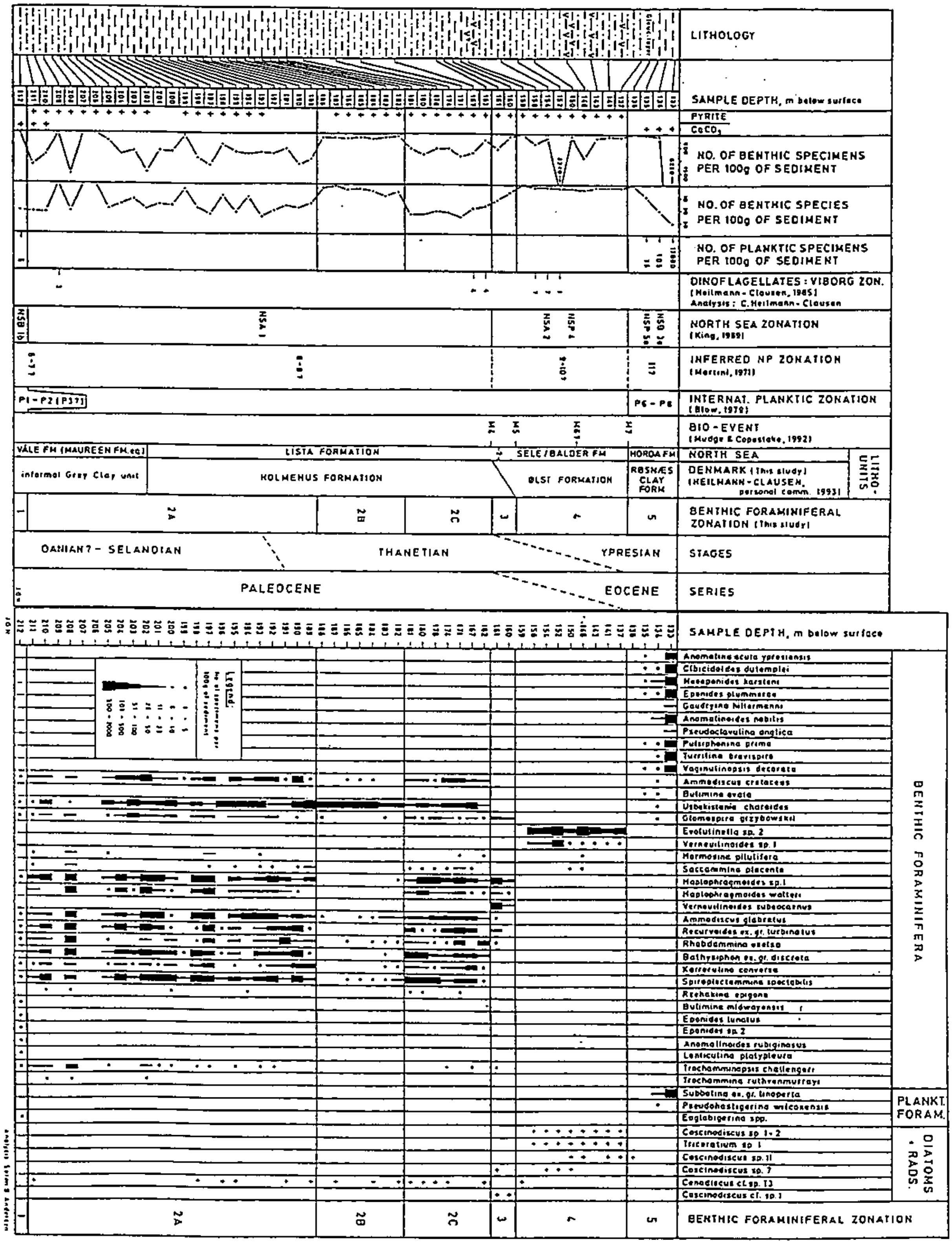

Fig. 1. Range chart of selected species from the Bovlstrup well combined with correlation scheme of biostratigraphy and lithostratigraphy.

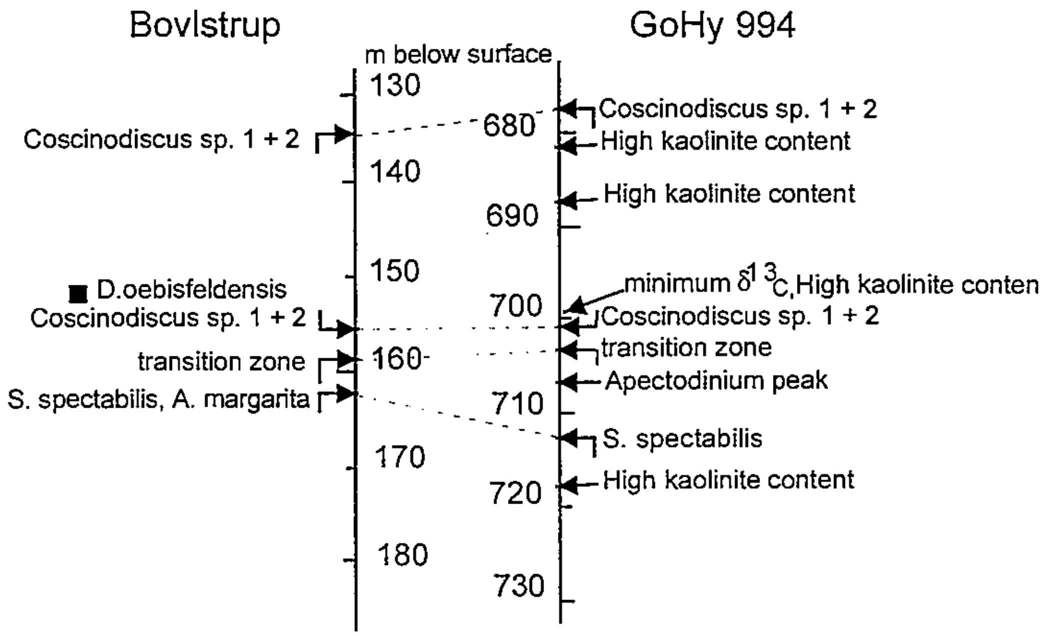

Fig. 2. Correlation of the Bovlstrup well in Denmark (Laursen & Andersen, 1997) with the GoHy 994 borehole in Germany (Dill et al., 1996)

There are very few published papers examining in detail single sections. One of the few exceptions is the GoHy 994 borehole (Dill et al., 1996) which presents a thorough palaeontological and sedimentological analysis of a borehole through the estaurine deposits in Northern Germany.

A number of events have been suggested as Palaeocene/Eocene boundary events (e.g. Knox, 1996): 1) a pronounced short term negative shift in the carbon isotope values, 2) a benthic extinction event, 3) a kaolinitic influx. The following events are registered in the German study: 1) a distinct negative excursion is seen in the minimum d^{13}C just when the faunas change from being dominated by agglutinated foraminifera to the influx of pyritized *Coscinodiscus,* 2) clear reduction in the number of species, 3) through Formation A and B kaolinite accumulated to be replaced by chlorite higher in the section.

In the established zonation schemes for the North Sea area the faunal turnover at the Palaeocene/Eocene transition is also registered by King (1989) and Gradstein et al. (1994), even though they are presented in a less detailed manner than in the Bovlstrup and Gohy 994 sections. The benthic extinction event is also demonstrated in the taxonomic work of Charnock & Jones (1990). The agglutinated faunas of the Late Palaeocene are fairly rich in advanced species. After this point the faunal turnover occurs and the only surviving species, or species restricted to this limited time interval are mostly the epifaunal opportunists *Thurammina papillata* and *Evolutinella* sp. 2. Only one form is interpreted as infaunal, namely *Verneulinoides subeocaenus* which is observed in this time interval. This form also occurs in less stressed environments higher in the section.

Combining the information of the above mentioned references the following list of events occur in ascending order over the Palaeocene/Eocene transition (Fig. 2): The kaolinitic increase beginning in the *Spiroplectammina spectabilis* zone, the extinction of *Spiroplectammina spectailis* which coincides with the top of the dinoflagellate zone of *Alisocysta margarita* (data from the Bovlstrup well, Laursen & Andersen, 1997), followed by a transition zone with a decreasing number of foraminiferal species and specimens (in this transitions zone a peak abundance of the dinoflagellate *Apectodinium* is observed in the GoHy 994 borehole, Dill et al., 1996). The next event is the influx of the pyritized diatoms followed by a pronounced negative excursion in the carbon isotope values (data from Dill et al., 1996) and the influx of the dinoflagellate *Deflandrea oebisfeldensis* (data from Laursen & Andersen, 1997).

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LATE PALAEOCENE MAMMALS FROM THE TREMP FORMATION (SOUTHERN PYRENEES, LLEIDA,SPAIN)

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The Tremp Formation in its type area (Tremp basin, Lleida, Spain) consists of up to 900 m thick coastal and continental deposits ranging in age from Late Cretaceous to Late Palaeocene. In its upper part, four localities have yielded four samples of vertebrate microfossils (figure 1). They are situated near the base of the Ilerdian stratotype, which is a stage younger than the Thanetian and overlapping the Ypresian (Hottinger & Schaub, 1960; Molina et al., 1992; Serra-Kiel et al., 1994). This situation allows to correlate these continental fossil sites with the marine bio- and chronostratigraphy and with the global palaeomagnetic scale.

Fossil teeth and bones of mammals together with thick eggshells from giant birds (*Ornitholithus*) and minor rests of other taxa (crocodiles, frogs, *Microcodium*) have been recovered by washing and screening 21 tons of sediments from four localities, namely Claret 0, Claret 4, Tendruy and Palau (López-Martínez *et al.*, 1995, 1996). The fossiliferous beds consist of grey marls intercalated in a thick succession of red clays with conglomerates, sandstones and gypsum levels. The lowermost locality is 80 m below the base of the Ilerdian. The three other localities are situated in the last 30 m of the Tremp Fm. at the base of the sedimentary sequence initiating the Ilerdian transgression (fig. 3).

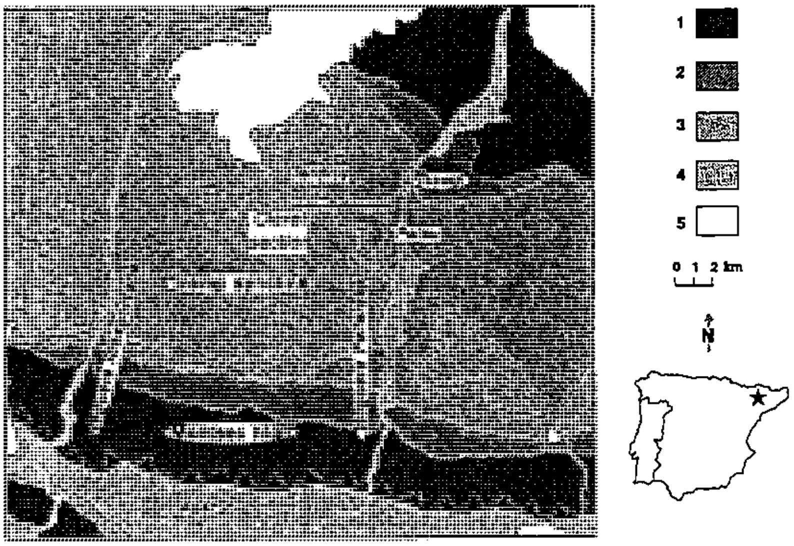

Figure 1.- Geological map of the Tremp region with situation of the Mammal Palaeocene localities. 1- Late Cretaceous and older; 2- Arén Formation; 3- Tremp Fm.; 4- Ilerdian; 5- Late Eocene conglomerates.

About 250 teeth and fragments have been identified, most of them from locality Claret 4. Eight species of Mammals have been recognized so far, belonging to the orders Multituberculata (cf. *Hainina*), Insectivora (?*Afrodon* sp.; *Adapisorex* sp.), Proteutheria gen. nov. and Condylarthra (*Paschatherium cf. dolloi*, *Paschatherium* sp., *Microhyus* cf. *musculus*, *Landenodon* sp.). The first poor indices of the presence of Chiroptera have not yet been confirmed (López-Martínez *et al.*, 1995). Many teeth are badly corroded by abrasive chemicals in contrast with the eggshell fragments, suggesting that digestion processes affected some of the mammalian rests before burial (López-Martínez *et al.*, 1996).

Multituberculates are represented by two teeth closely resembling the genus *Hainina* from Hainin (Belgium, Early Palaeocene, MP 1-5) and Cernay (France, Late Palaeocene, MP 6).

The Insectivora are represented by two taxa, one of them resembling the genus *Afrodon* described in the Palaeocene of Morocco, also present in the European site Walbeck dated as early Late Palaeocene (MP 5 or 6; Gheerbrandt & Russell, 1989; Gheerbrandt et al., 1997).

The condylarthra are by far the most abundant mammal in the assemblage, especially the species *Paschatherium cf. dolloi* which constitutes up to 75% of the sample. The metric and morphological characters of this population closely relate it to the homonymous taxon from Dormaal (Belgium, MP 7), where it is extremely abundant also (Denys and Russel, 1981). In the neighbouring site of Campo, a population described as *Paschatherium yvettae* slightly differs in having a large talonid in m3 (Gheerbrandt *et al.*, 1997). All three populations differ from *Paschatherium marianae*

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ESTRAVIS & RUSSELL 1992 from Silveirinha (Portugal, MP 7) and *P. plaziati* MARANDAT 1989 from Fordones (France, MP 7) by a slightly larger size and more bunodont cusps. Another small-sized *Paschatherium* species with sharpened cusps in its teeth is present in less number in the Tremp localities. The species *Microhyus cf. musculus* represented in Claret 4 and Tendruy by several teeth is also closely related to that of Dormaal, both being slightly larger than *Microhyus reisi* from Silveirinha, (Estravis, 1992). Moreover, a relatively large condylarthre is represented in Palau by a single lower molar with a lophodont pattern similar to that of *Landenodon* from Dormaal. The Spanish form is however slender in shape and smaller in size, probably representing a decidual tooth.

The overall composition of these assemblages is peculiar in having a low number of taxa. This is not only the case of these four localities but also of the Late Palaeocene site Campo, situated westwards in the central Pyrenees. Other Late Palaeocene-Early Eocene fossil sites from Europe, Africa and North America having a similar sample size bear three or four times more species than the Pyrenean localities (figure 2). Thus the poor diversity of the Spanish Palaeocene faunas seems to be a consistent pattern, although more data are needed in order to interpret this palaeoecological and/or chronological signal.

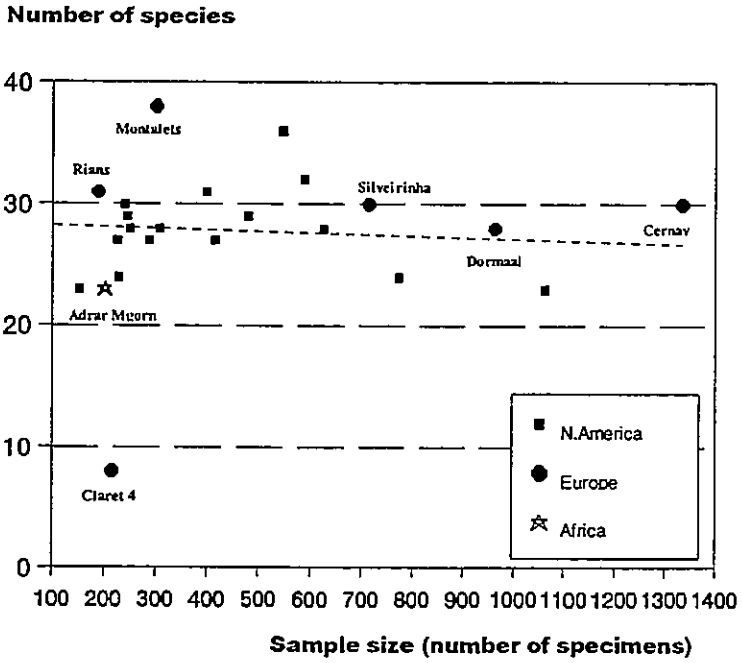

Figure 2.-
Plot between number of taxa against number of specimens of mammals in several localities from Palaeocene to Early Eocene age. There is no correlation between both variables, as indicated by the low value of the regression slope. The low diversity value in Claret 4 (Tremp Fm) is thus the signal of a significant chronological and/or palaeoecological difference.

The chronostratigraphic position of the Tremp localities is well defined by means of the marine biostratigraphy and palaeomagnetism (see figure 3). As it has been stated by Molina *et al.* (1992) and Serra-Kiel et al. (1994), the base of the Ilerdian belongs to the NP 9 nannoplancton zone, and is included in the chron 24R, thus Palaeocene in age. The Palaeocene-Eocene boundary is situated in the Middle Ilerdian (base of zone P6), about 150 m above the Ilerdian lower limit. The localities of Claret 4, Tendruy and Palau, situated about 30 m below the base of the Ilerdian, belong to the first Ilerdian sedimentary sequence (Krauss, 1990). In this sequence, Serra-Kiel *et al.* (1994) interpret the lack of the chron 25N as an *hiatus* in the sedimentation. In the Tremp basin, another *hiatus* affects the Palaeocene record as shown in the figure 3. This *hiatus* is related to the widespread presence of a sedimentary rupture (the "Tremp-Colmenar horizon", Eichenseer, 1988), which appears like a soil catena and evaporite collapse breccias approximately situated between the Danian and the Thanetian.

The mammalian biochronology for the Palaeocene in Europe is still unstable due to the lack of fossil successions. Nevertheless, a major division seems well established between the Cernaysian (MP 6) and the Sparnacian (MP 7 to 9) mammal ages, separated by the first appearance of marsupials, carnivores, rodents, bats, dermopteres, euprimates, artiodactyls and perissodactyls. The localities of Dormaal and Silveirinha belong to the Early Sparnacian (MP 7) because of the presence of these immigrants. The localities of Cernay and Berru (Paris basin) belong to the Cernaysian (MP 6) because of the absence of these immigrants. In North America, the immigration event is less sharp. Carnivorous and dermoptera were already present during the Tiffanian; afterwards, rodents arrived in the Clarkforkian mammal age, before a larger migratory wave in the Wasatchian.

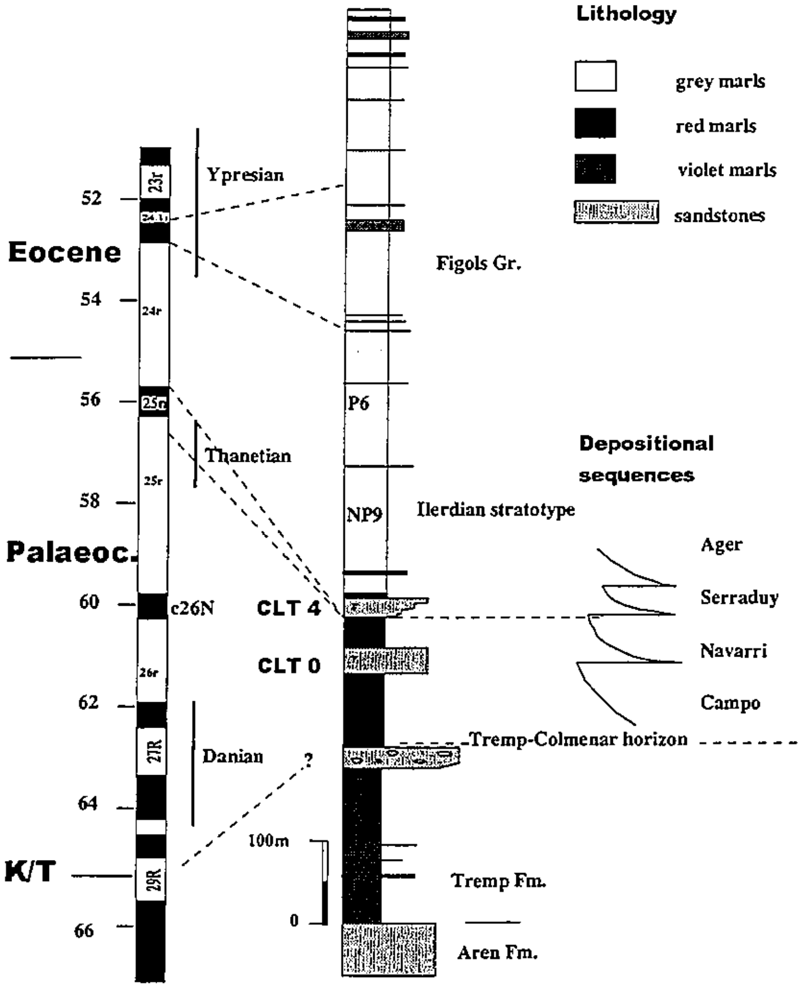

Figure 3. - Stratigraphical situation of Claret 0 and Claret 4 mammalian localities in a section from Northwest Tremp to Montllobar (Ilerdian stratotype). Biozones of nannoplancton and planctonic foraminifera, and correlation with the Geomagnetic time scale are indicated according to Serra-Kiel *et al.* (1994). Depositional sequences are from Krauss, 1990 and Eichenseer, 1988.

The boundary between Cernaysian and Sparnacian in Europe and that between the Clarkforkian and Wasatchian in North America have been usually correlated to each other and with the Palaeocene-Eocene boundary. This correlation has been rejected by Hooker (1991). According to this author, the Palaeocene-Eocene boundary (marine boundary between zones NP9-NP10) correlates with the middle Sparnacian (MP 8), and not with the base of it. This statement has been confirmed by Sinha *et al.* (1995), which situate the first Sparnacian levels from the Paris basin in the Late Palaeocene, following the position of the ^{13}C anomaly.

A problem arise when trying to include the Palaeocene localities of the Tremp Formation in such a scheme. The presence of several species in Tremp close to that of the Dormaal assemblage, such as *Paschatherium cf. dolloi, Microhyus cf. musculus* and a lophodont condylarthre close to *Landenodon*, strongly suggests they belong to the MP 7 unit (Early Sparnacian). On the other hand, the presence of the taxa cf. *Hainina* and *Afrodon* indicates an older age, either MP 5 or MP 6 (Cernaysian). The complete absence in the Tremp localities of the immigrant mammalian orders mentioned before, favours also an age older than MP 7. Thus the composition of the Tremp Palaeocene mammal assemblages shows a mixed character between MP 5-6 and MP 7 taxa. Including these assemblages in the biochronological scale would imply the definition of an intermediate unit (MP 6b, or MP 6-7).

The European Late Palaeocene mammals are chronologically ordered by external criteria, such as marine biostratigraphic or paleomagnetic correlations. If this succession holds, their biochronology shows an anomalous pattern. Some are dominated by *Paschatherium* populations, which shows strong fluctuations in their abundance and an erratic pattern of evolution. From older to younger, Walbeck has no *Paschatherium* species; Campo (nannoplancton zone NP8) has a derived *Paschatherium* dominating a poor assemblage; Cernay and Berru (MP 6, nannplancton zone NP9) hardly have any *Paschatherium* documented, in spite of very large samples with thousand specimens (Russell, 1964); the Tremp localities with mixed MP6-MP7 taxa are also dominated by *Paschatherium*, one of the species being similar to *Paschatherium dolloi* from Dormaal (Early Sparnacian, MP 7). In different localities of this later age, *Paschaterium* becomes widespread and evolves in a mosaic of combination of characters. The European Palaeocene mammal associations thus do not have an internal successional and evolutionary order, and they do not allow an accurate biochronological dating.

