Geologie en Mijnbouw **69**: 187–204, 1990.

Society affairs

1989 Staring Memorial Lecture

Meteorite impact, extinctions and the Cretaceous-Tertiary Boundary*

J. Smit
Institute of Earth Sciences, Free University Amsterdam, P.O. Box 7161, 1007 MC Amsterdam,
The Netherlands

Abstract

At present the two major hypotheses to explain the mass-extinctions at the Cretaceous-Tertiary boundary (KTB) are a large meteorite impact and widespread volcanism.

High resolution stratigraphy across the Cretaceous-Tertiary boundary in the Agost and Caravaca sections in Spain provides a test for these hypotheses. Hard to explain by any sort of volcanism are the shocked minerals, stishovite and the round form and distribution of microtektite remains, which are found worldwide at the KTB. It is now likely that several impacts occurred simultaneously. The low REE-abundances in the KT ejecta layer and quench crystals of clinopyroxene and other crystalline textures in the KTB microtektites betray an impact in ocean-floor basalt, whereas shocked quartz and stishovite favour a continental target. Stable oxygen and carbon isotope studies of both sections offer a glimpse of what catastrophic environmental changes may in fact have caused the extinctions. Carbon isotopes show that photosynthesis was strongly reduced at the KTB. A 8° C rise in ocean surface temperatures indicates that a greenhouse atmosphere followed the impact event. This greenhouse atmosphere may have lasted for several thousands of years.

Introduction

In 1975 I studied and sampled the Cretaceous-Tertiary Boundary (KTB) interval in the Barranco del Gredero, 1.5 km SW of the town of Caravaca in the subbetic zone of southern Spain. Van Veen (1969) realized in his thesis that the KT section in the Barranco del Gredero offered a unique opportunity to study the KTB in detail, because of its rich and well preserved planktonic faunas and continuity of deposition across the KTB. And indeed, the Gredero (Caravaca) section is one of the most complete and undisturbed KTB sections in the world, with the possible exception of the KT section near El Kef in Tunisia. Of particular interest is a 7–10 cm thick dark clay layer which marks the KT boundary. This clay layer (here termed the boundary clay) was earlier sampled by Prof. Hermes in 1962, but he did not realize its significance because Fallot and colleagues (Fallot et al., 1958) had misplaced the KT boundary by about 30 m. It was known for years that in Denmark, in particular at Stevns Klint, a similar boundary clay layer, the 'fiskeler', marked the KT boundary (Birkelund & Bromley, 1979). Luterbacher & Premoli Silva (1964) also described a thin clay layer at KTB from several sections in pelagic 'Scaglia Rossa' facies, in the Italian Appenines.

---

* The 1989 Staring Memorial Lecture of the Royal Geological and Mining Society of The Netherlands was presented in Utrecht, October 19, 1989