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BIOSTRATIGRAPHICAL DISTRIBUTION AND BIOCHRONOLOGICAL SIGNIFICANCE OF EUROPEAN PALEOGENE *CORYPHODON* (MAMMALIA, PANTODONTA)

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Three valid species of the large pantodont genus *Coryphodon* are present in European Paleocene-Eocene strata: *C. eocaenus* Owen, 1845, *C.anthracoideus* Blainville, 1846 (=*C. gosseleti* Malaquin, 1899) and *C. oweni* Hébert,1856. The three species are readily recognized from third molars (Fig. 1) and can be diagnosed from each other as follows:

1. *C. eocaenus* is a medium-sized species of *Coryphodon* (length M1-3 is about 79 mm). The large m3 entoconid and complete lingual cingulum on M3 also distinguish *C. eocaenus* from the larger *species C.anthracoideus* and from the smaller species *C. oweni.*
2. *C. anthracoideus* is a moderately large species of *Coryphodon* (length M1-3 is about 93 mm) morphologically identical to but significantly larger than *C. oweni.* The relatively small m3 entoconid, lack of a complete lingual cingulum on M3 and generally larger size *of C. anthracoideus* distinguish it from *C. eocaenus.*

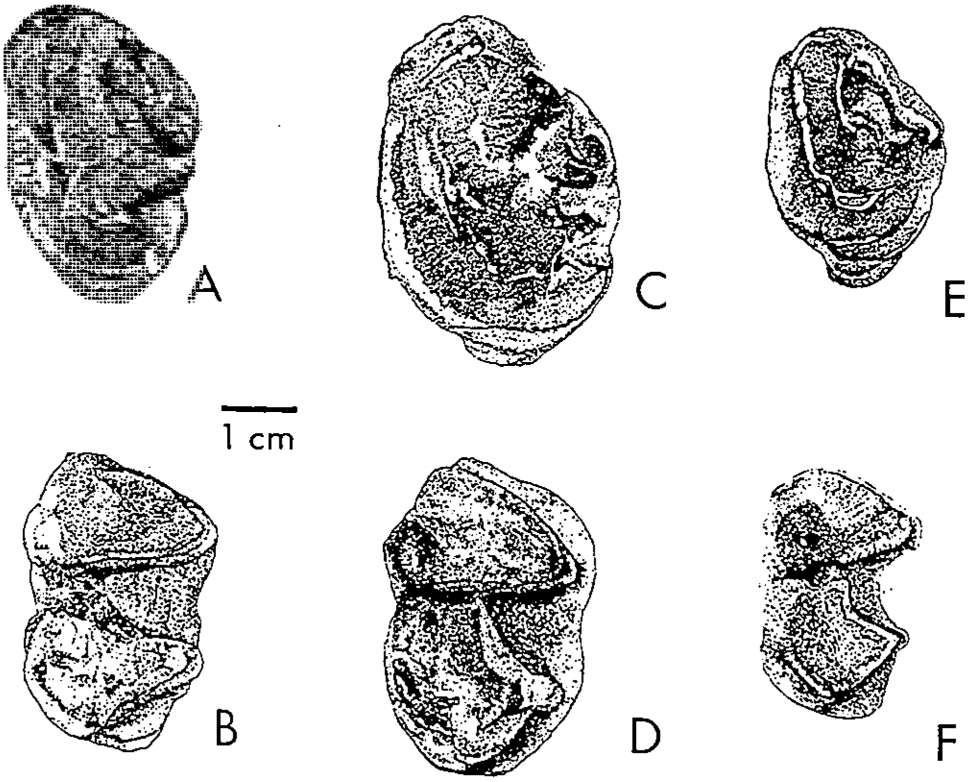

Figure 1. Left M3's and m3's of the three European species of *Coryphodon*. A-B, *C. eocaenus*. C-D, *C. anthracoideus*. E-F, *C. oweni*. A = BMNH M 13775, B= BMNH 27848, C = MNHN ARP 58, D = IRSNM IG 10.018, E = MNHN ARP= 10, F = MNHN ARP 9. BMNH = Natural History Museum, London, IRSNM = Institut Royale des Sciences Naturelles de Belgique, Brussels, MNHN = Museum National d'Histoire Naturelle, Paris.

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3. *C. oweni* is a small species of *Coryphodon* (length M1-2 is about 71 mm) most similar to *C. anthracoideus* but much smaller. The large m3 entoconid, complete lingual cingulum on M3, broader cheek teeth and, in most specimens, overall larger size of *C. eocaenus* distinguish it from *C. oweni*.

The biostratigraphic distribution of *Coryphodon* fossils in England, Belgium and France (Fig. 2) reflects nearly two centuries of collecting. In England, the Woolwich Shell Beds at Dulwich and Sydenham produced fragmentary postcrania and an upper canine of *Coryphodon* (Bott, 1875; Klaasen, 1883; Newton, 1883). The postcrania and the canine are small and may pertain to *C. oweni*. However, the demonstrable sexual dimorphism in canine and body size of *Coryphodon* make it possible that the *Coryphodon* specimens from the Woolwich Shell Beds are female *C. eocaenus*. Indeed, the holotype ulna of *C. "croydonensis"* from the Woolwich Shell Beds near Croydon, Surrey (Park Hill Railway Cutting) is larger than an ulna from Meudon that might, on the basis of its association with teeth of *C. oweni*, belong to that species. Thus, the Woolwich ulna may belong to *C. eocaenus*. However, at present only the identification *Coryphodon* sp. can be assigned with certainty to the *Coryphodon* specimens from the Woolwich Shell Beds.

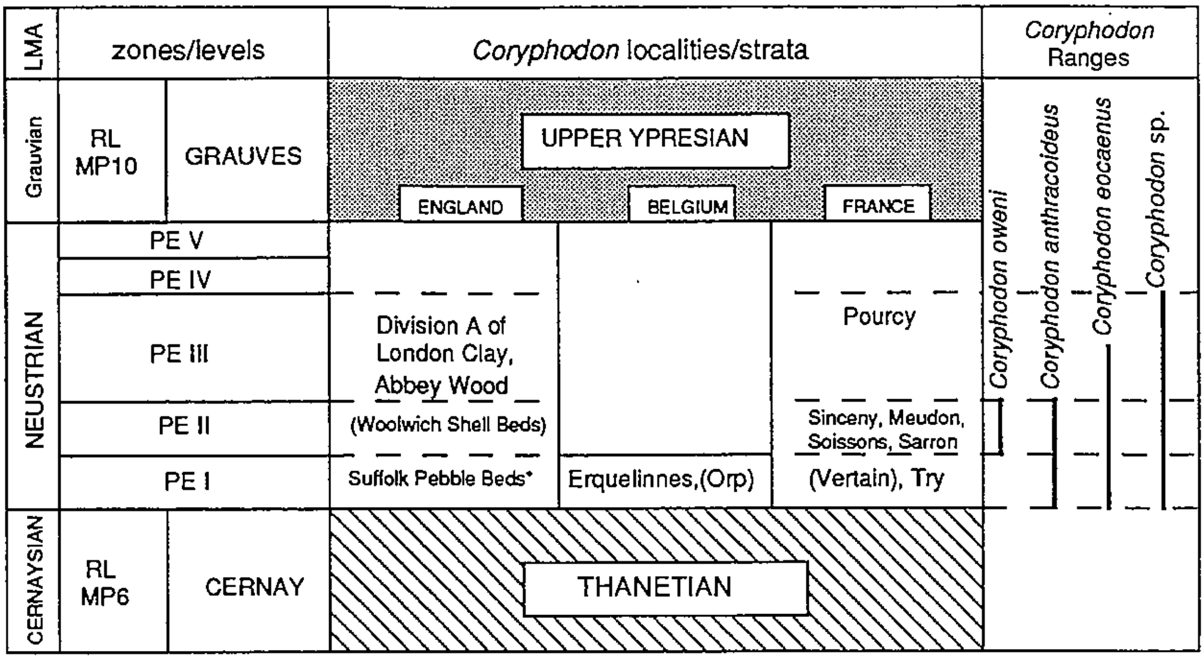
The Blackheath Beds at the famous Abbey Wood Locality have produced the most extensive British fossil mammal assemblage relevant to the P/E Series boundary. *Coryphodon eocaenus* is known from Abbey Wood (Forster Cooper, 1932).

The Suffolk Pebble Beds have produced an assemblage that may (all or in part) be reworked from older strata. Reworked teeth of *C. eocaenus* and *C. anthracoideus* found in the Pliocene Red Crag in Suffolk probably were derived from the Suffolk Pebble Beds and thus reworked twice. The lower part of the London Clay in the vicinity of Barnet(?), and just off the Essex coast near Harwich, produced teeth of *C. eocaenus*, including the holotype lower jaw fragment from the latter locality. The horizon in the London Clay that yielded the holotype was not recorded but is likely to be close to the Harwich Stone Band in division A (Hooker and Insole, 1980).

Dormaal is the best known Belgian fossil mammal locality relevant to the P/E Series boundary. It is the type assemblage of RL MP 7, and has more recently been placed in Biozone PE I, but it lacks *Coryphodon*. The fluvial Erquelinnes Sand Member of the Landen Formation at Erquelinnes (=Jeumont) near the Franco-Belgian border produced an upper premolar and a nearly complete femur that Teilhard de Chardin (1927) identified as *C. eocaenus*. We have not been able to locate the premolar, but the measurements of this tooth reported by Teilhard de Chardin (1927) support assignment to *C. eocaenus*. At Orp-le-Grand, just south of Dormaal, a partial skull, lower jaw and skeleton of *C. anthracoideus* were collected from lignitic beds in the"Sable fluvio-marin" (= upper Landen Formation) that evidently, and probably unconformably, overlies the Tuffeau de Lincent (Lucas, 1986). This too appears to be a record assignable to PE I. Similar fluvio-marine strata at Leval in southern Belgium produced two partial femora and a nearly complete astragalus of *Coryphodon*. Although Teilhard de Chardin (1927) assigned these bones to *C. eocaenus*, they are indeterminate at the species-level.

The most extensive fossil mammal assemblages from western Europe relevant to the P/E Series boundary are from France. The largest of these is from Avenay, which is the type locality for the Reference Level MP 8-9 fauna and belongs to Biozone PE V. This assemblage lacks *Coryphodon* as does the other large assemblage from nearby Mutigny (PE IV). Meudon, in suburban Paris, has long been considered by vertebrate paleontologists to be the oldest "Sparnacian" mammal locality. However, because its mammaliferous strata (Argiles Plastiques with basal conglomerate) sit unconformably on Cretaceous chalk and its fauna shares taxa with more recent sites, it is regarded as younger than Dormaal and is referred to Biozone PE II. Abundant *Coryphodon* from Meudon belong to *C. oweni*, a species not found elsewhere in Europe. Specimens of *C. oweni* were collected from the Conglomerat de Meudon at three or four localities (Meudon, Passy,Vaugirard, Moulineaux?) now within the suburbs of Paris (Hébert, 1856). Some of these teeth, as well as small postcrania that probably pertain to *C. oweni*, may actually have come from argillaceous and lignitic strata less than 1 m above the Meudon conglomerate (Plante, 1870), but provenance records are too nondescript to confirm or deny this possibility.

Two localities just north of Epernay (the type locality of the Sparnacian Stage), Pourcy and Mont de Berru near Cernay (northeast of Reims), purportedly produced specimens of *C. oweni*. Depéret (1906, 1907) originally reported these specimens, but they were not illustrated and cannot be located. The femur and metatarsal from the "Sables et argiles ligniteuses" at Mont de Berru reported by Depéret (1906) were assigned by him to *C. oweni* because of their small size. Depéret (1907) later mentioned a lower premolar he assigned to *C. oweni* from the "Sables supérieurs" at the "Gite de Pourcy". Depéret (1907) did not describe this tooth or justify its identification, so we identify it only as *Coryphodon* sp. The postcrania from Mont de Berru are also identified as *Coryphodon* sp. The Pourcy record may be the youngest European *Coryphodon*. The oldest from the Paris Basin and contemporaneous with those from Belgium is that from the Conglomerat à *Coryphodon* at the base of the Marnes de Dormans at Try (Marne). It is apparently associated with a PE I mammal fauna, although the position of the latter in the sequence is uncertain.



* reworked assemblage; (...) assemblage age extrapolated by superposition

Figure 2. Correlation of European *Coryphodon* localities/horizons of Neustrian mammal Biozones I-V (spanning Reference Levels MP7 and MP 8-9) with ranges of *Coryphodon* species.

Other French occurrences of *Coryphodon* of note are: (1) fluvio-marine sands at Vertain in northern France produced the lectotype M3 and other teeth of *C. "gosseleti"* Malaquin, 1899, which we consider to be a synonym of *C. anthracoideus*; (2) a premaxillary fragment from the Sables de Sinceny at Sinceny (west of Laon along the Oise River) pertains to *C. eocaenus* (Loius and Michaux, 1962; Michaux, 1963); (3) the Argiles a Lignites in the Soissons (Soissonais) and Laon (Laonnois) districts have produced teeth of *C. eocaenus* and *C. anthracoideus*, including the lectotype of the latter; (4) a single incisor from the Argiles a Lignites(?) at Muirancourt (west of Laon along the Oise River) probably pertains to *C. eocaenus*; (5) the Sables de Sinceny in the vicinity of the old railway depot at Sarron along the Oise River have produced teeth of *C.eocaenus* and *C. anthracoideus*; and (6) several teeth of *C. eocaenus* and *C. anthracoideus* were collected from the Argiles Plastiques at an unspecified locality (or localities) that may have been in the vicinity of Paris or, more likely, along the Oise Valley.

Coryphodon ranges can be derived from the European record (Fig. 2). *C. oweni* is restricted to a single biozone - PE II, *C. anthracoideus* ranges through PE I-II, and *C. eocaenus* occurs in PE I-III. Younger records of the genus are specifically unidentifiable, and extinction in Europe seems to have taken place before biozone PE IV. The genus *Coryphodon* defines only the early-middle part of the Neustrian LMA, but its FAD can be used to define the beginning of the Neustrian.

In North America, *C. eocaenus* ranges from late Clarkforkian through middle Graybullian, *C. oweni* is early Graybullian and *C. anthracoideus* is early-middle Graybullian. The LO of *C. eocaenus* may be older in North America than in Europe, but this depends on the exact age of the Pourcy specimen (possibly partly reworked fauna). If the two LO's are equated, then the late Clarkforkian LMA correlates with Biozone PE I. This, however, would make the major evolutionary turnover at the beginning of the Wasatchian and Neustrian LMA's diachronous. It is more likely that the Clarkforkian LMA correlates with RL MP 6, and that the FAD of *C. eocaenus* in North America predates that in Europe.

In Asia, *C. eocaenus* and *C. anthracoideus* have rare records in the youngest Nongshanian ("Gashatan") mammal assemblages. If the centre of origin of *Coryphodon* was North America, as is indicated from cladistic analysis of pantodont genera, Asian sites with *C. anthracoideus* are unlikely to be older than earliest Wasatchian. This suggests that the beginning of the Bumbanian is more recent than the beginning of either the Neustrian or Wasatchian.

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THE INTERNATIONAL SUBCOMMISSION ON PALEOGENE STRATIGRAPHY ACHIEVEMENTS AND PROBLEMS

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The International Subcommission on Paleogene Stratigraphy (ISPS) established in 1972 has three main goals:

(1) to agree on an international set of stages and series for the Paleogene
(2) to establish boundary stratotypes of the Paleogene stages and series
(3) to encourage research into the Paleogene by setting up working groups and regional committees to study and to report on specific problems.

A set of Paleogene stages has been voted and agreed on by the ISPS at the International Geological Congress in Washington in 1989 (Jenkins & Luterbacher,1991). Subsequently, working groups have been set up to find a Global Stratotype Section and Point (GSSP) for the lower boundary of each of these stages. At present, the GSSPs of the Danian (= Cretaceous/Paleogene boundary), of the Rupelian (= Eocene/Oligocene boundary) and of the Aquitanian (= Paleogene/Neogene boundary) have been established and ratified by the International Union of Geological Sciences (Fig.1). The forthcoming proposal of a GSSP for the base of the Ypresian (= Paleocene/Eocene boundary) will leave us with well defined boundaries of the three series (Paleocene, Eocene, Oligocene) of the Paleogene. The next step which is already well underway is the search for GSSPs of the remaining stages. We are optimistic that we may reach this goal within the next five years.

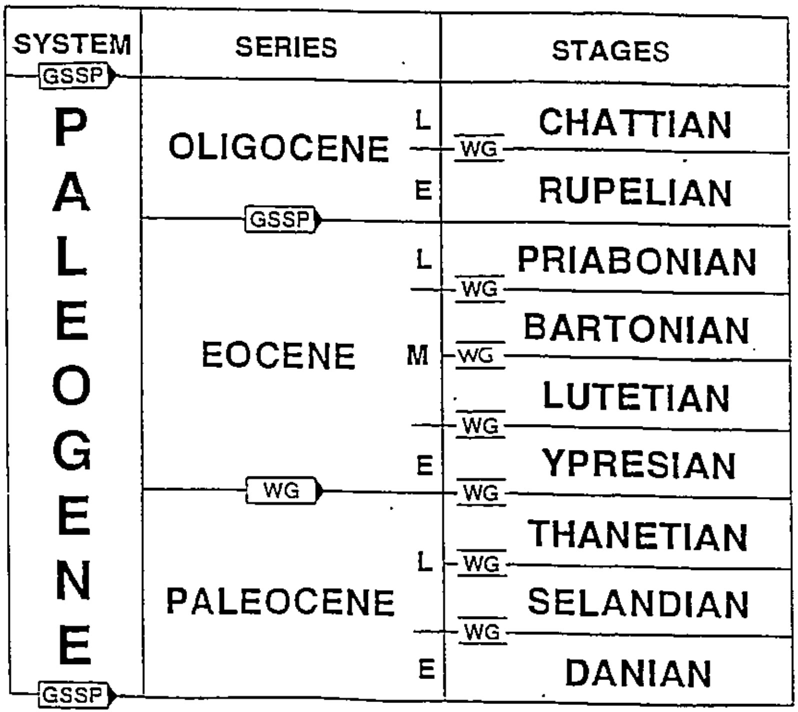

Fig.1- Subdivision of the Paleogene accepted by the International Subcommission on Paleogene Stratigraphy in 1989 (GSSP = established and ratified boundary stratotype, WG = working groups).

Strata Série 1., vol. 9 p 87-88, 1 fig. 1998

The guidelines for the selection of GSSPs have been published by the International Comission on Stratigraphy in 1996 (Remane et al. 1996). The checklist which an ideal GSSP has to satisfy is rather long :

> (1) good accessability, (2) permanent marker ('golden spike'), (3) marine, if possible open-marine succession, (4) continuous deposition and exposure across the boundary, (5) absence of structural or diagenetic complications or redeposition, (6) absence of facies changes across the boundary, (7) good paleontologic record. (8) reliable magnetostratigraphy, (9) chemo-stratigraphy, (10) possibility of direct radiometric dating.

Evidently , the ideal GSSP does probably not exist. Our approach should therefore be somewhat pragmatic and not become an endless search for the holy grail. The boundary stratotype for the Eocene/Oligocene satisfies the majority of the required criteria (Premoli Silva et al., 1988), but is still far from being perfect.

However, the rather formalistic definition of these GSSPs is a byproduct of a much wider and also more rewarding enterprise. The work of the different groups focusing on specific parts of the Paleogene has led (and hopefully continues to do so) to a very impressive increase of our knowledge and understanding of the chronology and interrelation of paleoceanographic, paleoclimatic and evolutionary events, to an improvement of stratigraphic resolution and correlations. This progress in our knowledge is demonstrated by the achievements of the working groups studying the Paleocene/Eocene and the Eocene/Oligocene boundaries (for the latter see e.g. Krasheninnikov & Akhmetiev,1996 ; Pomerol & Premoli Silva, 1986 ; Prothero & Berggren, 1992).

One of the main problems which hamper the work of the ISPS is the lack of funds necessary to support meetings of its different working groups and regional committees. We depend entirely on the goodwill and the initiative of the members and in particular of the leaders of the working groups and regional committees.

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THE PALAEOCENE/EOCENE BOUNDARY IN THE MALAGUIDE DOMAIN. INTERNAL BETIC ZONE (S SPAIN)

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ABSTRACT

In the Palaeocene/Eocene boundary in the Malaguide Domain, nowadays the upper and unmetamorphosed complex of the Internal Betic Zone (Wester Alpine Belt), a stratigraphic unconformity, with a gaps that affects at least the Ilerdian, accounted. A change in the sedimentation both in marine (mixed to carbonatic platform) and continental (fluvial to lacustrine) realms occur. This change is parallel with a climatic change from mediterranean-like to subtropical-like climate that allowed (in the marine realms) the passage from eutrophic condition during the Palaeocene to oligotrophic conditions during the Eocene.

INTRODUCTION

The Betic Cordillera, the more occidental of the European Alpine belts, has classically divided into two tectono-paleogeographic domains (the Internal and the External Zone: Julivert *et al.* 1977), with several differences in the sedimentary and tectono-metamorphic evolution, and in the age in which they have supported the Alpine deformation. The Malaguide Complex, the upper and unmetamorphised nappe of the Internal Zone has support a compressive evolution after the later Cretaceous to the Lower Miocene. Nevertheless, It has been during the Upper Oligocene and the Lower Miocene when the compressive tectonic activity has been stronger (Martín-Martín et al., 1997). The study of the stratigraphic and sedimentologic changes in the Maláguide Tertiary is highly important for the reconstruction of the paleogeography of this Domain, in the period in which it acquired its main tectonic features.

This communication shows the events that accounted during the boundary Palaeocene/Eocene in the Maláguide. This study has been located in the Sierra Espuña area (province of Murcia: SE Spain), where the Malaguide Tertiary is well exposed and with a relative low tectonisation.

THE MALAGUIDE DOMAIN DURING THE PALAEOCENE AND LOWER EOCENE

From a stratigraphic-sedimentologic standpoint, the Tertiary of Sierra Espuña can be divided into sedimentary cycles that include several depositional sequences bounded by unconformities and their laterally correlative conformities towards the centre of the basin. The Palaeocene to early Oligocene sediments were defined by Martín-Martín (1996) as belonging to *The Malaguide Tertiary Lower Sedimentary Cycle*. This mainly preorogenic sedimentary cycle is boundered, at the bottom, by a mayor discontinuity with biostratigraphic gap (Cretaceous/Tertiary boundary), and at the top by a mayor discontinuity with a biostratigraphic gap, related to a compressive tectonic phase that can affect to the underlying malaguide succession. This sedimentary cycle is composed by the Palaeocene, the Cuisian to early Lutetian, the middle Lutetian to Priabonian, and the early Oligocene depositional sequences.

The aim of this communication is to show the sedimentary, biostratigraphic, palogeographic and climatic changes during the periods of the Palaeocene and the Cuisian to early Lutetian depositional sequences.

The Palaeocene Depositional Sequence. It is composed of a stratigraphic-depositional unit. The lower limit of this unit paraconformably overlies the Upper Cretaceous, while the upper limit is marked by the deposits of the Lower Eocene (the Cuisian to early Lutetian depositional sequence), at the base of which a paraconformity is recognized, associated with a stratigraphic gap which consistently affects at least the Ilerdian that is missing in the Malaguide Domain. Two kinds of succession can be recognised in the area (Fig. 1):

a) Basin towards, the succession is formed by ritmic deposits. At the bottom, it is made of calcareous sandstones containing *Microcodium*, *phosphates*, quartz and abundant bioclasts, of microconglomerates and even by calcareous conglomerates of dark colours, which alternate with levels a few centimetres thick of azoic bluish sandy marls and silts. This level is early Palaeocene in age (Danian). In the middle part of the succession are present calcarenites or calcirudites of quartz pebbles, lydite and limestone, massive or with parallel stratification and dark grey colour. In this level are present some larger foraminifera that indicate an early Palaeocene (Selandian) age. Finally, at the top, are present calcarenites and marly levels analogous to those of the bottom of the succession, with the exception that the marls present plancktonic fauna that indicates a late Palaeocene (Thanettian) age.

b) In the marginal zones of the basin are present the "Garumnian-like" facies, formed by red lutites with *Microcodium*, conglomeratic intercalations fed from the malaguide mesozoic, and calcarenites with *Microcodium*. All this succession can be interpreted as the deposits in continental environments.