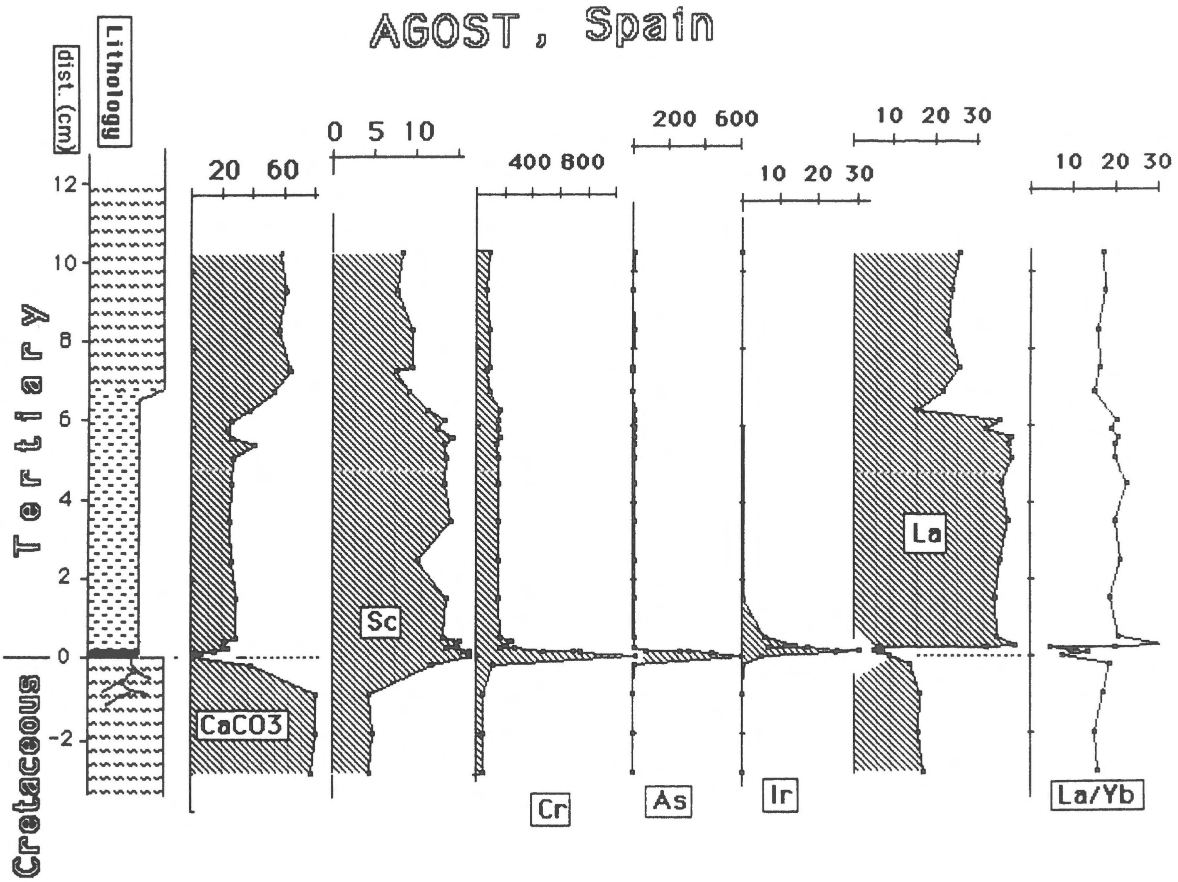

Fig. 1. Patterns of some important trace elements across the KTB in the Agost section (Carbonate in %, Sc-La in $\mu g/g$, Ir in ng/g). Most lithophile elements follow the distribution of Sc, which is highly (negatively) correlated with carbonate content. La is highly correlated with Sc, with the exception of the fallout lamina values. The low REE and low La/Yb suggest a contribution of excavated REE-depleted mantle in the fallout lamina.

In the Caravaca section the KTB interval and especially the KT boundary clay layer can be subdivided in a number of distinct lithological subunits. Although elsewhere these subunits are usually mixed, the succession of these lithologies can be recognized in all complete KTB sections around the world (Smit & Romein, 1985). Some exceptions to this rule occur, for instance in most sections in the Negev desert (Ein mor, HorHaHar) where the KTB is lithologically not clearly visible. The KTB interval is there completely homogenized by bioturbation. Because of the illusion of a gradual KTB transition the suggestion was once made that these sections could be the most complete in the world (Magaritz et al., 1984).

The Caravaca section has some special benefits over other, well known KTB sections, such as the Brazos River section in Texas (Hansen et al., 1985), the mentioned Negev sections in Israel, and the majority of the Deep Sea Drilling Project Holes through the KTB. Sedimentation is continuous and sedimentation rates are high. Even more important, the rate of bioturbation at Caravaca slowed down considerably just at the KTB. This has obviously been caused by whatever happened at the KTB, because all other sediments in the section from the Senonian through the Eocene have been completely homogenized by digging and foraging critters. Empirical studies and theoretical work on the dispersion of thin ash beds or microtektite layers (Officer & Lynch, 1982) in bioturbated pelagic sediments has shown that sedimentary particles or geochemical anomalies can be smeared over 40–60 cm. In the Caravaca section the bioturbational mixing came to a stop at KTB, and remained at low levels for a few thousands of years thereafter, and we are thus able to identify mm thin laminae. A few discrete *zoophycos* or *planolites* burrows do exist, but plowing echinoids, for instance, were temporarily removed from the area, although there are no obvious extinctions known among echinoderm taxa at KTB (Stokes, 1979). These preserved individual laminae allow us to study and discriminate time-slices of less than 100 years and unravel the successive KTB events with high precision.