The facies and the fossils of the Palaeocene indicate that this is a deposit in a mixed carbonate-siliciclastic platform with abundant contributions from a continent in which was developed a fluvial-like sedimentation. The presence of *Microcodium* indicates that the continental areas located near the marine environment, supported a vegetal cover weathered under a mediterranean-like climate (Esteban and Klappa, 1983)). On the other hand, the presence of phosphate, and the scarce presence of larger foraminifera, indicate that the marine zone underwent eutrophic conditions (Hallock, 1988) due to upwelling of deep marine currents to this platform.

This malaguide palaeocene sedimentation can be divided into three successive stages: a rapid transgression, accompanied by an erosive phase and the deposit of distal facies; progradation of the environments closest to the platform; and finally, the develop of distal and deepest marine facies, caused by a new transgression.

The Cuisian to early Lutetian Depositional Sequence. This depositional sequence boundered, at the bottom by a serious discontinuity with a biostratigraphic gap that affect at least to the Ilerdian (missing in the Malaguide Domain), is constituted by two laterally related kinds of facies (Fig. 1):

a) In the basinal part, the succession is mainly composed by alveolina-rich cream-coloured sandy limestones, nummulites-rich cream-coloured and white limestones, and cream coloured and white algae limestones. This sediment represent the sedimentation in a inner carbonatic platform with a great development of larger foraminifera that indicates oligotrophic conditions (Hallock, 1988).

b) In the marginal part of the basin, and mainly upward related to the above facies are present grey-blue marls and grey marly limestones with gastropod-rich and lignite-bearing levels. These types of sediments, described for the Malaguide Complex by Serrano et al. (1995), are interpreted as lagoon to marsh deposits. Sometimes alveolina-rich calcareous levels are intercalated within these deposits that allow to date this sediment as Cuisian to early Lutetian. In the lignite levels palinologic studies have been done. The species found indicate tropical climatic conditions (Kedves *et al.*, 1997) in good accordance to the presence of crocodile dents in the same lignitous levels (Ruiz-Bustos: *pers. com.*).

The above two kinds of sediments represent a great transgressive marine evolution onto the Palaeocene and Mesozoic emerged sediments, and later, a shallowing upward evolution for the depositional sequence.

CHANGES IN THE PALAEOCENE/EOCENE BOUNDARY IN THE MALAGUIDE DOMAIN

Some significative changes have accounted in the Palaeocene/Eocene boundary in the Malaguide. They will be exposed following:

1.- A stratigraphic gap occurs in this boundary that affect the Ilerdian. Although this period is well represent in others paleogeographic domains in the Alpine belt (Pirinean belt: see Martín-Martín et al., 1997, and references therein), it is absent in the Malaguide Domain. This allows to exclude a glacioeustatic change to explicate this gap. Regional changes in the plate tectonics must be proposed, since, during this period, a general change in the tectonic regime occurred in the western alpine belts and specially in the Malaguide Domain. This Domain was located during the early Tertiary in a Southeastern position respect to the actual coordinates (Martín-Algarra, 1987). This change transformed the tethysian divergent continental margins into compressive areas. Although in the lowers units of Internal Zone compressive phenomenons allowed the tectogenesis and the eoalpine metamorphism (Zeck et al., 1989), in the surface of this domain, preorogenic conditions yet accounted. Only slightly variation in the basin form occurred that allowed the developing of unconformities in the basin boundaries and local changes of facies.

2.- In this boundary in the continental realms a change from fluvial conditions to lacustrine condition accounted, meanwhile, in the marine realms a change from a mixed platform to a carbonatic platform is noticed.

3.- In the biologic condition in the marine platform, a change from eutrophic to oligotrophic conditions have been noticed above.

3.- The two previous changes (kind of sedimentation and biologic condition, respectively) must be related to a climatic change that conditioned the precipitations, the weathering and, by hence, the runoff from the continent and the upwellings from the deep marine zones. This data are in good accordance to the climatic change suggests above by mean of the presence of *Microcodiun* (in the Palaeocene sediments), and the continental palynology and the presence of crocodile dents (in the Lower Eocene materials): from stational mediterranean-like to subtropical-like climates.

ACKNOWLEDGEMENT

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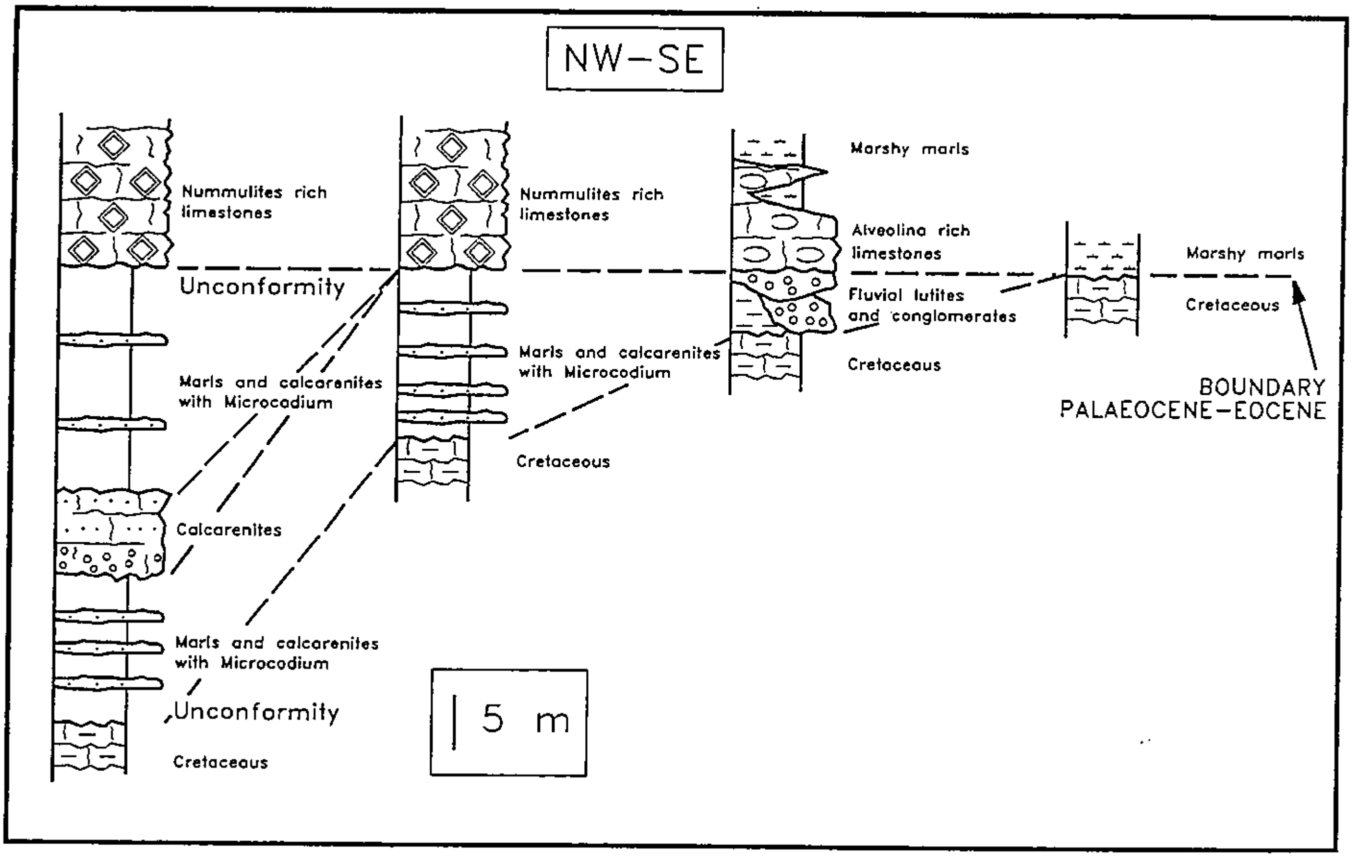
FIGURE CAPTION:

Figure 1: Litostratigraphic columns showing the Palaeocene-Eocene boundary in the Sierra Espuña area, and the lateral changes of facies of the palaeocene and lower eocene sediments.



PLANKTIC FORAMINIFERAL BIOSTRATIGRAPHY ACROSS THE PALEOCENE/EOCENE BOUNDARY : EVENTS AND CORRELATIONS

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A study of planktic foraminiferal assemblages of different sections in the Tethys and North Atlantic realms allows us to to revise the biostratigraphical zonations and to establish the correlation among biozones, sections and the geochronological time scale of Berggren *et al.* (1995). Many sections were studied in Belgium, Bay of Biscay, France, Israel, Italy, Kazakstan, Spain and Tunisia in order to search for the Paleocene/Eocene (P/E) Boundary Stratotype. The best ten sections were studied at high resolution and their results are summarized in this synthesis. Other sections, such as the Ypresian stratotype in Belgium (Pardo *et al.*, 1994) and the Aïn Settara section in Tunisia, have low planktic foraminiferal richness and were not included in this synthesis, although their results were taken into consideration. The best studied sections are located in Spain (Alamedilla and Caravaca in the Betic Cordillera and Zumaya, Campo and Tremp in the Pyrenees), the Bay of Biscay (DSDP Site 401), Italy (Bottaccione and Possagno), Israel (Ben Gurion) and Kazakstan (Kaurtakapy).

The quantitative study of the planktic foraminifera in all these sections allows us to identify several datum events along the P/E transition. These events are: first occurrence (FO) of *Muricoglobigerina soldadoensis*, last occurrence (LO) of *Luterbacheria pseudomenardii*, LO of *Igorina laevigata*, FO of *Acarinina berggreni*, *Acarinina sibaiyaensis* and *Acarinina africana* apparently simultaneous. LO of *Morozovella velascoensis* and FO of *Morozovella formosa*. These events have been used to recognize the biozonation of Berggren *et al.* (1995) that have been modified by Pardo *et al.* (in press) subdividing the P5 zone in two subzones: P5a of *Luterbacheria pseudomenardii-Acarinina sibaiyaensis* and P5b of *Acarinina sibaiyaensis/Morozovella velascoensis*. Some of these events have been used to establish the biozonation of Arenillas and Molina (1996) that have been slightly modified in this paper subdividing the *Luterbacheria pseudomenardii* in two biozones: *Luterbacheria pseudomenardii* Biozone and *Muricoglobigerina soldadoensis* Biozone. These subdivisions are useful in order to have a more detailed biostratigraphical scale for correlation between different regions and basins.

The planktic foraminiferal faunal turnover across the P/E boundary is quite gradual except at the benthic foraminiferal extinction event (BFEE) horizon. This extinction of the bathyal and abyssal small benthic foraminifera has been found in all deep sea sections just below a dissolution clay interval. This relevant extinction event and the concomitant lithological change from marls to clay constitutes an apparently very isochronous datum, which would be very suitable for the definition of the P/E Boundary Stratotype (Molina, 1996). The BFEE coincides with the LO of *Igorina laevigata* and the beginning of the increase of acarininids. In this interval, coincindent with the BFEE, an acarininid extra-tropical excursión is observed in middle and high latitudes. This excursion indicates increased sea waters temperatures coincident with the dissolution interval. The FO of the new acarininids (*A. africana, A. berggreni* and *A. sibaiyaensis*) is found just above the dissolution clay. An increase of the chiloguembelinid population has been found also above the dissolution clay, indicating hypoxic conditions in intermidiate waters.

The Caravaca section consists of marls at the P/E transition and it is not well exposed. Its foraminiferal assemblages were first studied by Von Hillebrandt (1974) establishing the correlation between planktic and large benthic. Later on, this section was evaluated as a potential P/E boundary stratotype by Molina *et al.* (1994) based on a high resolution and integrated stratigraphical study. Planktic foraminifera are very abundant and quite well preserved but this section is not optimal as potential stratotype for the P/E boundary, mainly because the likely presence of a short hiatus 2.5 m below the BFEE.

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The Alamedilla section is probably the best exposed and most continuos section in Spain. The preservation is not as good as in Caravaca, but planktic foraminifera are abundant (Arenillas and Molina, 1996) and have been correlated with the isotopic and sediment compositional changes (Arenillas *et al.*, 1996; Lu *et al.*, 1996). All the biozones from *M. soldadoensis* to *M. formosa* are present and there is no evidence of hiatuses in the P/E transition. A very distinctive red clay interval marks the BFEE at the bottom and the extra-tropical excursion at the top. Nannofossils and magnetostratigraphy are currently under study.

The Zumaya section is a quite expanded section especially from the *M. subbotinae* Biozone which consists of about 120 m of marl and calcarenitic sediments. Nevertheless, the P/E transition is not so calcarenitic and expanded. Von Hillebrandt (1965) first published the biostratigraphy of the whole section based on planktic foraminifera. Planktic foraminiferal faunal turnovers and biochronostratigraphy of the P/E transition were studied by Canudo and Molina (1992) and Canudo *et al.* (1994). The high-resolution iridium, isotopic, nannofossil and foraminifera correlations have been establish by Schmitz *et al.* (1997).

The Campo section constitutes the parastratotype of the Ilerdian stage and consists of platform facies at the base. The section is extremely expanded with a 70 m thick *M. velascoensis* Biozone (Arenillas and Molina, 1995). Although planktic foraminifera are scarce their assemblages were studied by several authors (Von Hillebrandt, 1965; Molina *et al.*, 1989; Canudo *et al.*, 1989). Many other microfossil groups are abundant (e.g., alveolinids, nummulitids, ostracodes, dinoflagellates, etc.) and the integrated stratigraphy across the P/E boundary have been revised and refined by Molina *et al.* (1992).

The Tremp section is the stratotype of the Ilerdian stage and consists of a shallow marine transgressive-regressive megasequence intercalated within terrestrial facies. Scarce planktic foraminifera are only present just in the middle part of the section, which represents the P-E transition. Planktic foraminiferal assemblages were studied by Gartner and Hay (1962), Von Hillebrandt (1965), Blow (1979) and Molina *et al.* (1992; 1995). The Tremp and Campo are very good reference sections for correlation between shallow environments and the deep see sections.

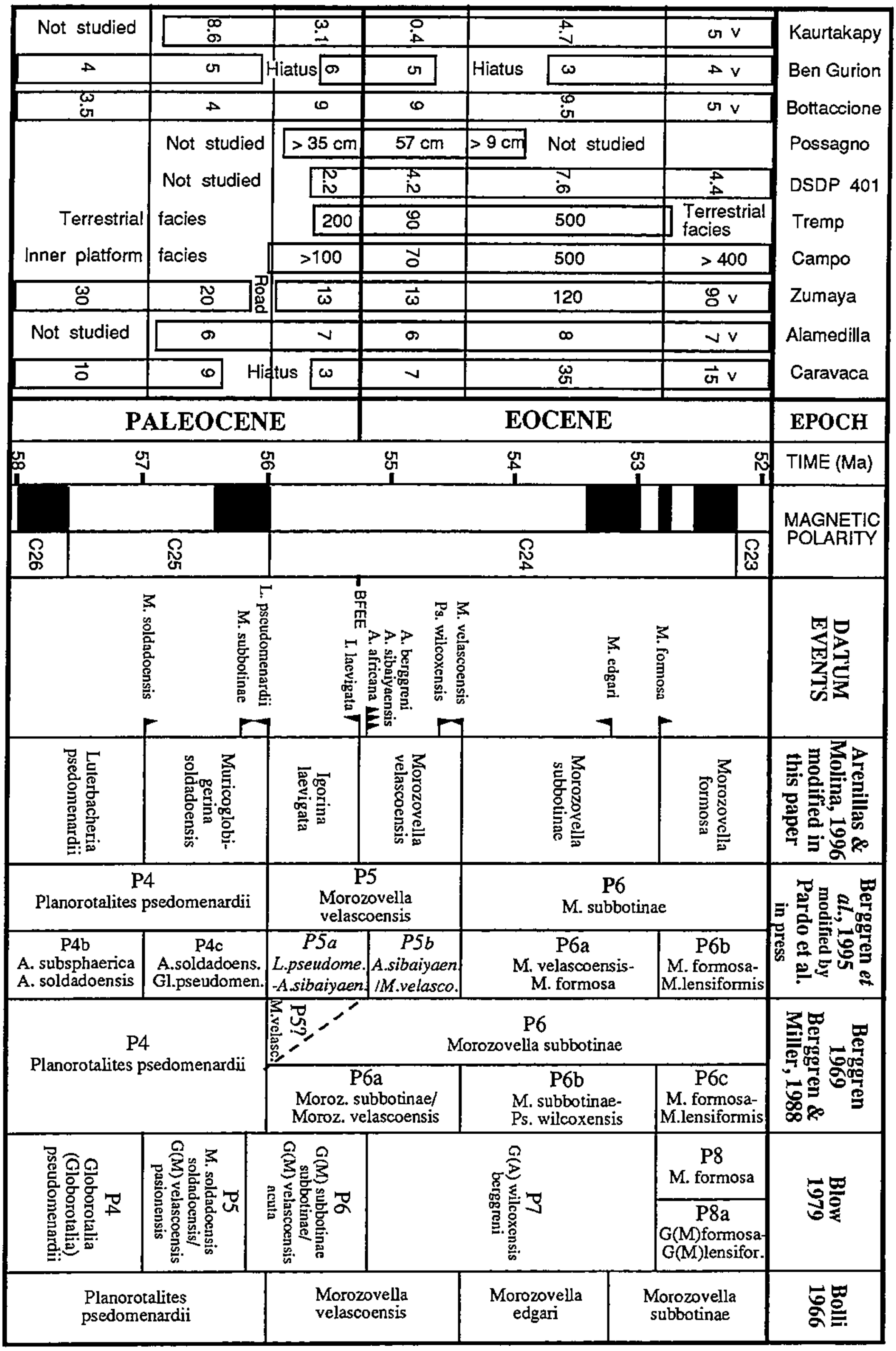
The DSDP Site 401 is located in the western abyssal plain of the Bay of Biscay. The P-E transition ranges from Zone P5 to P8, with some artificial hiatuses due to core recovery. Planktic foraminiferal assemblages show good preservation and diagenetic alteration is low. $\partial^{13}C$ analysis shows a shift coincident with a $\partial^{18}O$ shift, the BFEE and an increase in relative abundance of warm water acarininids and morozovellids. After the BFEE and the isotopic shift, chiloguembelinid population shows an increase, suggesting the onset of hypoxic conditions in intermediate waters (Pardo *et al.*, 1997).

The Possagno section micropaleontological studies were summarized in a monograph edited by Bolli (1975). This classical Italian section is located in a quarry, which is a major problem to consider it as a good candidate for the P/E boundary stratotype. Furthermore, the section is very condensed with a 4 cm thick red dark clay and a 57 cm thick *Morozovella velascoensis* Biozone. Nevertheless, we have identified the acarininids extra-tropical excursion in coincidence with the $\partial^{13}C$ shift.

The Bottaccione section, located in Gubbio, is another classical Italian section. Planktic foraminiferal assemblages from the Paleocene and lower Eocene were first studied by Luterbacher (1964). Preservation of planktic foraminifera is poor since the lithology is very calcareous, but nevertheless this section allows correlation between biostratigraphy and magnetostratigraphy.

The Ben Gurion is one of the best Israeli sections. Planktic foraminifera are frequent and were first studied by Benjamini (1980). Several dissolution levels are present across the P-E transition and two hiatuses have been recognized. Nevertheless, all the biozones from the *M. soldadoensis* to *M. formosa* Biozone have been identified.

The Kaurtakapy section, located in the boreal Paratethys, is a composite of two outcrops which are correlated by means of a 10 cm thick clay layer with marcasite nodules. Planktic foraminiferal assemblages are well preserved and characterized by a sharp increase in relative abundance and species richness of tropical acarininids, morozovellids and igorinids during the P-E transition. The maximum diversity of subtropical species, and the FO of *Acarinina sibaiyaensis* and *A. africana* coincides with a $\partial^{13}C$ shift and the onset of the clay layer. Prior to the main $\partial^{13}C$ shift a long term gradual decrease characterizes the upper Paleocene (Pardo et al., *in press*). The P5b Zone is only 0.4 m thick and suggests the presence of a hiatus.


Kaurtakapy
Not studied
8.6
3.1
0.4
4.7
5 v
Ben Gurion
4
5
Hiatus
6
5
Hiatus
3
4 v
Bottaccione
3.5
4
9
9
9.5
5 v
Possagno
Not studied
> 35 cm
57 cm
> 9 cm
Not studied
DSDP 401
Not studied
2.2
4.2
7.6
4.4
Tremp
Terrestrial facies
200
90
500
Terrestrial facies
Campo
Inner platform facies
>100
70
500
> 400
Zumaya
30
20
Road
13
13
120
90 v
Alamedilla
Not studied
6
7
6
8
7 v
Caravaca
10
9
Hiatus
3
7
35
15 v
PALEOCENE
EOCENE
EPOCH
58
57
56
55
54
53
52
TIME (Ma)
MAGNETIC POLARITY
C26
C25
C24
C23
DATUM EVENTS
M. soldadoensis
M. subbotinae
L. pseudomenardii
BFEE
I. laevigata
A. berggreni
A. sibaiyaensis
A. africana
M. velascoensis
Ps. wilcoxensis
M. edgari
M. formosa
Arenillas & Molina, 1996 modified in this paper
Luterbacheria psedomenardii
Muricoglobigerina soldadoensis
Igorina laevigata
Morozovella velascoensis
Morozovella subbotinae
Morozovella formosa
Berggren et al., 1995 modified by Pardo et al. in press
P4
Planorotalites psedomenardii
P5
Morozovella velascoensis
P6
M. subbotinae
P4b
A. subsphaerica
A. soldadoensis
P4c
A.soldadoens.
Gl.pseudomen.
P5a
L.pseudome.
-A.sibaiyaen.
P5b
A.sibaiyaen
/M.velasco.
P6a
M. velascoensis-
M. formosa
P6b
M. formosa-
M.lensiformis
Berggren 1969
Berggren & Miller, 1988
P4
Planorotalites psedomenardii
P5?
M.velasc.
P6
Morozovella subbotinae
P6a
Moroz. subbotinae/
Moroz. velascoensis
P6b
M. subbotinae-
Ps. wilcoxensis
P6c
M. formosa-
M.lensiformis
Blow 1979
P4
Globorotalia (Globorotalia) pseudomenardii
P5
M. soldadoensis soldadoensis/
G(M) velascoensis pasionensis
P6
G(M) subbotinae subbotinae/
G(M) velascoensis acuta
P7
G(A) wilcoxensis berggreni
P8
M. formosa
P8a
G(M)formosa-
G(M)lensifor.
Bolli 1966
Planorotalites psedomenardii
Morozovella velascoensis
Morozovella edgari
Morozovella subbotinae

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CALCAREOUS NANNOFOSSIL EVENTS AND TURNOVER AROUND THE PALEOCENE/EOCENE TRANSITION IN SOME SECTIONS OF SPAIN AND ITALY

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Several stratigraphic sections straddling the Paleocene/Eocene transition at different latitudes and paleogeographic settings have been investigated to improve calcareous nannofossil biostratigraphy and taxonomy. Semiquantitative analyses have been also performed in order to obtain some paleoecological information and to better correlate the observed fluctuations.

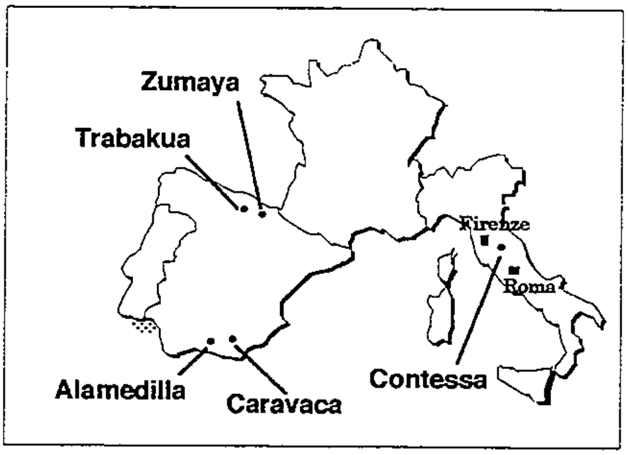

Fig. 1 Location map of the studied sections

The investigated sections have been studied with a multidisciplinary approach by several scientists in order to integrate the different events and characterize the important geochemical and biological changes that influenced the Paleocene/Eocene transition. In this report we present only the calcareous nannofossil results. Six sections have been investigated: Zumaya and Trabakua sections in the Basque-Cantabrian basin (northern Spain), Alamedilla and Caravaca sections in the Betic Cordillera (south-east Spain) and Contessa Road and Contessa Highway sections in the Umbria- Marche basin (central Italy) (Fig.1).

The Paleocene/Eocene transition is characterized by a major turnover in the calcareous nannofossil assemblages with several first and last occurrences. In particular, the entry of the genus *Rhomboaster* and *Tribrachiatus* and their evolutionary trend has been used by Martini (1971), Okada and Bukry (1980) and Aubry (1996) to define zones and subzones and to mark the Paleocene/Eocene boundary in terms of calcareous nannofossil biostratigraphy. In these last years, several authors (Bybell and Self Trail, 1995; Angori and Monechi (1996); Aubry 1996; Aubry et al 1996; Wei and Zhong 1996) have revised the taxonomy of the genera *Rhomboaster* and *Tribrachiatus*, but a general agreement has not yet been reached. Consequently, the lower boundary of NP10 may be placed differently (Fig.2).

The evolutionary trend of *Rhomboaster-Tribrachiatus* has been recognized in all the analysed sections. Three specific variations of *R. bramlettei* described by Angori and Monechi (1996) and *Tribrachiatus digitalis*, used by Aubry 1996 to subdivide the NP10 have been also found. Discoasters show an important diversification with several first occurrences: *D. mediosus, D. araneus, D. diastypus , D. mahmoudi, D. salisburgensis* and *D. binodosus*; the first occurrence of *Discoaster diastypus* has been used to define the CP8/CP9 zonal boundary of Okada and Bukry (1980) . On the contrary, the diversity and the abundance of *Fasciculithus* decrease notably and progressively untill their demis. The last species to get extinct are *F. thomasi* and *F. tympaniformis*. The last occurrence of the genus *Fasciculithus* has been also used to approximate the NP9/NP10 zonal boundary when *Rhomboaster bramlettei* is missing (Berggren et al 1995). Unfortunately, this approximation is not always correct.

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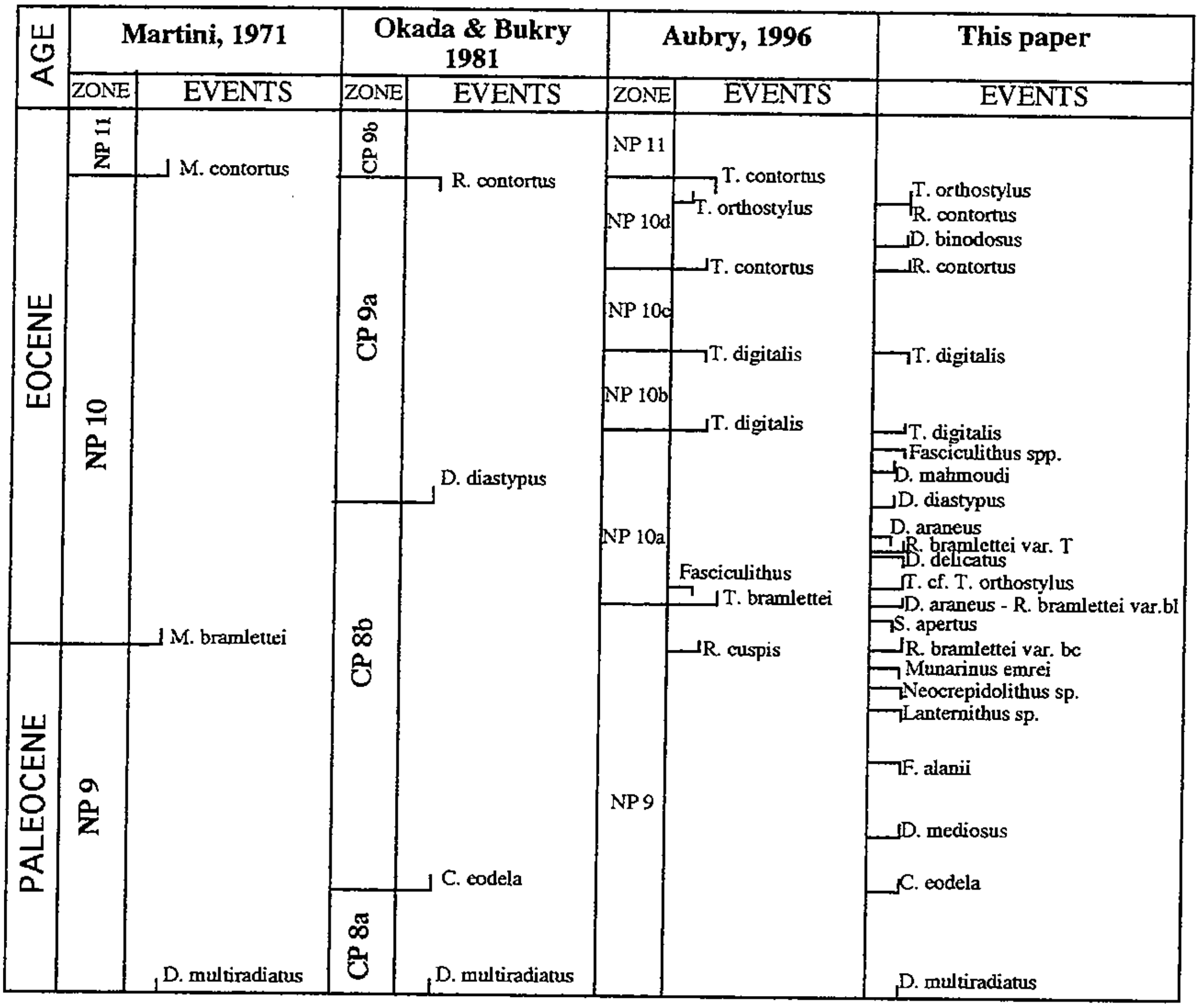

Fig. 2 Comparison of the major calcareous nannofossil zonation schemes and biostratigraphic events spanning the Paleocene/Eocene boundary

High resolution biostratigraphy has allowed to recognize and correlate a succession of 20 well differentiated events. Comparison between sections has revealed that Zumaya and Alamedilla are the most expanded and continuous even if show differences in distribution patterns of some genera as *Fasciculithus*, *Discoaster* and *Neocrepidolithus*. Fluctuactions in abundance of important paleoecological species has confirmed a warming trend around the Paleocene/Eocene boundary and revealed a peculiar distribution of fasciculiths, discoasters, holococcoliths and pentaliths that enhance additional information on paleoenvironmental changes.

Correlation s among calcareous nannofossil events, other fossil groups and geochemical features show that the lowest presence of *Rhomboaster bramlettei* (base of zone NP10) in all the studied sections just below the benthic foraminiferal extinction (BFE) and the carbon isotope excursion (CIE). In the Trabakua section the first occurrence of *R. Bramlettei* cannot be precisely placed relatively to the BFE and CIE due to the presence of a very extended dissolution interval, where nannoflora is absent (Fig. 3).

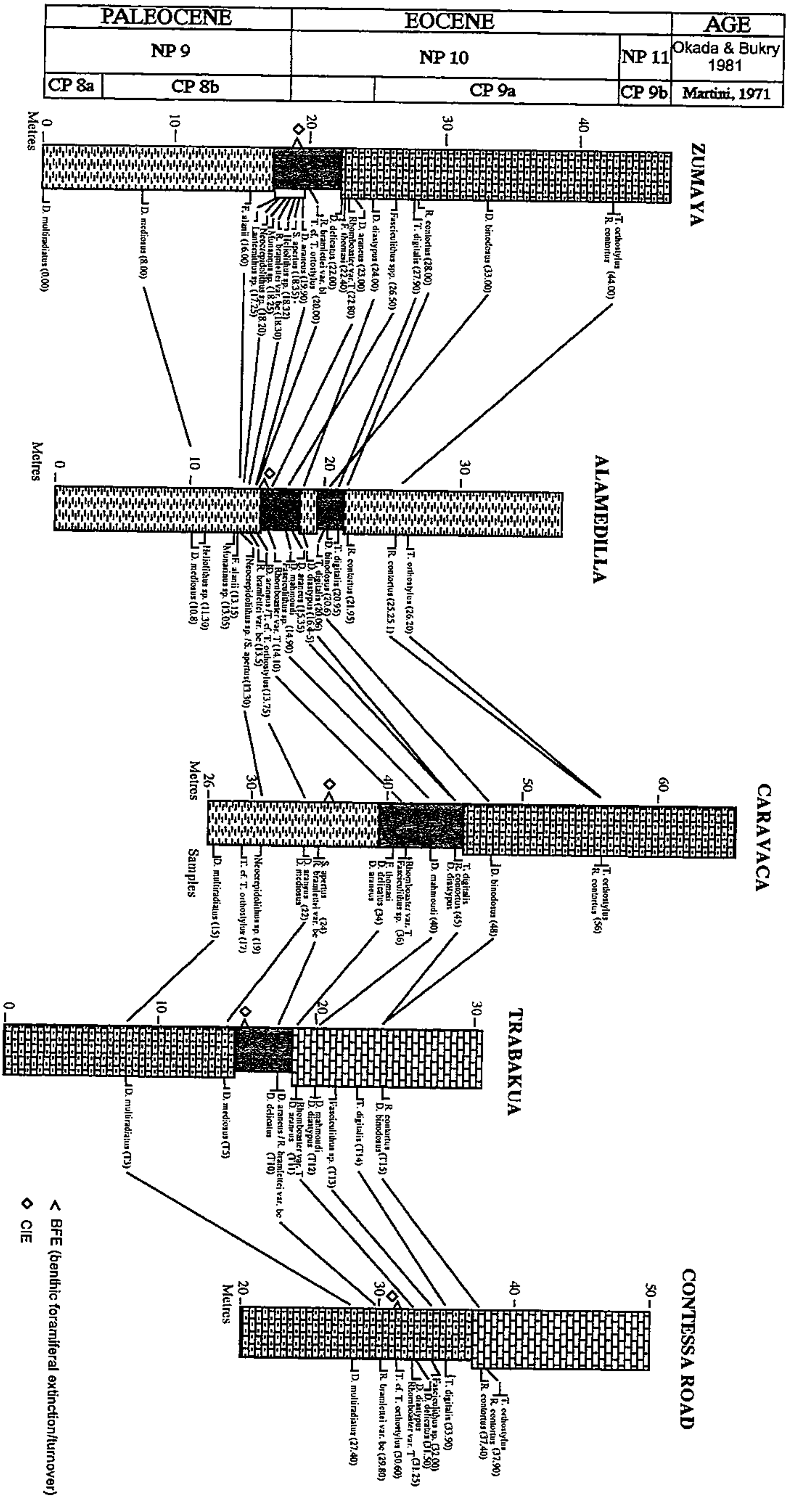

Fig. 3 Correlation of the major events recognized in the studied sections

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LATE PALEOCENE TO EARLY EOCENE BIO -, SEQUENCE -, AND VOLCANIC EVENT STRATIGRAPHY : CORRELATION OF BELGIUM AND NORTH SEA BASIN

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A litho -, bio -, volcanic event -, and seismic/sequence stratigraphy NS correlation (Fig.1, line A-A') is presented for the thick (deep water, more or less complete) succession of the central North Sea Basin, and the thinner (shallow water, and incomplete) lithocolumns of Belgium, allowing to recognise the major trans-, and regressive cycles in the area. Finer cycles (cf. Neal, 1996) are still debated.

Over the last three decades numerous North Sea wells have been analysed a.o. by the oil industry, and these (now largely published) results allow to present a detailed compilation of the Late Paleocene to Early Eocene strata of that area (Fig. 2).

Similarly a compilation is presented for the stratigraphic framework of the Belgian lithocolumns, including the more southern (strongly incomplete) outcrop sections and the (somewhat more complete) northern cored well sections (Fig.3). This scheme is of importance as it comprises the lithocolumns of the Ypresian in its type area (as e.g. compiled by Willems & Moorkens, 1991), and of the underlying Paleocene - Eocene transitional strata, which now appear of importance for the actual discussion on the P/E boundary. The present paper concentrates on this P/E transitional interval.

The juxtaposition of the two above mentioned schemes (Figs.2 & 3) leads to the correlation of the Belgian and Central North Sea sections (Fig.4), thus providing a detailed bio-, litho-,seismic / sequence -, and volcanic event stratigraphy of the North Sea Basin and contiguous areas.

Although the recognition of the now much discussed d13C peak level appears very difficult in the North Sea Basin sections - owing to the strongly decalcified nature of the sediments of that interval - it now appears possible to indicate the approximate position of that level in North Sea wells: This level approximately coincides with the appearance (FOD) of *Apectodinium augustum*, and of some other *Apectodinium* spp. Recently Heilmann-Clausen (1997), and H.-C. & Egger (in preparation; results as presented at the RCNPS meeting of Sept. 1997 at Karlslunde), found *A. augustum* in the NP9 calcareous nannoplankton zone in Austria, there coinciding with non-decalcified / volcanic ash layers and ranging approx. up to the NP 9/10 boundary. Since these outcrop sections appear to be well correlatable with Danish and Central North Sea well sections, one may also expect to find the "NP 9/10 boundary equivalent level" near to the LOD of *A. augustum* in the North Sea Basin, whereas the "d13C peak level equivalent" approx. coincides there with the FOD of that species.

In Denmark (a.o. the Bovlstrup well section, cf. Laursen & Andersen, 1997), this *A. aug.* FOD level coincides with the onset of the Olst Formation (=Sele Formation equivalent) volcanics, and with a strong impoverishment of the (agglutinated) foraminifera assemblages, leaving only a couple of taxa (*Evolutinella and Verneuilinoides* spp...) above that level. Also in North Sea wells this level coincides with a rather abrupt impoverishment of the foraminifera (or the transition to entirely barren samples). In the North Sea Basin (Fig.2) it appears from common pyroclasts , that this "d13C peak level equiv." approximately coincides with a (re-) onset of volcanic activity (Knox, 1994; 1996; i.p.) as occurring in the Northern Atlantic, or nearer by, in Scotland. (A considerably older volcanic event, the Andrew tuff, observed in the lower part of the Lista Fm, is not discussed here). But from the base of the Sele Fm on volcanic activity ranges more or less continuously up to the top part of the Balder Fm ("major volcanic event"), and slightly bejond. This "vol. event" also occurs in the basal Ieper Clay of Belgium (Geets, 1994). Also in the Belgian well- and outcrop sections, it appears that benthic foraminiferal assemblages show a rather strong turnover (affecting some 25 taxa) from Late Paleocene (Moorkens, 1982), to Early Eocene (Willems, 1980), but a rather broad barren/ decalcified interval (and/or hiatuses) prevent giving its exact position. However, the dinocyst assemblages in cored sec-tions are generally well preserved, and a.o. the onset level of common *Apectodinium spp.* appears conspicuous (De Coninck, i. p., and his results to be presented at this conference); it may provide a valid correlation with the FOD of *A. augustum* - level in North Sea sections. Analyses (a.o. on stable isotopes) should thus be concentrated on the *A. augustum* (FOD) level in the two areas, as this is the approx. stratigraphic position of the "d13C atmospheric global event", and (e.g. on pyroclasts) at the *A. aug.*LOD level, (near to the NP 9/10 boundary equivalent) within the "major volc. event" interval.

Strata Série 1., vol. 9 p. 101-105, 4 fig. 1998

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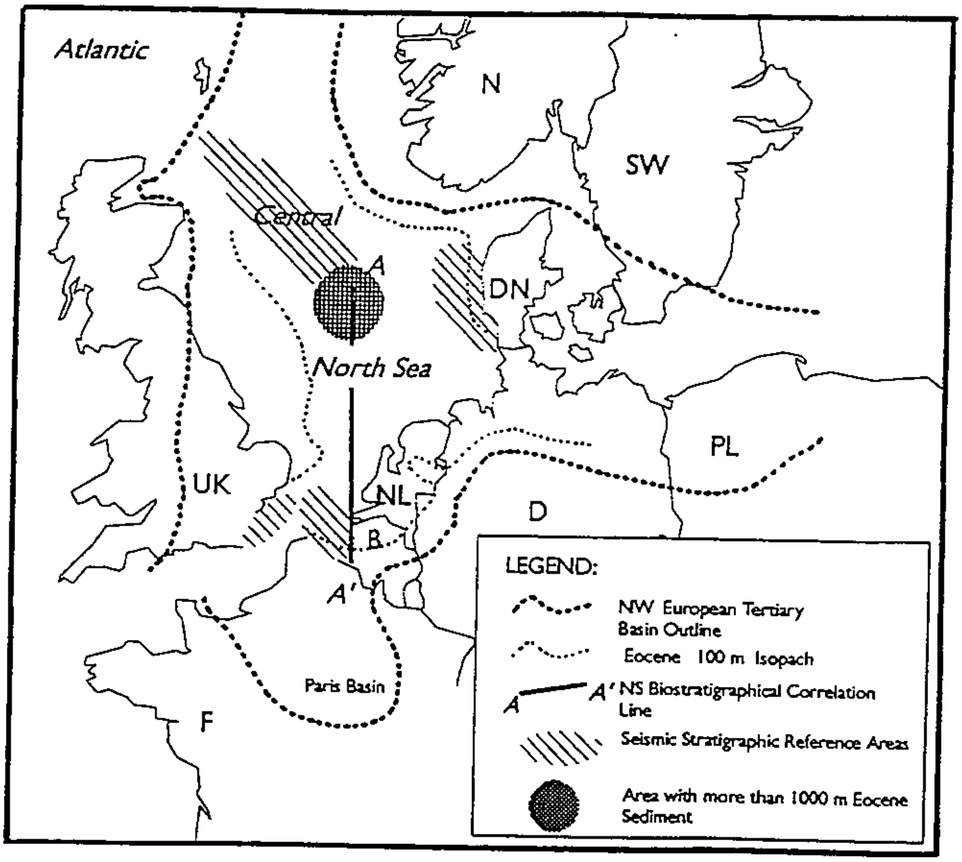

Fig.1. Locality details of the NW European / North Sea Basin.

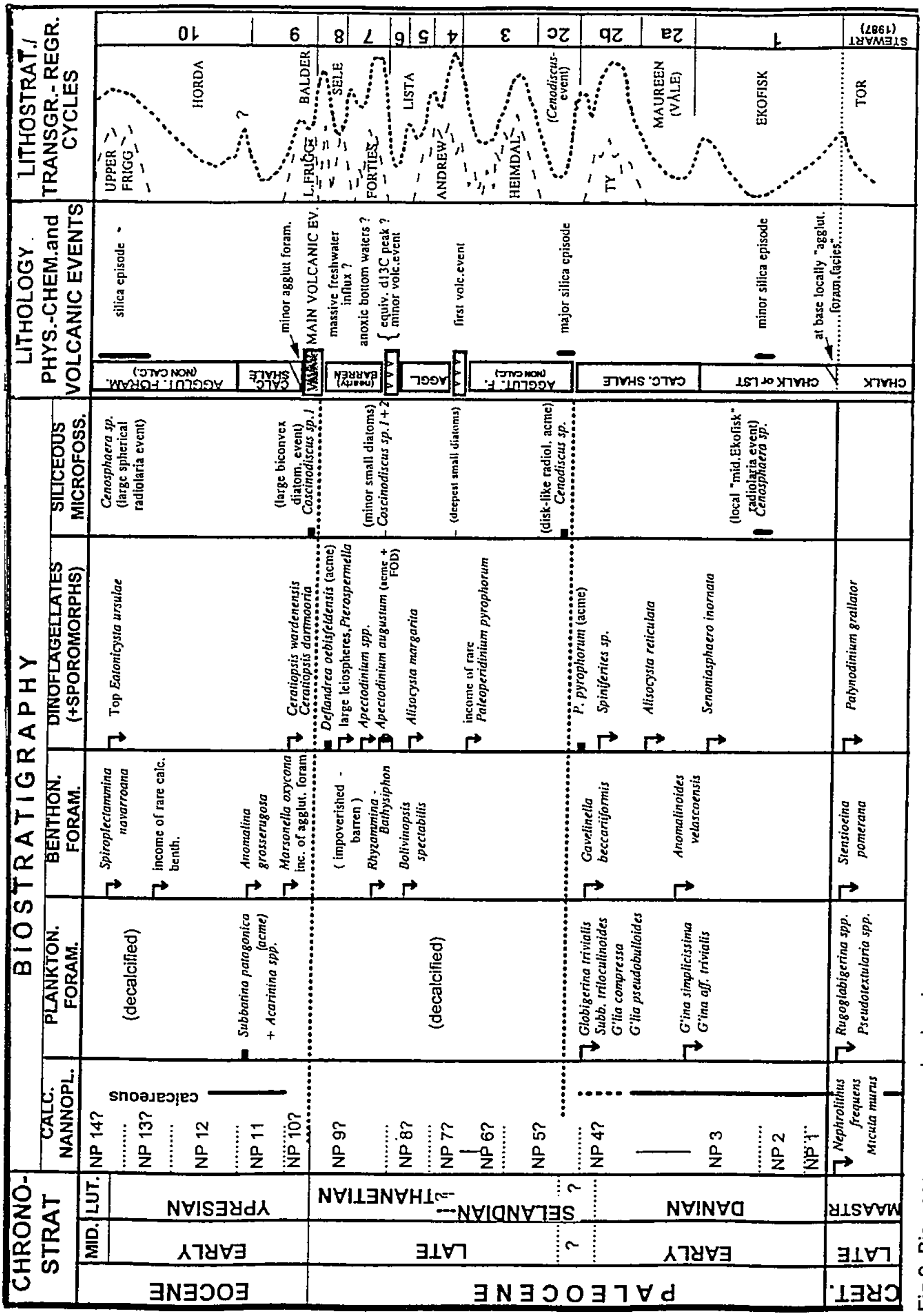

Fig.2. Bio-, sequence-, and volcanic event stratigraphy of the Central North Sea Basin.

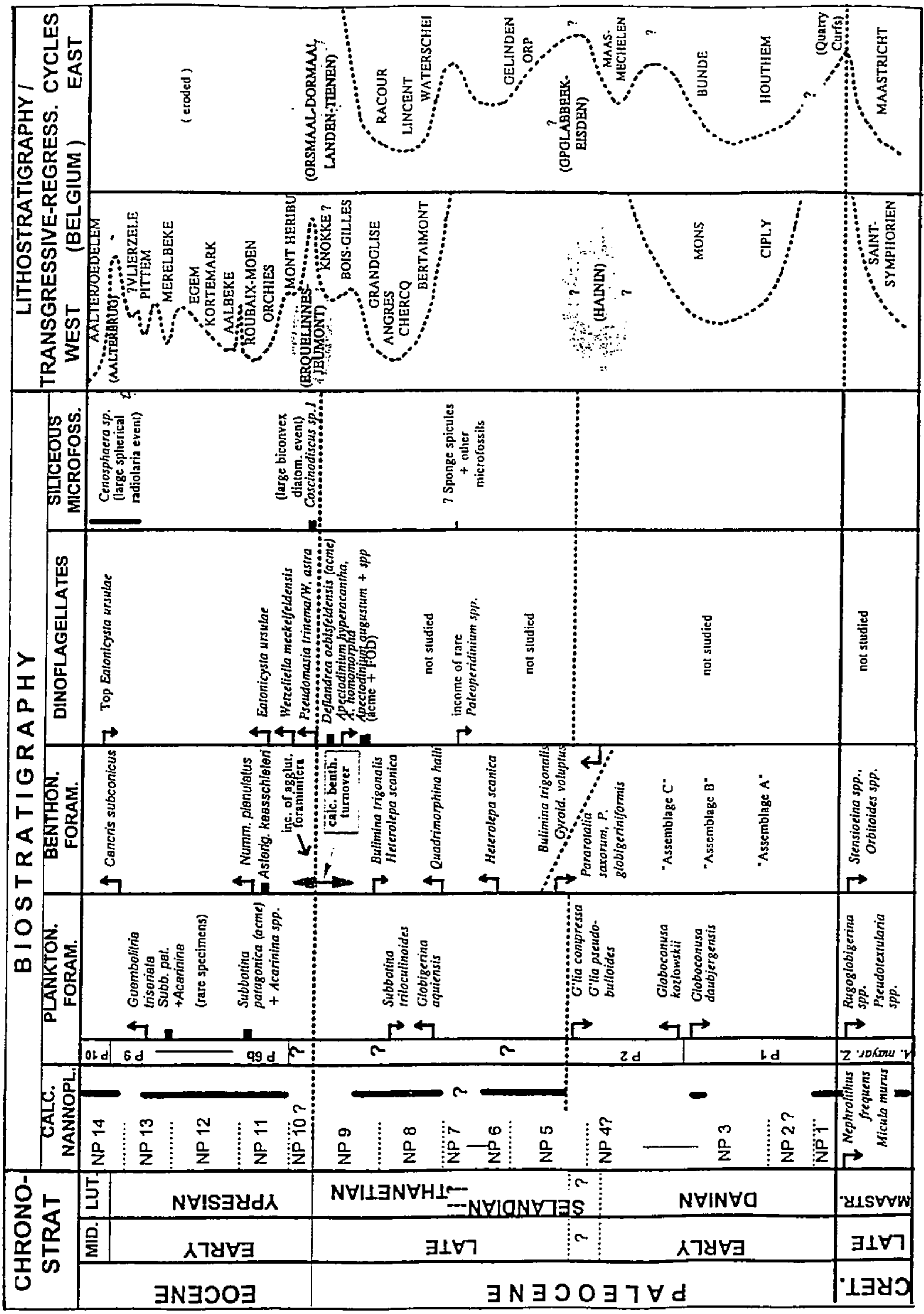

Fig.3. Bio-, and sequence stratigraphy of western and eastern Belgium.

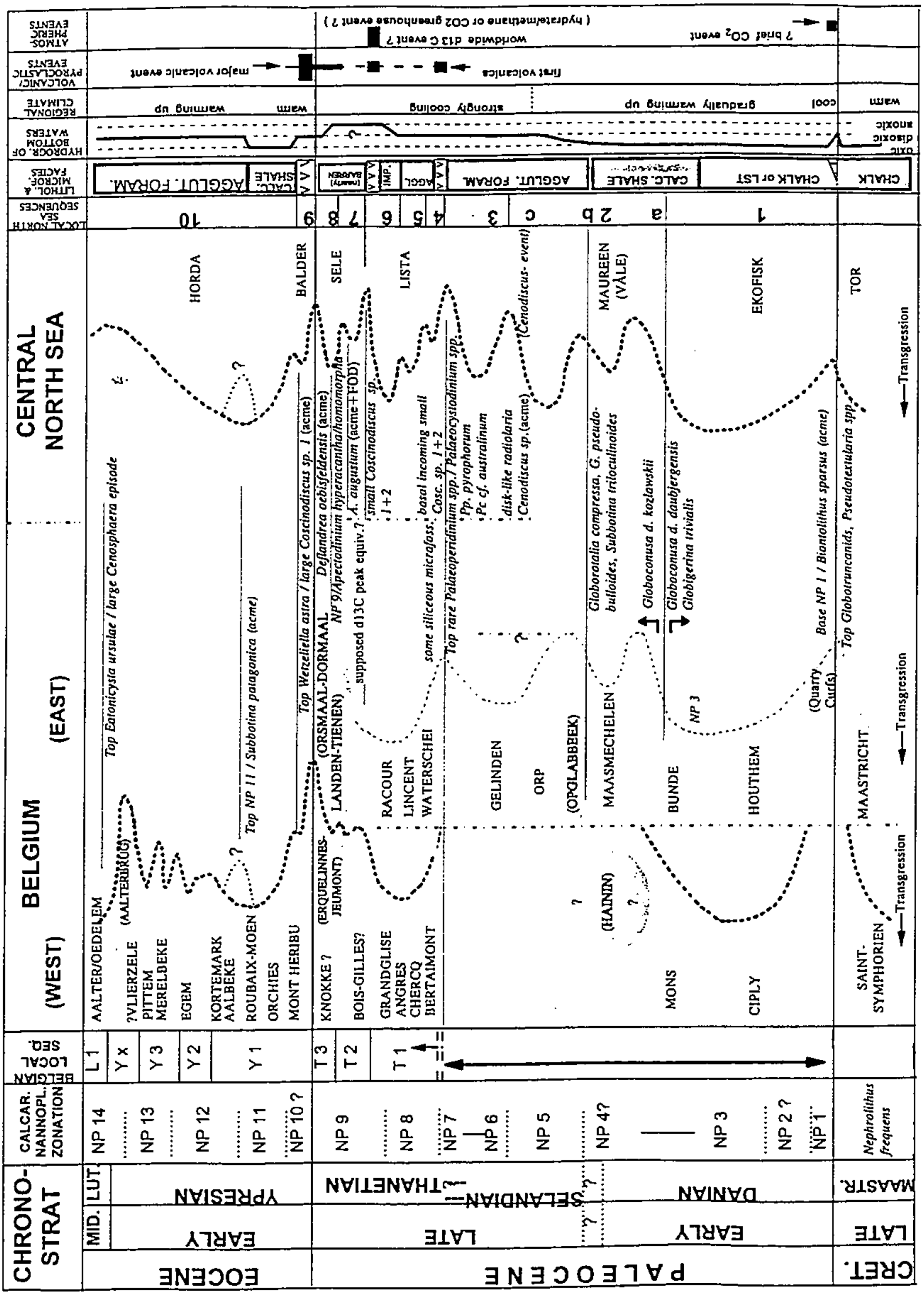

Fig.4. Correlation of the Paleocene-Early Eocene of Belgium and the Central North Sea.

NEW MICROPALEONTOLOGICAL DATA (PALYNOMORPHS AND PLANKTIC FORAMINIFERS) FROM SELECTED P/E BOUNDARY SECTIONS OF THE WESTERN PYRENEES

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The Paleocene/Eocene (P/E) boundary sections are mostly composed on the Western Pyrenees by carbonate sediments of a variety of facies, from bioclastic grainstones, rich in larger foraminifera characteristic of shallow settings, to alternations of hemipelagic marls and limestones in basinal zones (Baceta *et al.*, 1997). Very often, however, these carbonate-dominated successions contain a distinct fine-grained siliciclastic unit (SU), the origin of which is discussed by Pujalte *et al.* (this volume).

Previous micropaleontological analysis of basinal sections permitted the establishment of several biozonation schemes for these sections (Orue-Etxebarria *et al.*, 1996). Most of these biozone boundaries occur at different datum horizons and, therefore, there is no unique horizon for the P/E boundary, although it lies within or close to the SU. For instance, the base of the *Wetzeliella astra* dinoflagellate zone, which has been used to place the P/E biozone at other basins, occurs right at the top of the SU whereas the base of the *Morozovella subbotinae* planktic foraminifer biozone is present several metres higher. Other biozone boundaries, as well as isotopic peaks, occur within the SU. Since the biozonation schemes are already established and constitute a reference framework for further research, the purpose of the present study is to describe the palynological and plankton foraminifer biostratigraphy of a wider variety of settings, namely the outer platform (Urbasa), base-of-slope carbonate apron (Ermua and Aixola), and intermediate-deep basinal facies (Trabakua Pass and Zumaia) with emphasis on the SU.

The palynological biozonation published in Orue-Etxebarria *et al.* (1996) was based on marine palynomorphs but even if terrestrial forms were also present in the samples, they were not used in the establishment of any zonation scheme. The present study, although a continuation of the previous one, now addresses both types of palynomorphs and has further been extended to a section not previously analysed, namely the Urbasa section which represents a depositional setting not represented in the previous study. Samples from the five sections included in the present research, vary in palynomorph content but, in general, are moderate to rich, although there are some few samples that are either barren or very poor. Not all the sections have a similar yielding and, although variation is also present among the samples from one same section, the best samples come from the Ermua and Urbasa sections which are representative of two different depositional settings.

Marine palynomorph taxa include dinoflagellate cysts and acritarchs. The dinoflagellate cysts are consistently intermediate to well-preserved in all settings whereas acritarchs are variably preserved from poor to good in relation to both sample and facies. Terrestrial plynomorphs are represented by miospores, fungal spores, and pollen taxa. All these types are variably preserved, intermediate to well, at all sections but Urbasa where they are very well preserved. Spore and pollen taxa are more abundant and diverse at the outer platform setting of the Urbasa section whereas the marine palynomorphs are more diverse and abundant at the base-of-slope setting of Ermua and Aixola and at the basinal facies of Trabakua Pass and Zumaia. Marine dinoflagellates account for 85-90 % of the total assemblages, both in terms of species and specimens at all sections, except Urbasa where the opposite occurs.

All together, about 50 species of marine palynomorphs and 30 of terrestrial taxa were recognized in this study. Among the dinoflagellate forms proximochorate cysts are dominant and the most common species are those within the genus *Apectodinium*. Regarding acritarchs, *Michrystridium* is the most abundant genus among the four acritarch taxa identified in this study. However, acritarchs are not consistently present in this study. They appear very occasionally through the sections with the exception to the SU were they become fairly common coinciding with an increase in numbers of specimens and taxa of terrestrial palynomorphs as well as in reworked palynomorphs. These reworked taxa belong, among others, to genera such as *Oligosphaeridium, Diconodinium* and *Gonyaulacysta* but there are also reworked pollen and spore species.

Terrestrial palynomorphs are consistently present in all sections but show a peak abundance, both in terms of numbers of specimens and of taxa, at the SU. These palynomorphs are dominant at the outer platform setting of the Urbasa section where at the afore mentioned interval they account for 90-95 % of the total assemblages. Although *Caryapollenites* is the most abundant genus, *Kurtzipites*, *Tricolpites* and *Tiliapollenites* are also common. Fungal spore taxa such as *Dyadosporites*, *Monoporisporites* and *Pluricellaesporites* are also found in the studied samples but they appear very sporadically.

Regarding the planktic foraminifers, on the other hand, the richness of the samples and the good preservation of the specimens at the Zumaia section allow to undertake thorough quantitative and qualitative analyses. Although some species such as *Globanomalina pseudomenardii*, *G. chapmani*, *G. imitata* and *Subbotina pseudoeocaena* have their first occurrence or their last appearance in the nearby proximity of the SU, there is no indication of any kind of unusual extinction event for the planktic foraminifers in relation to this unit. However, a quantitative analysis allows to observe changes in the relative ratios of several genera, both below and above the mentioned unit. For instance, above the SU there is a concomitant decrease in the *Subbotina* and *Globanomalina* ratios and an in increase in those for the *Acarinina* assemblages.

Finally, sedimentological analyses (Pujalte et al., this volume) as well as micropaleontological studies based on palynomorphs and planktic foraminifers indicate an unusual origin for the SU. The increase in both absolute and relative ratios of terrestrial palynomorphs as well as in acritarchs and reworked forms points towards a sudden increase in sediments from surrounding emerged areas and concomitant stress conditions in the marine environment for a brief period of time that did not deeply influence the dinoflagellate and planktic foraminifer assemblages but that induced an increase in acritarchs. The stress conditions, therefore, did not generate any major extinction event for either the marine palynomorph taxa or the planktic foraminifers, but did produce quantitative changes in the assemblages and species turnover for the dinoflagellate cyst taxa. On the other hand, the terrestrial palynomorphs indicate a hot and wet climate period at the end of the Paleocene that allowed the establishment of a Paratropical Forest in southwestern Europe as has been reported also from other European regions.

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UN NOUVEAU GISEMENT PALEONTOLOGIQUE *KONSERVAT-LAGERSTATTE* A LA BASE DE L'EOCENE DANS LA REGION DE CREIL (OISE)

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Même si plusieurs gisements à ambre sont cités dans la partie française du Nord du Bassin Parisien (Lacroix, 1910; Feugueur, 1963; Du Gardin, 1990), aucun ambre fossilifère n'y avait été signalé. La découverte d'un gisement de ce type dans la vallée de l'Oise présente un grand intérêt. Ce site a livré une faune d'arthropodes dans l'ambre et, dans le sédiment encaissant, une faune de vertébrés continentaux et une flore bien diversifiées, de la base de l'Eocène.

La stratigraphie

Ce gisement, dit de Creil, est situé dans la vallée de l'Oise sous des alluvions détritiques quaternaires dans une gravière en exploitation. Les premières études sédimentologiques sur le terrain ont mis en évidence un ensemble de corps sédimentaires lenticulaires intriqués correspondant à des remplissages de chenaux ayant une orientation NE-SO. Les dépôts qu'ils contiennent sont fréquemment agencés en lits progradant vers le NE et évoquent une dynamique sédimentaire liée à un régime fluviatile qui drainait un domaine émergé d'âge Crétacé supérieur (remaniements de silex et galets de craie).

Le remplissage sédimentaire des chenaux présente deux faciès: des sables argileux très riches en lignites souvent pyritisés (avec des niveaux riches en concrétions pyriteuses) associés à de l'ambre; des sables gris plus ou moins argileux, pauvres en lignite recélant une faune de vertébrés continentaux.

Les chenalisations ravinent et cannibalisent le sommet des sables gris vert glauconieux (formation des sables de Bracheux) du Thanétien (Feugueur, 1963) qui affleurent en fond de vallée, latéralement au système chenalisant. D'origine marine, les sables glauconieux livrent une ichthyofaune variée dont 12 espèces de Squaliformes (et Teleostei ?). Les espèces de requins déterminées sont connues dans le Thanétien; deux d'entre elles sont signalées dans l'Yprésien. Trois dents de *Heterodontus* (sélacien) pouvant correspondre à une nouvelle espèce ont été récoltées. Dans son ensemble, cette ichthyofaune possède un cachet nettement Thanétien (Dutheil, 1991).

Les Mammifères

La liste faunique provisoire est la suivante: **Multituberculata**: gen. sp. indet.; **Marsupialia**: gen. sp. indet.; **Lipotyphla**: Dormaalidae, cf *Macrocranion*; **Plesiadapiformes**: 1) Plesiadapidae, un grand *Platychoerops* ou *Plesiadapis.* 2) Paromomyidae, cf *Arcius*; **Rodentia**: Ischyromyidae, 2 sp.; **Primates**: 1) Omomyidae, *Teilhardina* cf. *belgica.* 2) Adapidae, *Cantius* sp.; **Creodonta**: Hyaenodontidae (une petite espèce); **Carnivora**: Miacidae, cf. *Miacis*; **Condylarthra**: Arctocyonidae, *Landenodon* sp.; **Perissodactyla**: cf *Pachynolophus*; **Artiodactyla**: Diacodexeidae, *Diacodexis* sp.; **Pantodonta**: Coryphodontidae, *Coryphodon* sp.

La faune est diversifiée et peu tronquée sur le plan taphonomique, avec des formes de petite et de grande taille. Une quinzaine d'espèces sont reconnues à ce jour mais l'inventaire actuel est encore très incomplet (les gisements comparables tels Dormaal comportent de 30 à 45 espèces). Signalons la présence de nombreux poils et de deux plumes inclus dans l'ambre. Un des poils est associé à un poux (Mallophaga).

Cette faune présente un cachet éocène inférieur (Sparnacien *s.l.* : chrons 25N-24N d'après Cavelier & Pomerol, 1986, *ca.* - 56 à - 53 M.a. d'après l'échelle de Cande & Kent, 1992). D'une part, elle comporte des taxons tels que les Rongeurs, les Euprimates, les Périssodactyles et les Artiodactyles, faisant typiquement apparition à la base de l'Eocène dans les domaines laurasiatiques. D'autre part, la découverte du primate omomyidé *Teilhardina*, ainsi que du condylarthre arctocyonidé *Landenodon*, rappelle la faune belge de Dormaal qui caractérise le premier niveau-repère à mammifères de l'Eocène en Europe: MP7. Cette faune représente donc un niveau ancien de l'Eocène du Bassin de Paris (Godinot *et al.*, ce volume). Creil serait la faune éocène bien diversifiée la plus proche de la limite Paléocène-Eocène dans le Bassin Parisien. Sa position précise par rapport à Dormaal reste toutefois à établir.

Les autres vertébrés

Les reptiles sont représentés par des crocodiles (2 espèces au moins, dont un gavialidé), des tortues et des squamates (au moins trois familles de lézards, Gekkonidae, Lacertidae, Varanidae et trois familles de serpents, Boidae, Tropidophiidae, Russellophiidae, typiquement éocènes). La faune d'amphibiens comprend des urodèles (salamandroïdes) et peut-être des anoures. Les poissons sont largement dominés par le genre d'eau douce *Lepisosteus*. Les coprolithes sont attribuables à des gros vertébrés (plus abondants et en meilleur état dans les niveaux à lignite).

L'ambre

L'ambre est très abondant dans les bancs à lignite et argile (de 5 à 10 % du volume) où il se présente en masses importantes (max. 200 g.) ou, le plus souvent, sous la forme de plaques et de "coulures stalagmitiques" plus ou moins fines. La présence de perles d'ambre sphériques suggère que de la résine pouvait s'écouler directement dans l'eau (Nel, obs. pers.). La découverte de deux exuvies larvaires d'Ephemeroptera confirme que les essences végétales productrices d'ambre croissaient en bordure d'eau. La présence de "stalagtites" d'ambre extrèmement fines (1 mm de diamètre), parfois très longues (30 cm) et d'associations bois/résine prouvent que la résine n'a subi qu'un faible transport avant son dépôt et n'a pas été remaniée. Par contre, les couches à vertébrés continentaux recèlent, en faible proportion, des petits débris d'ambre plats et allongés, ce qui suggère un tri mécanique consécutif à un transport.

Les premières analyses par infrarouge de cet ambre le distinguent nettement de l'ambre balte, tant par la composition que par l'origine végétale. L'étude en lame mince des bois associés démontre que la résine était produite par un Angiosperme encore indéterminé, mais qui appartenait probablement aux Papilionacées, ce qui ferait de cet ambre le plus ancien connu de ce type.

La flore

La flore observée dans le gisement comporte: dans les sables: des pollens, graines, bois et troncs d'arbres flottés; dans l'ambre: des feuilles, tiges, brindilles et fleurs. Conservées en volume, les très nombreuses graines appartiennent à 5 ou 6 espèces. Certaines présentent des perforations liées à l'activité d'insectes. Les premières analyses des bois montrent la présence de Conifères et d'Angiospermes. Un cône de *Pinus* sp. conservé en volume a été observé. Six fleurs appartenant à 3 espèces ont été récoltées dans l'ambre. Les feuilles d'Angiosperme incluses dans l'ambre proviennent d'une seule espèce (arbre producteur ?).

L'entomofaune

L'entomofaune était très diversifiée et correspond à une époque pour laquelle aucun ambre insectifère n'avait été trouvé jusqu'ici. L'inventaire préliminaire comporte 17 ordres et plus de 200 espèces nouvelles: Lepidoptera, Trichoptera, Hymenoptera, Diptera, Coleoptera, Neuroptera, Mallophaga, Hemiptera, Psocoptera, Thysanoptera, Mantodea, Blattodea, Isoptera, Orthoptera, Dermaptera, Phasmatodea, Collembola. Les faits saillants de l'analyse préliminaire sont la découverte d'une nouvelle famille de Dermaptera (pince-oreilles), la présence d'Hymenoptera Scolebythidae (guèpes néotropicales, australiennes, malgaches et sud-africaines) et celle d'Isoptera Mastotermitidae (famille relicte de termites actuels en Australie, bien représentée dans les niveaux lacustres de l'Oligo-Miocène paléarctique et néarctique mais qui n'a jamais été signalée dans les ambres cénozoïques). Les Diptera et les Hymenoptera sont abondants, fait classique dans les gisements à ambres. Les Coleoptera sont très bien représentés, avec plus de 10 % des effectifs, ce qui est exceptionnel pour un ambre insectifère. L'abondance relative des Trichoptera et des Ephemeroptera (entre 3 et 4 %) tend à prouver l'existence d'un cours d'eau à proximité immédiate de la forêt productrice.

Conclusion

L'ensemble des données géologiques, faunistiques et floristiques recueillies à ce jour permet d'ébaucher une reconstitution du paléoenvironnement. Les bois, l'ambre et les restes de mammifères n'ont subi qu'un faible transport et se sont déposés rapidement, dans un milieu fluviatile, sous climat intertropical chaud et humide. La végétation luxuriante et variée permettait le développement d'une entomofaune abondante. Les vertébrés aquatiques, en particulier les tortues et les crocodiles, devaient pulluler dans cet environnement favorable. Les mammifères étaient également variés.

La fossilisation d'une telle association est d'un grand intérêt *paléoécologique*, et sans équivalent connu dans l'Eocène inférieur européen. Il est exceptionnel de trouver associées dans un même gisement flore, faune d'insectes et faune de vertébrés de la même chaîne trophique. Il s'agit de l'une des communautés fossiles continentales les mieux conservées et les plus complètes connues pour l'Eocène inférieur européen.

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AN EARLY "ILERDIAN" TRANSIENT SWITCH-OFF IN SHALLOW AND DEEP-WATER CARBONATE DEPOSITION IN THE WESTERN PYRENEES, SPAIN: ORIGIN AND RELEVANCE FOR THE PALEOCENE/EOCENE BOUNDARY DEBATE

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A major (punctuated) transgression took place in the Western Pyrenees throughout the Paleocene and early Eocene, causing the encroachment of wide carbonate platforms. These platforms were able to keep-up with the sea-level rise during Danian time, but gave-up and back-stepped as the transgression proceeded. The overall transgression peaked during the latest Paleocene-earliest Eocene interval, when the bulk of the so-called "Alveolina limestone" was deposited. On that interval, which roughly coincides with the "ILerdian" stage, seven different depositional settings can be recognized in a N-S cross-section (Fig. 1).

In six of these settings (the exception being the deep-sea channel axis), the "ILerdian" stage is predominantly represented by limestones or limestone/marl alternations, of either shallow-water origin, planktic derivation or a mixture of both sources. However, the study of 22 sections covering all these settings (Fig. 1, inset) also reveals the occurrence of a comparatively thin but laterally extensive carbonate-poor unit. That unit is a clear proof that, for a short while, carbonate-dominated sedimentation was replaced by siliciclastic deposition (mostly fine-grained) over most of the western Pyrenees.

Biostratigraphic analyses further demonstrate that the accumulation of this Siliciclastic Unit (SU): (i) took place within the time span of the SB-5 largest foraminifera biozone (sensu Serra-Kiel *et al.*, in press), within the middle-upper part of the P-5 planktic foraminifera biozone (sensu Berggren *et al.*, 1995), and within the dinoflagellate cyst *Apectodinium hyrecanthum* zone, and, (ii) the calcareous nannoplankton and deep-water benthic foraminifera zonal boundaries NP-9/NP-10 and BB-1/BB-2 are situated at or near the base of the SU. On the other hand, isotopic analyses also show that the base and top of the SU are respectively placed just above the onset of the prominent δ13C negative excursion, and below its conclusion (Schmitz *et al.*, 1997, and this volume). This later evidence indicates that the SU was laid down in a comparatively short time span (up to 400 kyr. but possibly less, since the isotopic excursion is thought to have lasted just about 100.000 years).

It seems likely, therefore, that the changes, elsewhere reported, in oceanic circulation connected with the P-E boundary induced a period of hot and wet climate also in the Pyrenees. Chemical weathering would then be enhanced, as would stream runoff and the proportion of suspension load in rivers. These, and a concurrent fall of sea level, greatly increased the flux of fine-grained siliciclastic sediments to the sea, a possibility further reinforced by the large proportion of terrestrial palynomorph within the SU (see Núñez-Betelu *et al.*, this volume). As an outcome, previous carbonate sedimentation was almost entirely suppressed, or greatly diluted. In addition to this, some carbonate dissolution seems to have taken place in basinal settings, evidenced for instance by partial dissolution of microfossil tests.

Whatever its genesis, in the Western Pyrenees the SU constitutes an excellent marker horizon to subdivide and correlate shallow and deep-water "ILerdian" carbonate successions. Therefore, if the deep benthic foraminifer extinction event and concurrent isotopic excursion are chosen to pinpoint the P/E boundary, Paleocene and Eocene deposits would be nicely separated in the Pyrenees.

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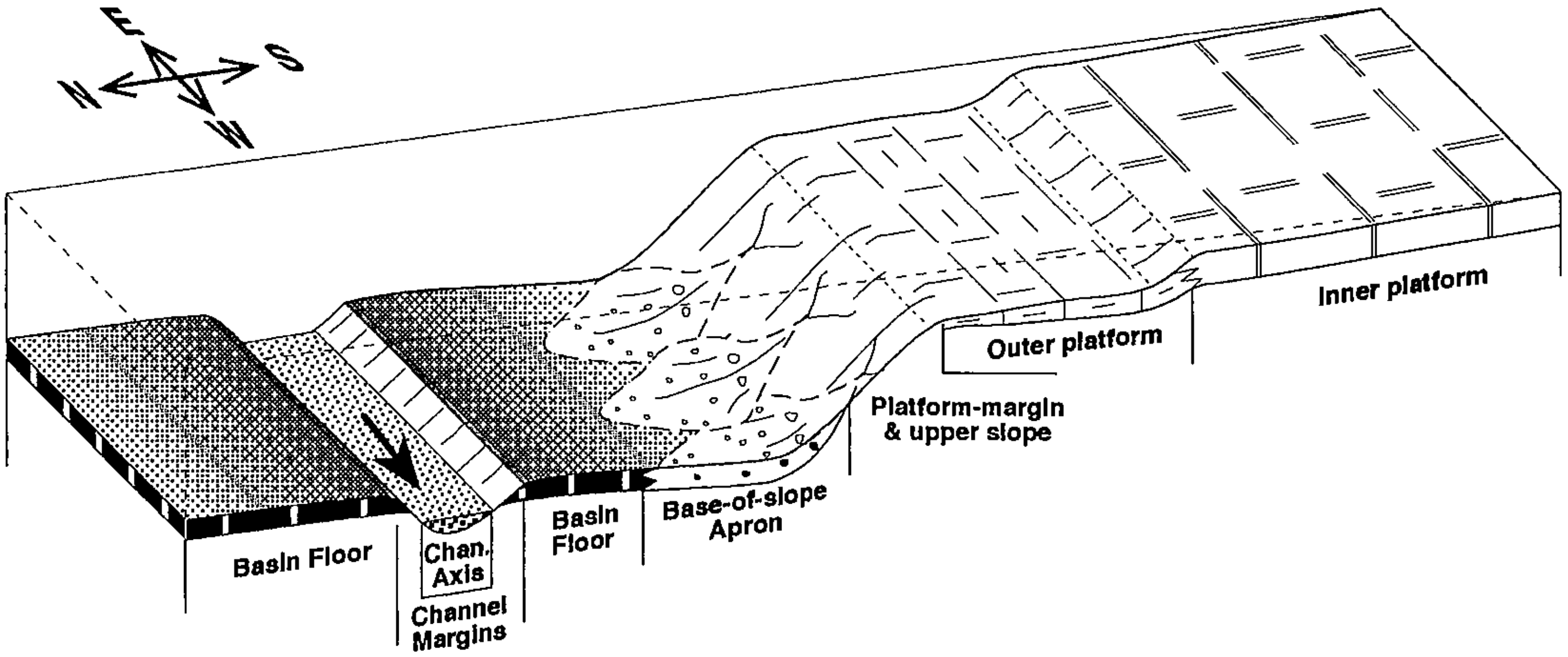

Figure 1.- Reconstructed cross-section showing the different late Paleocene depositional settings recognized on the Western Pyrenees.
A carbonate-dominated sedimentation, of either shallow-water, pelagic or mixed character, took place in most of them (i.e., except in the deep-sea channel axis).
The situation changed abruptly inmediately after the Benthic Extintion Event and, for a short while, a fine-grained siliciclastic deposition was the rule.
The talk discusses the character and relevance of that siliciciclastic interval.

SETTINGS	STUDIED SECTIONS
Inner platform	Campo (1), Laminoria (2), Leortza (3) Corres, (4), Villalain (5)
Outer platform	Ordesa (6), S. Adrian (7), Urbasa (8)
Ptm. margin & u. slope	Lizarraga Pass (9), Unanua (10)
Base of slope apron	Irati (11), Beorburu (12), Ermua (13)
Basin floor	Zumaia (14), Berano (15), Trabakua Pass (16)
Channel margin	Ibaeta (17), Amilategi (18), Urrutxua (19)
Channel axis	Orio (20), Gonzugarai (21), Unbe (22)

CALCAREOUS NANNOFOSSILS, FORAMINIFERA AND STABLE ISOTOPE STUDIES FROM P/E BOUNDARY SECTIONS IN EGYPT

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The terminal meeting of IGCP 308 on the Paleocene/Eocene (P/E) Boundary was held 1997 at the Penrose Conference in Albuquerque. But the search for a GSSP, a Global Stratotype Section and Point, for the P/E Boundary is not concluded. One of the reasons for the difficulty in deciding both on a criterium for the definition of this boundary and for an actual location of the GSSP is the fact "that many of the classical northern European sections are terrestrial to shallow marine and contain few biostratigraphically useful fossils" (GSA Today, 8.1996). We are, therefore, grateful that we can present results of our studies from Egypt at a meeting concentrating on the European record.

Three possible candidates for a GSSP, the sections at

- **Gabal Owaina** in the southern Nile Valley,
- **Gabal Duwi** at the Red Sea coast and
- **Gabal Abu Had** (Gabal Qreiya of some authors) in the East of Qena in Middle Egypt,

are presently under investigation as part of the PhD thesis of A.A.T. including the study of the lithology, the calcareous nannofossil and planktic foraminiferal biostratigraphy and the stable isotope stratigraphy of the Maastrichtian through Lower Eocene.

Schmitz et al. (1996) have shown the presence of well developed P/E boundary sequences at Gabal Duwi and Gabal Owaina by describing both the BFEE (benthic foraminifera extinction event) and the negative $\partial^{13}C$ shift measured on benthic foraminifera and on bulk samples. In the present study six sections were studied, the three completest of which will be presented.

At all these three sections the **calcareous nannofossil** and **planktic foraminifera** assemblages include the important marker species for the P/E Boundary interval. The BFEE and the negative $\partial^{13}C$ shift were found to lie within the *Morozovella velascoensis* Zone. The first occurrence of *Globoanomalina luxorensis* occurs at or just above the BFEE at all six sections studied and marks the base of the new *G. luxorensis* Zone. Contrary to the observations by Spejer & Samir (1997), *G. luxorensis* was not encountered below the BFEE.

At **Gabal Owaina** — Fig. 1, with a sample spacing of ca. 100cm — *Rhomboaster cuspis, R. spineus, R. cf. R. bramlettei, Rhabdolithus solus* and *Discoaster mahmoudii* and *D. diastypus* all appear in the same sample just above the BFEE, while the genus *Fasciculithus* is still present but has decreased in abundance and diversity. The negative $\partial^{13}C$ shift of about 3%o measured on bulk samples starts just above the Tarawan (chalk) Formation, some 5m below the BFEE.
This placement of the P/E Boundary interval is considerably lower in the Esna (shale) Formation than previously described. Schmitz et al. (1996) show the NP9/NP10 Boundary ca. 8m above the negative $\partial^{13}C$ shift, while we found it ca. 3m above it. Both the BFEE and the NP9/NP10 boundary occur within the negative $\partial^{13}C$ shift.

Also at **Gabal Duwi** — Fig. 2, the sample spacing is ca. 50cm, *R. cuspis, R. spineus, Tribrachiatus bramlettei* and *Discoaster mahmoudii* appear together in the sample above the BFEE, while *D. diastypus* appears and becomes common in the next sample, the highest to include common *Fasciculithus* and followed by a barren sample. *F. bobii* and *F. alanii* became extinct before the BFEE. The BFEE occurs within the negative $\partial^{13}C$ shift of about 3 (-5?)%o . Our NP9/NP10 boundary — based

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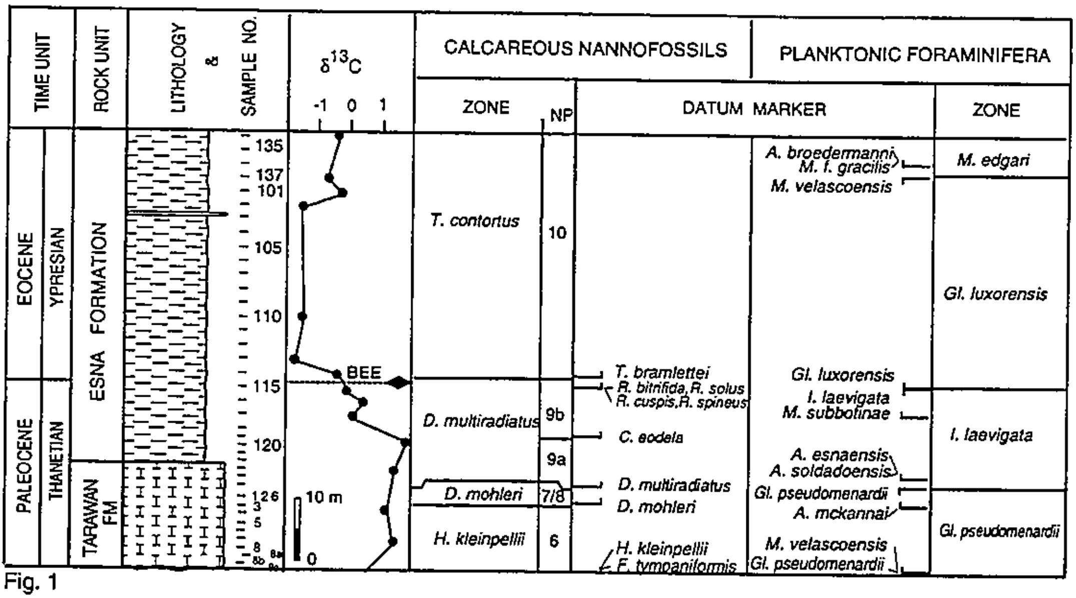

Fig. 1

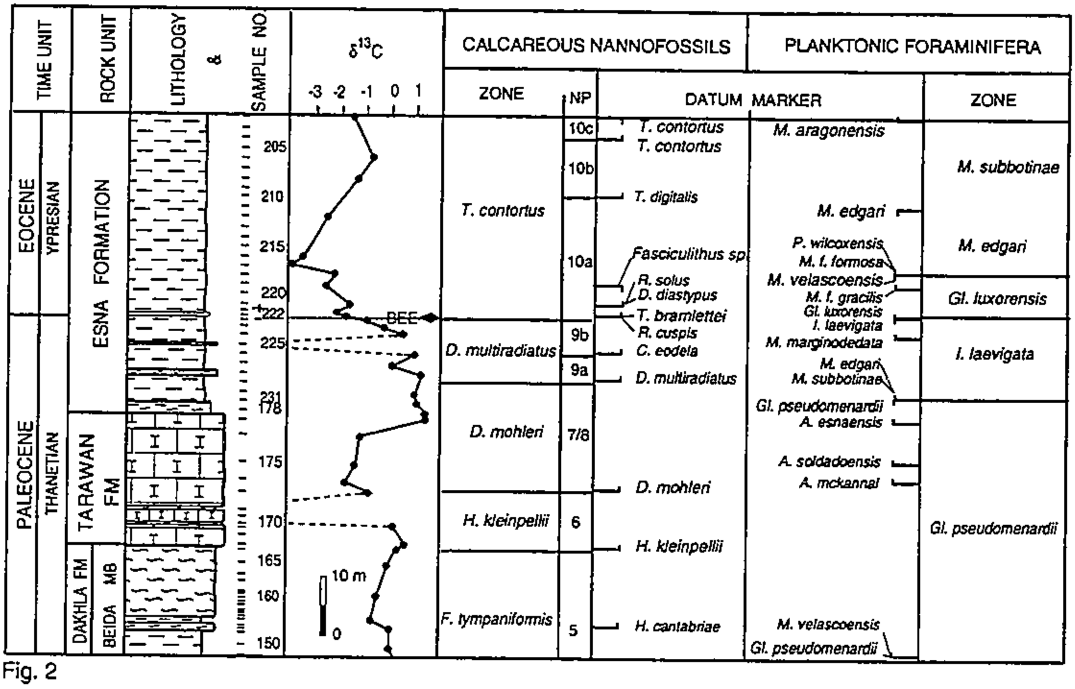

Fig. 2

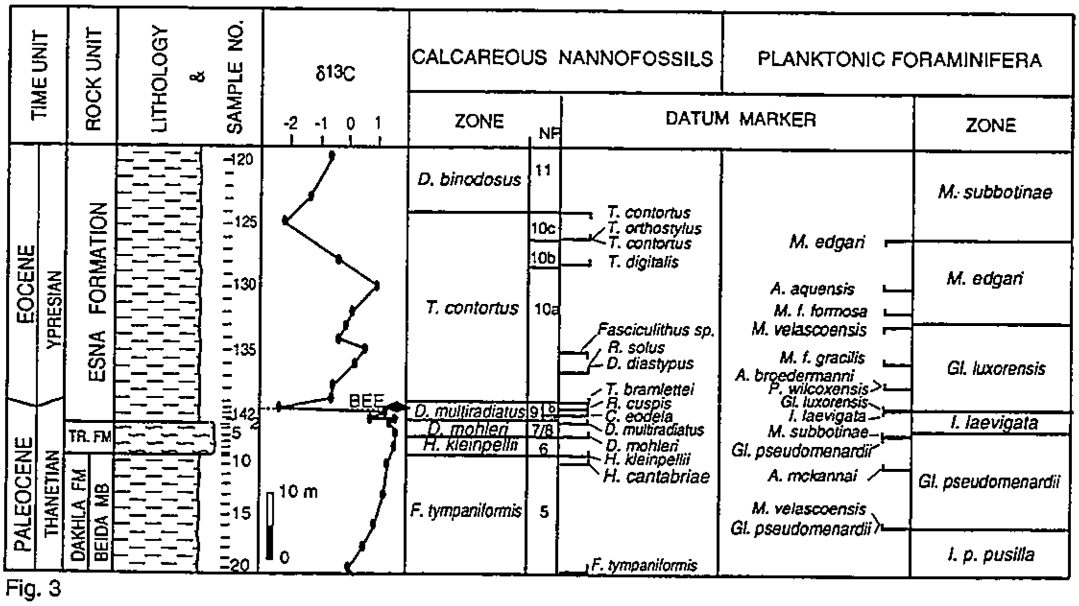

Fig. 3

on the FO of *T. bramlettei* — occcurs just above the BFEE, while Schmitz et al. (1996) show it about 1m above it. The *Frondicularia*-bed referred to by Speijr et al. (1996) lies about 200cm below the BFEE.

At **Abu Had**, Fig. 3, at a sample spacing of 50 to 100cm, *R. cuspis, R. spineus* and *D. mahmoudii* appear together just above the BFEE and towards the end of the negative ∂^{13}C shift of about 2.5%o . Above this we found the bas NP9/NP10 boundary with the first occurrence of *T. bramlettei* and *Neochiastozygus junctus* followed by the FO of *R. solus* and *D. diastypus*. *Fasciculithus tonii* and *F. alanii* became extinct before the BFEE.

The presence of few to common *Hornibrookina*, a calcareous nannofossil genus usually restricted to high latitudes and found both at Gabal Owaina, Gabal Abu Had and in a single sample at Gabal Duwi is surprising but not unique, since it was also found in Tunisia P/E Bounadry sections. There it seems to be connected to the early (but not the initial) phases of transgressions, while it's meaning in Egypt has not yet been analysed.

T. Bralower and KvS recently compared notes on their findings that, across the P/E Boundary interval, the representatives of the calcareous nannofossil family of the Prinsiaceae decreased considerably in favor of *Coccolithus pelagicus*. Such observations were made both in an ODP-Site in the Pacific, in the northern Tethys, in Tunisia and now in Egypt (Fig.4). This "*C. pelagicus* acme" can be used to recognise the P/E Boundary interval in sections, where markers are very rare or absent.

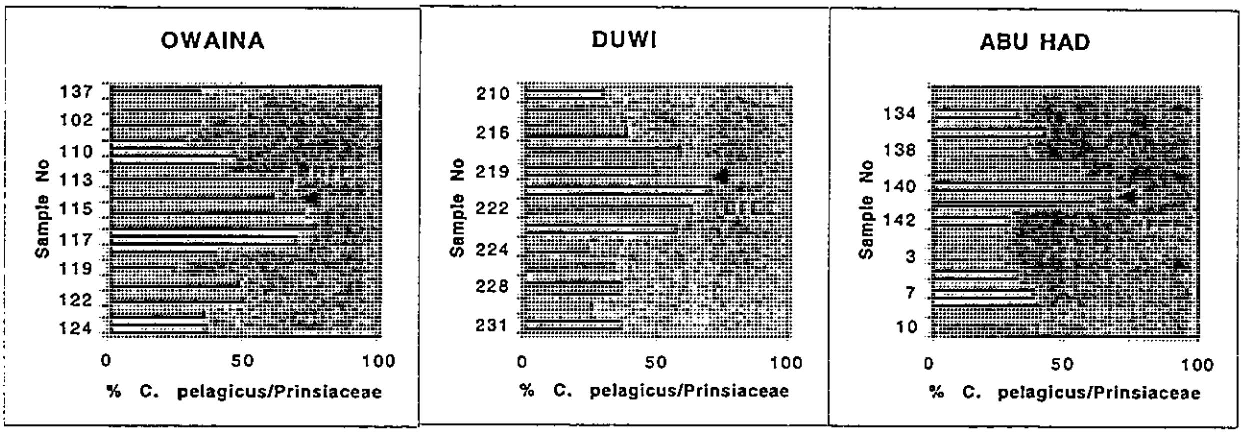

Fig. 4 *Coccolithus pelagicus* vs. Prinsiaceae at Gabal Oweina, Gabal Duwi and Gabal Abu Had across the P/E Boundary interval. Only in this interval does *C. pealgicus* reach over 50%.

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HIGH RESOLUTION CARBON ISOTOPIC CORRELATION OF THE LATEST PALEOCENE BENTHIC EXTINCTION EVENT : THE ERMUA AND ZUMAYA SECTIONS, NORTHERN SPAIN

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Very expanded and continuous marine sections across the latest Paleocene benthic extinction event occur at Zumaya and Ermua, northern Spain. The two sections reflect basinal and base-of-slope successions, respectively. Stable isotopic work is difficult in these sections because of common diagenetic overprinting of original isotopic signals. Calcareous foraminifera tests are in general badly preserved or infilled by extraneous calcite. However, a previous isotopic study of the Zumaya section revealed that whole-rock calcite can be used for carbon isotopic reconstructions. The whole-rock $\delta^{13}C$ record for the Paleocene at Zumaya showed the same major features as records measured on well-preserved deep-sea foraminiferal calcite. Limestones and marls that have been indurated or compacted during early diagenesis may represent closed systems with respect to carbon isotopes, and therefore have a great potential for retaining original $\delta^{13}C$ signals. This is particularly true for sediments where the calcite/organic matter ratio is high, as is often the case at Zumaya and Ermua. Here we present the results of a stable isotopic study of the Ermua section and compare the results with those for Zumaya.

At Ermua the isotopic signatures have generally been more altered than at Zumaya. Many samples give diagenetic signatures. Turbidite beds and some marl layers show $\delta^{18}O$ values as low as -5 ‰. Carbon isotopic values are sometimes more negative than at Zumaya and correlate with oxygen isotopic values, a strong indication of diagenesis. However, many marl intervals and all indurated limestones give isotopic values in agreement with those at Zumaya. We established a whole-rock carbon isotopic curve for Ermua based on measurements of limestone samples and those marl samples that have $\delta^{18}O$ values similar to the limestones. By this "filtered" carbon isotopic curve we can correlate the Ermua and Zumaya sections and show that they have the same general lithological successions in the latest Paleocene, however, the Ermua section is much more expanded and contains abundant turbidites which are very rare at Zumaya. At Zumaya, marls of late Paleocene (Zone P5) age are interupted by a 0.7 m thick limestone bed. This bed is overlain by a 0.4 m thick greenish marl interval which grades into a 4 m thick clay interval.The highest occurrence of many late Paleocene deep-sea benthic foraminifera coincides with the transition from greenish marl to calcite-free clay. A gradual 1.6‰ negative shift in $\delta^{13}C$ is developed over the greenish marl interval, probably reflecting the global negative $\delta^{13}C$ shift in connection with the benthic extinction event. At Ermua, the late Paleocene (Zone P5) section is dominated by marls and turbidites (and one big slump bed), that is interupted by a 2 m thick interval of limestone and turbidites. Above this follows 0.9 m of marls and turbidites in which $\delta^{13}C$ declines.Thereafter follows circa 25 m of clay corresponding to the 4 m of clay associated with the benthic extinction event at Zumaya.

LES FAUNES A VERTEBRES DE DORMAAL (BELGIQUE) : RENOUVELLEMENT, CORRELATIONS ET PALEOBIOGEOGRAPHIE

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Le gisement de Dormaal (Formation de Tienen, Brabant, Belgique) a fourni d'importantes faunes de vertébrés dont des mammifères permettant de définir la faune-repère MP7 de l'échelle biochronologique mammalienne du Paléogène européen (Schmidt-Kittler, éd., 1987). La faune-repère MP7 se caractérise par les premières occurrences de groupes modernes en Europe (euprimates, artiodactyles, périssodactyles, rongeurs) alors que persistent encore des groupes archaïques (condylarthres, arctocyonides, apatothères,...). Bien qu'une dizaine de fouilles aient déjà été menées à Dormaal depuis sa découverte en 1883, l'âge du dépôt et sa biostratigraphie restent encore à préciser. Les dernières fouilles (R. Smith, 1989-1990) permettent non seulement de présenter des faunes plus diversifiées qu'auparavant, mais également d'apporter des données biostratigraphiques nouvelles concernant la limite Paléocène/Eocène en Belgique.

Rappelons qu'il convient de distinguer au sein du dépôt continental trois faunes d'origines différentes: une faune terrestre composé de mammifères et de reptiles terrestres; une faune dulcicole de poissons, amphibiens et reptiles d'eau douce; enfin une faune marine remaniée, riche en élasmobranches.

Près de 40 taxons de mammifères sont identifiés. Ceux-ci peuvent être corrélés avec les faunes de la limite Clarkforkien/Wasatchien (Cf3 à Wa1) des "Bighorn and Clarks Fork Basins" du Wyoming et Montana (U.S.A.). C'est la faune la plus ancienne du Wasatchien, celle du Wa0 (où se situe l'excursion négative du d13C) qui présente le plus d'affinités avec celle de Dormaal. La faune de mammifères de Dormaal est plus proche des faunes nord-américaines que de celles du sud de l'Europe. Ceci corrobore l'hypothèse quant à l'existence de 2 domaines européens isolés (?) géographiquement et/ou climatiquement à cette époque (Marandat, 1997) et permet de reconsidérer les voies de migration est-ouest pour les différents groupes de mammifères.

Une seconde faune abondante est celle des élasmobranches représentée par environ 4000 dents de requins et raies, pour un total de 45 taxons. Parmis ces derniers, 30 espèces se retrouvent dans le sommet des sables de Bracheux (Rollot, Ressons, Elincourt). Cette faune marine semble avoir été remaniée d'une assise juste inférieure au dépôt continental et qui aurait disparu à Dormaal. Cette assise serait donc équivalente au niveau supérieur des sables de Bracheux dans le Bassin parisien et également aux sables de Bois Gilles présents sous les sables continentaux d'Erquelinnes dans le Bassin de Mons. La présence significative de *Odontaspis winkleri, "Carcharias" vincenti, Physogaleus secundus, Rhinobatos matzensis* et *R. bruxelliensis* témoigne de l'apparition de formes "yprésiennes" dès le Thanétien supérieur et correspondrait à une ouverture progressive du Bassin de la mer du Nord laissant migrer des formes plus méridionales.

UPPER PALEOCENE TO LOWER EOCENE
LITHO -, BIO -, ECO -,
AND CHEMOSTRATIGRAPHY IN EGYPT AND ISRAEL

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Lower Paleogene marine deposits cover vast areas of Egypt and Israel, usually in a subhorizontal or slightly tilted position, but hardly affected folding. In combination with a near absence of vegetation, this setting provides unique opportunities to study Paleocene-Eocene stratigraphy along paleoenvironmental (e.g., paleobathymetric) gradients. Consequently, the region has strongly contributed to the understanding of early Paleogene stratigraphy, particularly low-latitude biostratigraphy, and continues to do so in relation to global change and events, such as during the latest Paleocene.

Traditionally, the Paleocene/Eocene boundary in Egypt has been delineated by the marl-limestone contact between the Esna Fm. and Thebes Fm. (or their lithostratigraphic equivalents), or by the calcareous nannofossil NP9/NP10 zonal boundary, or by the planktic foraminifera P5/P6a zonal boundary (terminology after Berggren et al., 1995), both situated within the Esna Fm. (e.g., Nakkady, 1950; Said, 1990). A similar situation exists in southern Israel (e.g., Arkin et al., 1972) although different lithostratigraphic terminology is being used there (Taqiye and Mor Fms instead of Esna and Thebes Fms, respectively).

Recent studies on stratigraphy and paleoenvironments in connection to latest Paleocene events were performed on the following sections:

- Ben Gurion (Nahal Avdat), Negev Desert (Benjamini, 1992; Speijer 1994a,b; Lu et al., 1995; Schmitz et al., 1997).
- Wadi Nukhl (Abu Rudeis), Gulf of Suez Coast, Sinai Desert (Speijer, 1994a,b; Speijer et al., 1997).
- Gebel Aweina, Nile Valley, Eastern Desert (Speijer, 1994a; Charisi and Schmitz, 1995; Speijer et al., 1995; Schmitz et al., 1996, 1997).
- Gebel Duwi, Red Sea Coast, Eastern Desert (Speijer, 1994a; Schmitz et al., 1996; Speijer et al., 1996).
- Gebel Qreiya, Wadi Qena, Eastern Desert (Speijer et al., in prep.).

These sections are arranged on the generally northwest trending paleoslope of the North African continental margin, which in its northern part (Negev and northern Sinai) was interrupted by (predominantly submarine) highs and lows of the Syrian Arc. Benthic foraminifera assemblages indicate upper bathyal deposition at Ben Gurion and Wadi Nukhl, outer neritic at Gebel Aweina and Qreiya and middle neritic at Gebel Duwi.

Lithology

Within the studied interval (P4-P6a or NP8-NP10), monotonous marls prevail in all sections, providing rich and mostly well preserved foraminifera (benthic and planktic) and nannofossil assemblages, enabling high-resolution biostratigraphic, chemostratigraphic and paleoenvironmental studies. Three sections appear conformable in the interval of main interest, where a 20-50-cm-thick dark-brown laminated bed (Nukhl and Qreiya), or gray laminated bed (Duwi) is intercalated within the marls (Speijer et al., 1996, 1997). The Aweina and Ben Gurion sections do not show such a laminated bed, but instead show a minor unconformity at this level (Schmitz et al., 1997). In all sections, a foraminifera-rich limey bed rests on top of the laminated bed or unconformity.

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Stable isotopes

The globally recognized negative carbon-isotope trend and the superimposed negative $d^{13}C$ spike (~ 3 ‰ shift) have been recorded in whole-rock samples from all sections within Zones P5 and NP9. In Nukhl, Qreiya and Duwi, the base of the spike coincides with the laminated beds. In the other two sections, it coincides with the unconformity, suggesting that the lower part of the shift is truncated there. The $d^{13}C$ excursion terminates above the limey bed. For Duwi and Aweina, fragments of well-preserved thick-shelled *Frondicularia* and *Lenticulina*, provided benthic stable isotopic values, leading to similar $d^{13}C$ profiles and reliable $d^{18}O$ data. Thin shells of planktic foraminifera and benthics like *Nuttallides truempyi* or *Gavelinella beccariiformis* are generally unsuitable for this purpose, because of recrystallization and the presence of infillings that cannot be separated from shell material (Schmitz et al., 1996).

Benthic foraminifera

The base of the carbon-isotopic excursion coincides with prominent changes in benthic foraminifera assemblages. The bathyal sections show a sudden turnover similar to the deep-sea benthic extinction event (BEE), with an overall replacement of a *G. beccariiformis* assemblage by a *N. truempyi* assemblage. Numerous characteristic deep-sea taxa became extinct at this level. The neritic sections also show a distinct, though less dramatic turnover. At outer neritic depths a *Bulimina callahani* assemblage replaced an *Angulogavelinella avnimelechi* assemblage, whereas at the middle neritic Gebel Duwi a *Frondicularia phosphatica* assemblage was replaced by a *Valvulineria scrobiculata* assemblage (Fig. 1). Within the dark laminated beds (Nukhl, Qreiya, Duwi), a quite unusual oligotaxic *Anomalinoides aegyptiacus* assemblage is intercalated between the long-ranging assemblages indicated above. It consists of large numbers of *A. aegyptiacus*, *Stainforthia* sp., and *Valvulineria* sp., indicating temporary oxygen deficiency at the bathyal and neritic sea floor (Speijer et al., 1996, 1997). A detailed cm-scale sampling reveals that this assemblage is not present at Aweina or Ben Gurion.

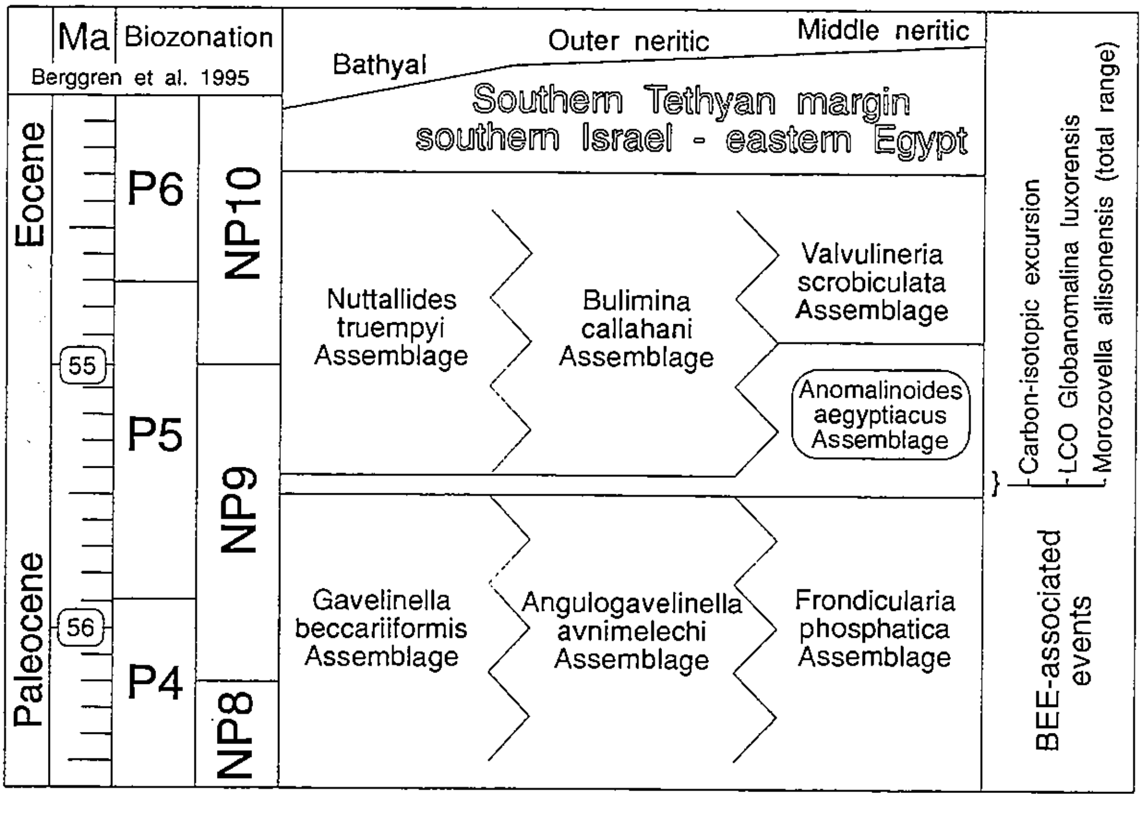

Figure 1. Spatial and stratigraphic distribution of benthic foraminiferal assemblages during the Paleocene-Eocene transition and other main events in isotopic and biotic evolution.

Planktic foraminifera

In all sections, the lowest common occurrence of *Globanomalina luxorensis* (Speijer and Samir, 1997) is found just above the base of the carbon spike, In addition, the laminated beds in the Nukhl, Qreiya and Duwi sections contain a unique planktic assemblage, largely consisting of *Acarinina* (e.g. *A. sibaiyaensis*) and with a minor but distinct component of *Morozovella allisonensis*. This unusual *Morozovella* species has until now only been found within the interval of the carbon isotopic excursion in the equatorial Pacific, ODP Hole 865 (Kelly et al., 1996), thus having an extremely short range (~ 100 k.y.). Neither this *Acarinina* dominated planktic assemblage, nor *M. allisonensis* were found in the Aweina and Ben Gurion sections. These biostratigraphic data enable a practical regional subdivision of the *Morozovella velascoensis* Zone (Zone P5) into a *G. chapmani* Zone (below the BEE) and a *G. luxorensis* Zone above the BEE. In the most complete sections, an additional and extremely narrow *M. allisonensis* Zone can be distinguished in between.

Summary

Our stratigraphic data from Egypt and Israel indicate the following:

- The level of latest Paleocene global change can be monitored lithologically, isotopically and faunistically in all sections studied, irrespective of their paleobathymetric positions. Presumably, other components, such as ostracoda, nannofossils, and clay-minerals show prominent changes in this interval too, but have not yet been documented in great detail.
- Although all sections show the (upper part of the) isotopic excursion, the ones containing the dark laminated interval with its unique biotic and paleoenvironmental features probably provide the best registration of the sequence of latest Paleocene events in this region.
- The extremely short stratigraphic and wide geographic (low latitude) range of *M. allisonensis* enables an accurate correlation between the equatorial Pacific and the southern Tethys, and assessment of stratigraphic completeness within low latitude sections. Absence of this marker species most likely indicates an incomplete record across the BEE.

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THE MID-PALEOCENE TO EARLY EOCENE DEPOSITIONAL HISTORY AND BIOTIC EVOLUTION IN THE SOUTHERN NORTH SEA

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High-resolution stratigraphical and micropalaeontological studies of outcrop and borehole sections throughout Belgium and northern France have documented a complex Mid-Paleocene to Early Eocene depositional history, marked by many disontinuities and changes in lithofacies. A series of 10 major stratigraphic events has been identified, delimiting several large scale depositional sequences. The temporal and spatial relationship of these sequences has been established using newly developed calcareous nannofossil and dinoflagellate cyst biozonations (Steurbaut et al., in press). Age estimations and hiatus durations have been acquired through calibration with the recently revised global polarity time-scale.

PALAEOECOLOGICAL EVOLUTION ACROSS THE PALEOCENE/EOCENE BOUNDARY IN BELGIUM

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The Kallo borehole (N Belgium, BGD hole 27E148) provides an excellent reference section through the Tienen Formation, recently introduced to replace the invalid Upper Landen Formation, the Upper Landenian or Sparnacian auct. (Steurbaut, in press). This formation, which is widely distributed in subcrop in northern Belgium, is believed to span the Paleocene/Eocene boundary. At Kallo, it consists of a ca 25 m thick, heterogenous lagoonal succession, unconformably resting on glauconitic sands of the Thanetian Grandglise Sand Member, and unconformably overlain by glauconitic fine sands, representing the base of the Kortrijk Clay Formation (formerly Ieper Formation).

Recent research on aspects of the micropalaeontology (including calcareous nannofossils, dinoflagellate cysts and miospores), palynofacies, clay mineralogy and geochemistry, based on a series of 40 close spaced samples, has greatly improved our knowledge of this previously poorly understood interval. This multidisciplinary approach has led to the recognition of 3 major transgressive-regressive cycles. These could be detected through analyses of specific parameters, such as, the amount of reworked clay minerals and organic matter (TOC), the percentage and biodiversity of aquatic palynomorphs, the degree of pyritisation and the ratio fresh versus oxidised continental organic matter. The fluctuations of the ratio miospores - fresh water palynomorphs seem to refer to substantial climatic changes, which occurred during the deposition of the Tienen Formation.

The palaeoenvironmental changes recorded across the Paleocene/Eocene boundary at Kallo are quite similar to these observed in coeval sections in Haute Normandie (France). Calibration of the biotic and abiotic events in both areas may lead to a better understanding of the temporal relationships between the changes in sedimentological signature as recorded in the southern North Sea Basin and the major events of global significance, which are useful for the definition of the Paleocene/Eocene boundary.

THE DORMAAL SANDS AND THE PALEOCENE/EOCENE BOUNDARY IN BELGIUM

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The Dormaal Sands, famous for their rich mammal fauna, are a series of fluviatile structureless gravel beds, cross-stratified lignitic sands and beige-coloured calcarenites, which are restricted to eastern Belgium. Their stratigraphic position, as well as their age, has been a focal point of controversy, essentially because of the complexity of their fauna, which includes a large variety of marine, freshwater and terrestrial taxa.

New excavations at the type locality of the Dormaal Sand Member, including detailed palaeontological and sedimentological investigations, have documented cyclic sedimentation within a fluvial system. The basal beds are rich in reworked material, including chalk pebbles, Cretaceous calcareous nannofossils and dinoflagellate cysts, and Late Paleocene shark and ray teeth, indicating substantial local and upstream erosion. The co-occurrence of indigenous biotic components (freshwater fish, amphibians and reptiles; freshwater algae) and surrounding contemporaneous groups (land mammals, tortoises, birds, land plants, etc.) refer to a relative wide river system, marked by rapidly changing sedimentation conditions, and bordered by dense warm-temperate forests, occasionally set afire by lightning.

Comparison with coeval sections of a more basinward position (e.g. Kallo) suggests that the stratotype of the Dormaal Sand Member corresponds to the lower part of the Tienen Formation (recently introduced by Steurbaut to replace the invalid Upper Landen Formation, Upper Landenian or Sparnacian auct.). It represents the most proximal or landward infill of a large valley, cut out during a major eustatically induced sea-level lowstand. This infill, which marks the onset of a subsequent relative sea-level rise, seems to be prior to event t4 (Steurbaut et al., in press). The palaeosol, associated with event t4 and documenting a continental episode, as well as the overlying major lignitic horizon, have now been recognised throughout the southern North Sea Basin. However, correlation between the stratigraphic events in this confined, marginal marine sedimentation area and the major events of global significance, potentially useful for the Paleocene/Eocene boundary definition, is still tentative, and awaits completion of ongoing mammalian and stable isotope studies.

TENTATIVE CORRELATIONS BETWEEN CONTINENTAL DEPOSITS OF THE ARGILES PLASTIQUES (PARIS BASIN) AND READING BEDS (LONDON BASIN), BASED ON CHEMOSTRATIGRAPHY

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The upper Paleocene-lower Eocene formations on the southern edge of the North Sea Basin consists mainly of continental clay-rich deposits referred to as "Argiles Plastiques" in the Paris Basin and Reading and Woolwich Beds in the London Basin. Stratigraphic correlations are extremely difficult and in fact poorly established. In addition to the absence of fossils in these continental deposits and the scarcity of stratigraphic markers in the correlative brackish deposits, geometric relationships are unclear due to sharp lateral facies variations and limited outcrops.

In the Paris Basin, 8 sections in quarries and drill holes through the Argiles Plastiques and extending over 350 km at the southern border of the basin have been selected for detailed isotopic study. In the London Basin, samples from the two Reading Beds units which alternate with the Woolwich Formation in the Jubilee 404 borehole have been analysed. Carbon isotope analyses were performed on the organic carbon occluded within the carbonate nodules as well as in the clay matrix. All sections were densely sampled with an average sample interval of 30 to 50 cm. In total 273 samples have been analysed for their carbon isotopic composition (Sinha, 1997).

The Argiles Plastiques $\delta^{13}C$ record

The Argiles Plastiques overly unconformably the Cretaceous Chalk and lower Paleocene marine limestones. Two distinct deposits have been differentiated: an older in the west, wich consists of smectitic mottled clays and nodular carbonates (the Argiles Plastiques bariolées) and a younger one in the East, comprising kaolinitic clays and sands (the Argiles Plastiques kaoliniques) (Thiry, 1981).

The Argiles Plastiques bariolées

Remarkably consistent variations of the $\delta^{13}C$ values of organic matter preserved in the Argiles Plastiques bariolées are observed between the six sections studied (Fig. 1). In five sections a marked decrease in $\delta^{13}C$ values of about 3 ‰ occurs near the base of the Argiles Plastiques bariolées. This negative peak is followed by three positive peaks. To facilitated our discussion the most significant events are numbered 1 to 4. The prominant negative (-27 ‰ to -28 ‰) peak at the base of the Argiles Plastiques bariolées and the most prominant positive (-21 ‰ to -23 ‰) peak at mid-section are refered to as P1 and P3 respectively. The intervening peak between P1 and P3 which average around -24 ‰ is referred to as P2. The positive peak above P3 which also average -24 ‰ is referred to as P4.

The four peaks can be recognised in three sections, those towards the center of the basin. Westwards, in the Limay section, the absence of the second peak (P2) is linked to the well developed mature nodular calcrete that formed over a relatively long time and thus corresponds to a period of non deposition. Similarly, in the Septeuil section the absence of the second peak P2 relates most probably to a stratigraphic gap indicated by a change in the clay mineral assemblage. Eastward, in the Meudon section, the lower part of the Conglomerat de

Meudon records positive values up to -22.5‰. In the upper part of the mammal bearing unit (Cendrier) values reach about -25.0‰. These most positive values probably correspond to peak P3.

The Argiles Plastiques kaoliniques

The values of -21‰ recorded at the base of Chalautre and the Sezanne boreholes are the same as the values of peak P3 in the Limay and Bougival sections (Fig. 1). Based on this we may interpret the record of the base of the Chalautre and Sézanne sections as being part of peak P3 and the clear peak above as corresponding to peak P4. If these correlations are correct, the Argiles Plastiques kaoliniques are equivalent to the upper part of the Argiles Plastiques bariolées.

Composite $\delta^{13}C$ variations of the "Sparnacian" in the Paris Basin

It seems that similar variations of the $\delta^{13}C$ record are exibited in the 8 Argiles Plastiques sections, and this, independently of facies. Characteristic $\delta^{13}C$ variations of the organic matter occur as well in the oxidized paleosol facies and in the fresh water euxinic deposits. Thus it may be possible to construct a composite $\delta^{13}C$ variation stratigraphy for the Sparnacian deposits. A few assuptions and principles underlie our construction.

First and foremost, we must admit that variations in the $\delta^{13}C$ composition of organic matter of 1.0 ‰ are meaningful, i.e., they do not correspond to "background variations".

Because of the lack of constraints which would permit us to determine sedimentation rates for different intervals of the sections, we cannot consider the rates of variation in isotopic composition between the sections; thus, we can only propose peak-to-peak correlations, aware of the danger of such a methodology. The most negative (P1) and the most positive (P3) peaks are characteristic enough to serve as markers and help to identify the secondary peaks (P2 and P4). Finally, we must consider that maxima and minima values may not have been recorded, either because of too broad sampling intervals, or because of the occurence of hiatuses. Thus, we must only consider the extreme values of the peaks.

A composite record can be put together. The individual peaks of the sections are put on the composite curve and a broad band representing a reliable interval including >80 % $\delta^{13}C$ records can be built (Fig. 2). The envelope casing the Argiles Plastiques bariolées is relatively narrow. For most of the events the width of the envelope does not exceed 1.0‰, corresponding to an interval of uncertainty > ± 0.5‰, whereas the variation from successive peaks is > 2.0‰, reaching some times 3.0‰. The $\delta^{13}C$ variations recorded in the Argiles Plastiques kaoliniques differ slightly from the Argiles Plastiques bariolées composite record. The Argiles Plastiques kaoliniques record appear parallel to the the Argiles Plastiques bariolées one , but shifted about 2‰ towards more positive values. This may be related to the different nature of the organic compounds preserved in the soils of the Argiles Plastiques bariolées and in the lacustrine deposits of the Argiles Plastiques kaoliniques.

The composite curve of $\delta^{13}C$ variations of the "Sparnacian" appears to have enough contrast to be use at least in the Paris Basin to make correlation between the different formations. Nevertheless, it has to be keep in mind that slight shifts of the composite curve may occur according to the deposit environment. The composite curve of $\delta^{13}C$ variations may be used as a local stratigraphic scale in the Paris Basin. Remains the question, how far this composite chart can be extended to over regions or basins.

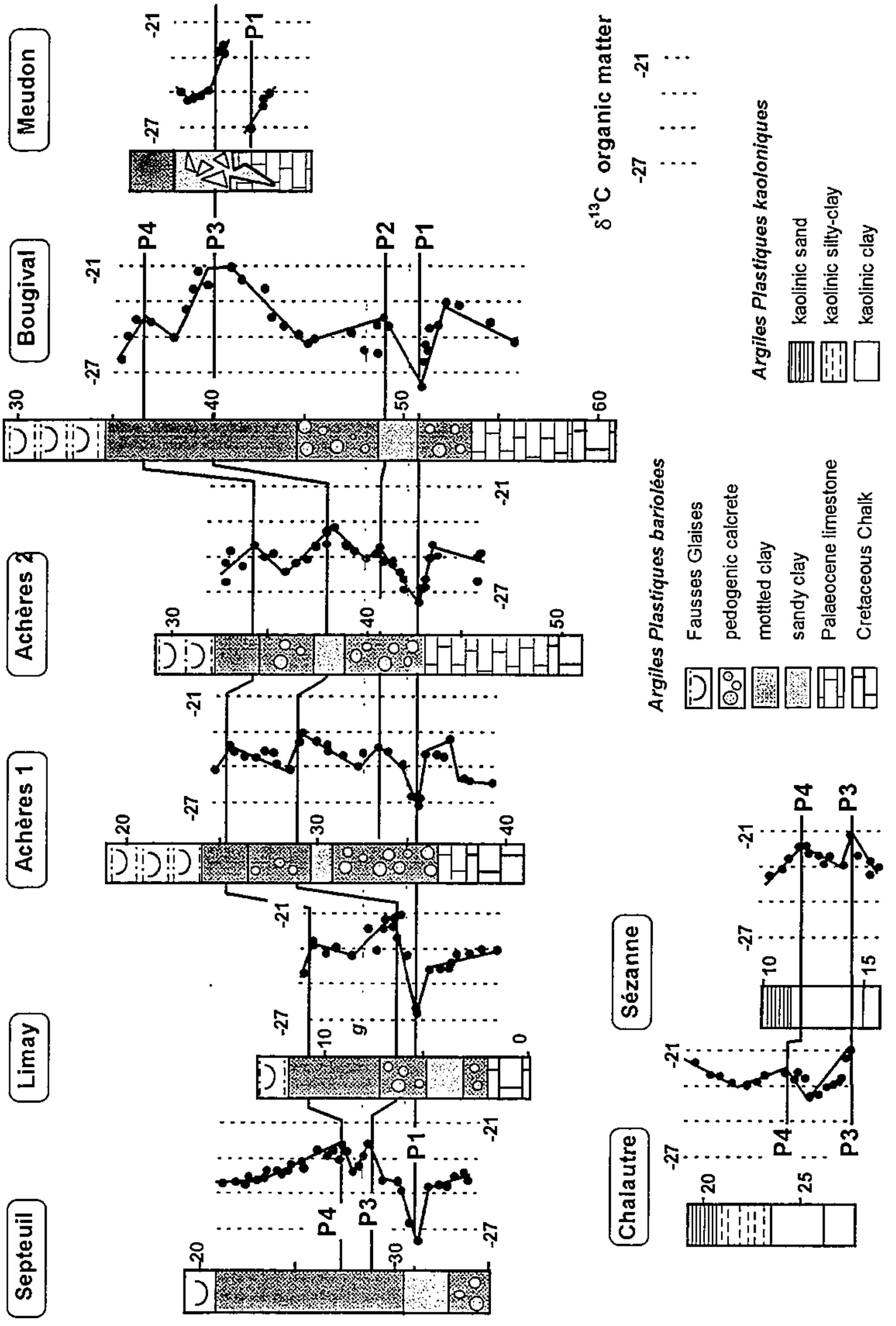

Figure 1 - Correlation of the δ^{13}C values recorded in the Argiles Plastiques sections of the Paris Basin. The sharp negative drop in isotopic values (P1) and the highly possitive isotopic values (P3) are used as "fingerprints" to correlate horizons from section to section. The Argiles Plastiques kaoliniques are equivalent to the upper part of the Argiles Plastiques bariolées.

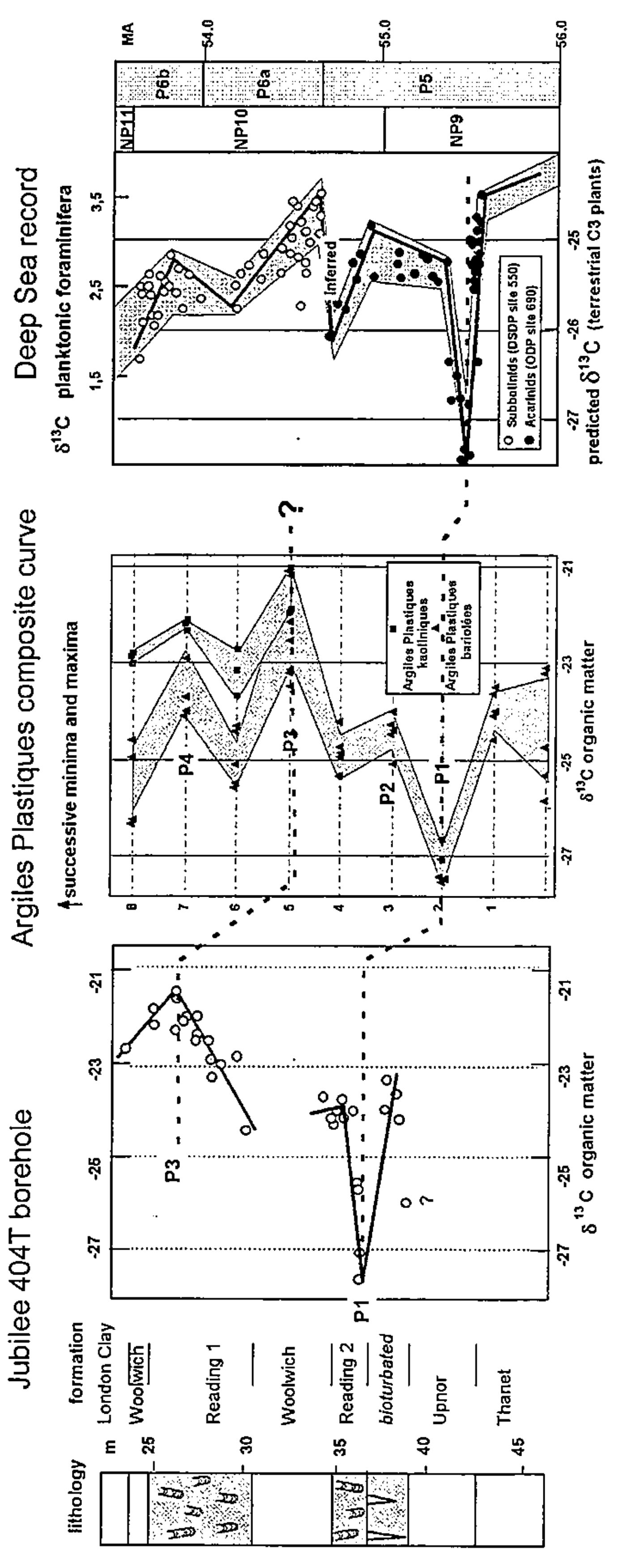

Figure 2 - Correlation of the $\delta^{13}C$ values recorded in the Reading Beds in the London Basin with the $\delta^{13}C$ composite curve constructed from the Argiles Plastiques sections in the Paris Basin. These correlations imply that the Reading Beds in London area are a lateral equivalent of the lower part of the Argiles Plastiques bariolées. The sharp negative drop in isotopic values (P1) at the base of the Reading Beds and Argiles Plastiques Formations has to be related to the negative excursion recorded in the marine realm (Aubry *et al.*, 1996; Stott *et al.*, 1996).

The Reading Beds $\delta^{13}C$ record

The Reading Beds are continental deposits that were deposited in the western London Basin whereas the marine/brackish Woolwich Formation was deposited in the East part of the basin (Ellison *et al.*, 1994). In the London area, the Reading Beds consist of two units. The upper one being progressively replaced eastward by the Woolwich Formation. The lower unit is sandwiched between the Woolwich beds and the Upnor Formation (=Reading Bottom Bed equivalent). Samples from the Jubilee 404 borehole in the Central London have been analysed.

The $\delta^{13}C$ values from the lower Reading unit display a decrease from -24.0‰ to -27.5‰ at 36 m depth and then a progressive increase to about -24.0‰ (Fig. 2). The $\delta^{13}C$ values from the upper Reading unit displays an increase from about -25.0‰ to extremely positive values of -21.5‰ at 26 m depth. The values and amplitude of the negative peak in the lower unit are similar to those of the prominant negative peak at the base of the Argiles Plastiques bariolées. The values of the upper unit are similar to those of the most prominant positive peak. Thus these two peaks represent, respectively, peaks P1 and P3 identified in the Argiles Plastiques bariolées. These isotopic correlations imply that the Reading Beds in London area are a lateral equivalent of the lower part of the Argiles Plastiques bariolées.

Correlation within the continental facies and with the deep-sea record

The $\delta^{13}C$ record of the Argiles Plastiques bariolées displays a coherent pattern of isotopic changes that can be identified in all the sections studied. The $\delta^{13}C$ record consists of several "unique" $\delta^{13}C$ events (i.e. sedimentary horizons characterized by a unique $\delta^{13}C$ values) that can be used as fingerprint to geochemically correlate specific horizons from section to section. When traced across the facies boundaries these unique $\delta^{13}C$ events are also identifiable in the Argiles Plastiques kaoliniques and in the Reading beds. This may allows to develop a regional stratigraphic framework on which the superposition and/or the lateral equivalence of the Sparnacian can be demonstrated.

Moreover, the sharp decrease in $\delta^{13}C$ values from around -24.0‰ to values between -27.0‰ and -28.0‰ has to be related to the negative excursion recorded in the marine realm (Aubry *et al.*, 1996; Stott *et al.*, 1996) (Fig. 2). This excursion constitues a precise datum level that can be recognized and correlated between marine and terrestrial Paleocene/Eocene boundary sequences. It implies that the basal Argiles Plastiques and Reading Beds correspond to the interval centered on 55.5 Ma in mid-Biochron NP9. Remains to discussion the significance of the isotopic record between the level of the excursion and the top of the Sparnacian series.

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TERRESTRIAL AND MARINE PALYNOMORPH ASSEMBLAGES FROM THE DANISH FUR FORMATION, AT THE PALEOCENE/EOCENE TRANSITION.

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Introduction

A palynological study of the marine Late Thanetian to Ypresien Fur Formation and top of underlaying laminated clay in Denmark has been undertaken. The purpose was to make local a spore and pollen zonation and a palaeoenvironmental interpretation of the palynomorph assemblage.

The Fur Formation.

The Fur Formation is a 60 meter thick diatomite deposited near the NE coast of the North Sea Basin. The formation contains a series of ash layers, which are divided into at lower negative series numbered −33 to −1 and an upper positive series numbered +1 to +140 (Bøggild, 1918). The laminated clay or Stolle Klint Clay makes up the lower part of the negative ash series from −33 and downwards in the Haslund Member, lower part of the Ølst Formation (Heilmann-Clausen, 1995).

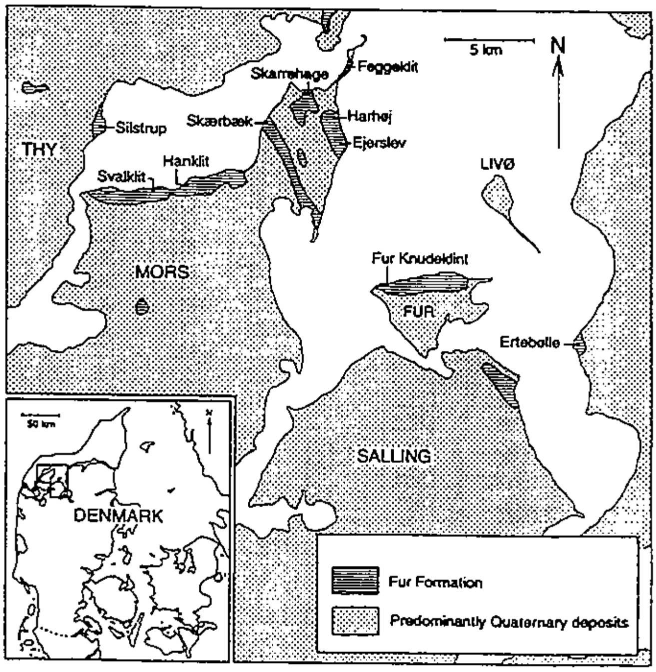

Fig.1. Map of localities where the Fur Formation is exposed (modified from Pedersen & Buchardt 1996).

The study of the palynomorphs is based on 84 samples collected from three different outcrops of the Fur Formation in the northern part of Jutland: Stolleklint, Silstrup Firkanten and Silstrup Sydklint (Fig.1.). The samples were collected with a one meter spacing.

Results.

The samples yielded 45 species of spores and 95 species of pollen. The microfloral assemblage is divided into six pollen zones, A to F (Fig. 2.), and the zones can be correlated with the regional

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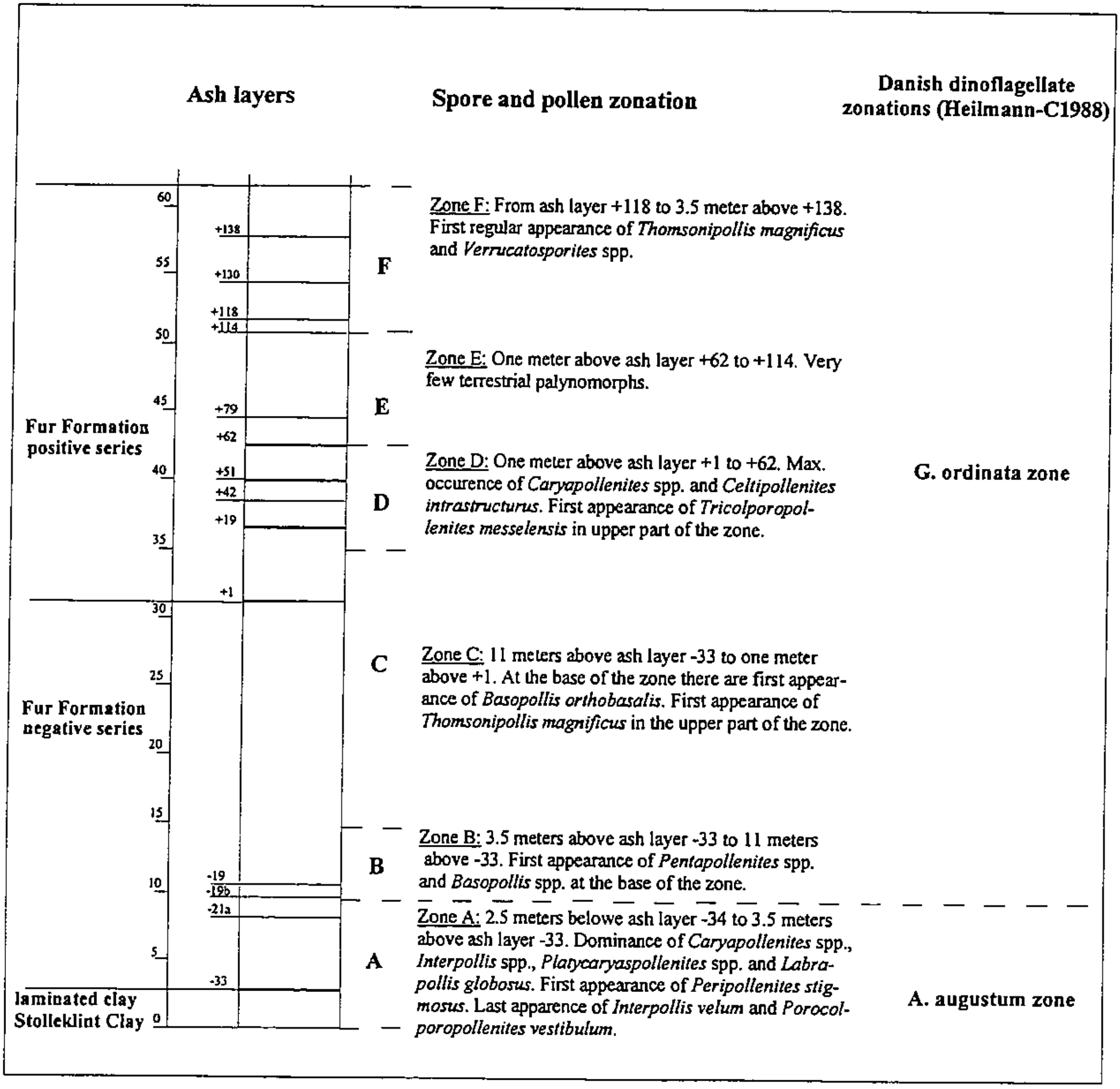
pollen zones 9 to 13a of Krutzsch (1966, 1967, 1970) and the spore and pollen zones SP3-SP4b/ Wu2-Wu3b of Meyer (1988, 1989). Several species of pollen appears earlier than previously recorded by Thiele-Pfeiffer (1988) and Krutzsch (1966, 1970).

The *Apectodinium augustum* (Aau) and *Glaphyrocysta ordinata* (Gor) dinoflagellate zones of Powell (1992) and the Viborg Zone 6 and 7 of Heilmann-Clausen (1985, 1988, 1994) were recognised in the sample succession. These zones correlates with the nannofossil zones NP9 to mid NP10, and supports a Late Thanetian to Lower Ypresien age of the deposits.



Zone F: From ash layer +118 to 3.5 meter above +138. First regular appearance of *Thomsonipollis magnificus* and *Verrucatosporites* spp.

Zone E: One meter above ash layer +62 to +114. Very few terrestrial palynomorphs.

Zone D: One meter above ash layer +1 to +62. Max. occurence of *Caryapollenites* spp. and *Celtipollenites intrastructurus*. First appearance of *Tricolporopollenites messelensis* in upper part of the zone.

Zone C: 11 meters above ash layer -33 to one meter above +1. At the base of the zone there are first appearance of *Basopollis orthobasalis*. First appearance of *Thomsonipollis magnificus* in the upper part of the zone.

Zone B: 3.5 meters above ash layer -33 to 11 meters above -33. First appearance of *Pentapollenites* spp. and *Basopollis* spp. at the base of the zone.

Zone A: 2.5 meters belowe ash layer -34 to 3.5 meters above ash layer -33. Dominance of *Caryapollenites* spp., *Interpollis* spp., *Platycaryaspollenites* spp. and *Labrapollis globosus*. First appearance of *Peripollenites stigmosus*. Last apparence of *Interpollis velum* and *Porocolporopollenites vestibulum*.

Fig. 2. Correlation of spore and pollen zones with the ash layers in the Fur Formation and top of the laminated clay. Dinoflagellate zonation of Heilmann-Clausen (1994) indicated in the figure.

The palynomorph assemblage and palynofacies in zone A are very similar to the Paleocene Subzone 19.4 of Schröder (1991) found in Central North Sea Basin.

Close to 60% of the total spore and pollen assemblage in zone A is constitued by the pollen genera: *Caryapollenites* spp., *Labrapollis* spp., *Interpollis* spp., *Platycaryapollenites* spp. *and*

Tricolporo-pollenites spp. First appearance of *Peripollenites stigmosus*. The pollen species *Interpollis velum* and *Porocolporopollenites vestibulum* have last appearance in zone A. The genus *Apectodinium* dominates the dinoflagellate assemblages in the zone and the species *A. augustum* has last appearance 3.5 meter above ash layer –33 (approximately just belowe ash layer –19b).

At the base of zone B the pollen genera *Basopollis* and *Pentapollenites* have first appearance and the relative amount of disaccate pollen increases from < 1% in zone A to approximately 70% in zone B.

Among the dinoflagellates, the genus *Glaphyrocysta* and the species *Deflandrea oebisfeldensis* dominates the in zone B to F.

The first appearance of the species *Basopollis orthobasalis* is at the base of zone C. *Thomsonipollis magnificus* has first appeareance in the upper half of the zone.

In zone D *Celtipollenites intrastructurus* and the genus *Caryapollenites* spp. have peak occurrences and constitutes up to 70% of the total amount of terrestrial palynomorphs. First and last appearance of *Tricolporopollenites messelensis* in upper part of the zone.

Zone E is nearly barren of terrestrial palynomorphs and is dominated by dinoflagellates.

In zone F *Thomsonipollis magnificus* and *Verrucatosporites* spp. occurs regularly and *T. magnificus* makes up approximately 2% of the total amount of terrestrial palynomorphs.

Paleocene-Eocene boundary interval.

The last appearance of the species *A. augustum* has been used as a marker for the Paleocene-Eocene boundary in the Central and Northern North Sea Basin (Powell, 1988, Schröder, 1991). Accepting this definition the Paleocene/Eocene boundary in the Fur Formation can be placed 3.5 meter above ash layer -33 (approximately just belowe –19b).

Palaeoenvironmental interpretation of the palynomorph assemblage in the Fur Formation.

Following the vegetation model of Boulter & Hubbard (1982) one can interpretate the changes in the microflora assemblages as representing two warm and two cooler periods (Fig.2.). The warm periods are dominated by Eocene paratropical vegetation; the cooler periods are dominated by Eocene ferns and coniferous forest.

The first warm period occurs in zone A, in the laminated clay and in the lowermost part of the Fur Formation, where the palynomorph assemblage is characterised by the pollen genera *Caryapollenites* spp., *Interpollis* spp., *Labrapollis* spp., *Platycaryapollenites* spp. and the dinoflagellate cyst genus *Apectodinium*.

The latter genus is considered to reflect warm conditions (Knox, 1996, Powell et al. 1996), and the species make up 50-70% of the total assemblage of dinoflagellate cysts in zone A.

The changes in both terrestrial and marine palynomorphs indicates pronounced palaeoenvironmental changes. This change could be interpretated as a combination of climatic cooling and rising sea level with the result, that the marine waters in the North Sea Basin became less restricted.

In the zones B and C the climate was probably cooler than in zone A, and the microflora is dominated by coniferous pollen and the *Glaphyrocysta* group and *Deflandrea oebisfeldensis*. The *Glaphyrocysta* group may represent open marine, inner-neritic water masses (Powell et al., 1996) and *D. oebis-feldensis* characterise areas with high primary production (Powell et al., 1996).

In zone D the pollen genus *Caryapollenites* constitutes up to 70% of the total amount of terrestrial palynomorphs, indicating a climatic shift towards warmer conditions or a changed distance to the shore. A decline in the relative frequency of *Glaphyrocysta* and *D. oebisfeldensis* supports that a environmental change takes place from zone C to D.

In zone F the terrestrial palynomorphs indicate a cooler climate, and the relative frequency of *D. oebisfeldensis* and *Fromea gracilis* increases. The genus *Apectodinium* reappears in zone F, although in relative small frequency, 1-3%, of the total amount of dinoflagellate cysts.

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CARBON ISOTOPE AND OTHER CHANGES ACROSS THE PALEOCENE-EOCENE BOUNDARY IN MARINE AND CONTINENTAL SEQUENCES

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One of the more extreme and abrupt episodes of global warming occurred approximately 55 Mya at the Paleocene/Eocene (P/E) boundary. This climatic event (a.k.a. Late Paleocene Thermal Maximum) was accompanied by a prominent 3 to 4‰ negative carbon isotope excursion (CIE) as recorded in both marine and continental sediments. This excursion as well as other geochemical anomalies have become crucial to developing a quantitative understanding of what might have caused this unusual event. More importantly, the carbon isotope excursion has facilitated correlation of this brief horizon in both marine and terrestrial settings thereby allowing investigators to reconstruct, in exceptional detail, both the climatic signal and environmental consequences of the P/E boundary event on a global scale.

This paper examines the CIE as it is preserved in marine and continental sequences. We will discuss how the CIE was recorded in each environment, and how the CIE has facilitated the correlation of sediments from a wide variety of depositional settings. In marine settings, this analysis will include 2 pelagic sequences from the Caribbean (ODP Sites 999 and 1001), as well as a shallow marine sequence from the New Jersey margin. In continental settings, we will examine the CIE as recorded in a sequence of flood plain paleosols from the Bighorn Basin. We will also discuss some of the more significant paleoclimatic and paleoenvironmental implications of these specific correlations.

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