Recently we discovered two new complete KT sections in southern Spain, near the towns of Agost and Relleu, in the prebetic zone. Lithology, (bed-by-bed), induration, sedimentation rates, the KT boundary clay layer, foraminiferal content and degree of bioturbation, and even the peak iridium values are identical to the Caravaca section (24.6, 26 and 26.6 ng/g), although both new sections are more than 100 km away from Caravaca. By comparing the data of the new sections with Caravaca we may have important tools to separate local or even regional phenomena (diagenesis, faunal differences, depth of deposition) from supra-regional and global effects.

Stratigraphy of the KT boundary interval

In southern Spain the KTB interval consists of (hemi)pelagic uppermost Maastrichtian and Paleocene marls and limestones with a diverse and abundant microfauna, interrupted by the KT boundary clay layer. The marls are dominated by pelagic calcareous organisms, mixed with a 20% detrital clay and silt. The pelagic marls/limestones extend for more than 100 m down in the Maastrichtian and were deposited at a rate of 40 mm/kyr. In the uppermost Maastrichtian in the Caravaca and Agost sections no significant lithological changes occur towards the KTB. This lack of change is confirmed by trace element, carbonate (Fig. 1) (Smit & Ten Kate, 1982) and stable isotope analysis of both sections (Fig. 2). From this lithological evidence alone one can infer that no significant environmental changes up to the KTB occurred. The upper Maastrichtian marls are sharply overlain by a pure clay layer or lamina, barely 2 mm in thickness.This lamina contains the remains of meteorite and impact debris and is here called the fallout lamina because it probably represents the distal fallout of a large impact event. Both the lower and the upper boundary of the fallout lamina are sharp, indicating that it was deposited in one event.

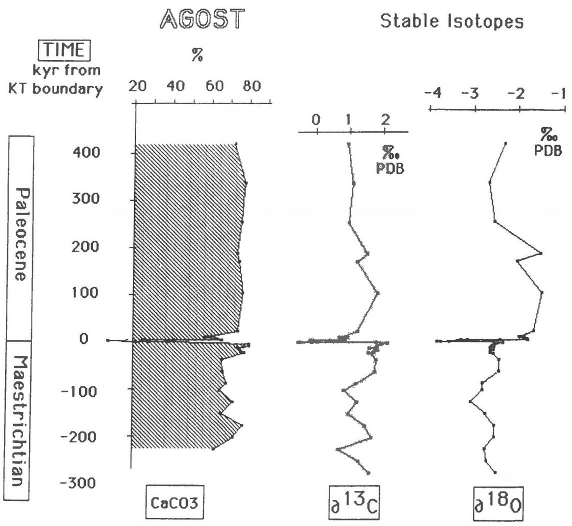

Fig. 2. Stable Carbon and Oxygen isotope profiles across the KTB in the Agost section. δ 13C co-fluctuates with carbonate in the Maastrichtian. The stratigraphic vertical scale is linear with time, converted from sedimentation rates which are calculated from magnetic reversal ages of the Agost section (Groot et al., 1989).

The fallout lamina is directly overlain by the dark boundary clay, actually a clayey marl with about 25% carbonate. The boundary clay is characterized by very low planktonic foraminiferal and nannofossil abundance. New planktonic foraminiferal faunas are not present yet. The thickness of the boundary clay layer is correlated with normal sedimentation rates in the upper Maastrichtian and lower Paleocene, and varies from 3 cm in the Relleu section to 6.5 in the Caravaca and Agost sections and is highest, 20 cm, in the Kef section in Tunisia. It therefore represents a normal sedimentary influx of clay and carbonate and not material from other sources, such as volcanic eruptions or impact debris, because such input would have a thickness independent of the local sedimentation rates. The

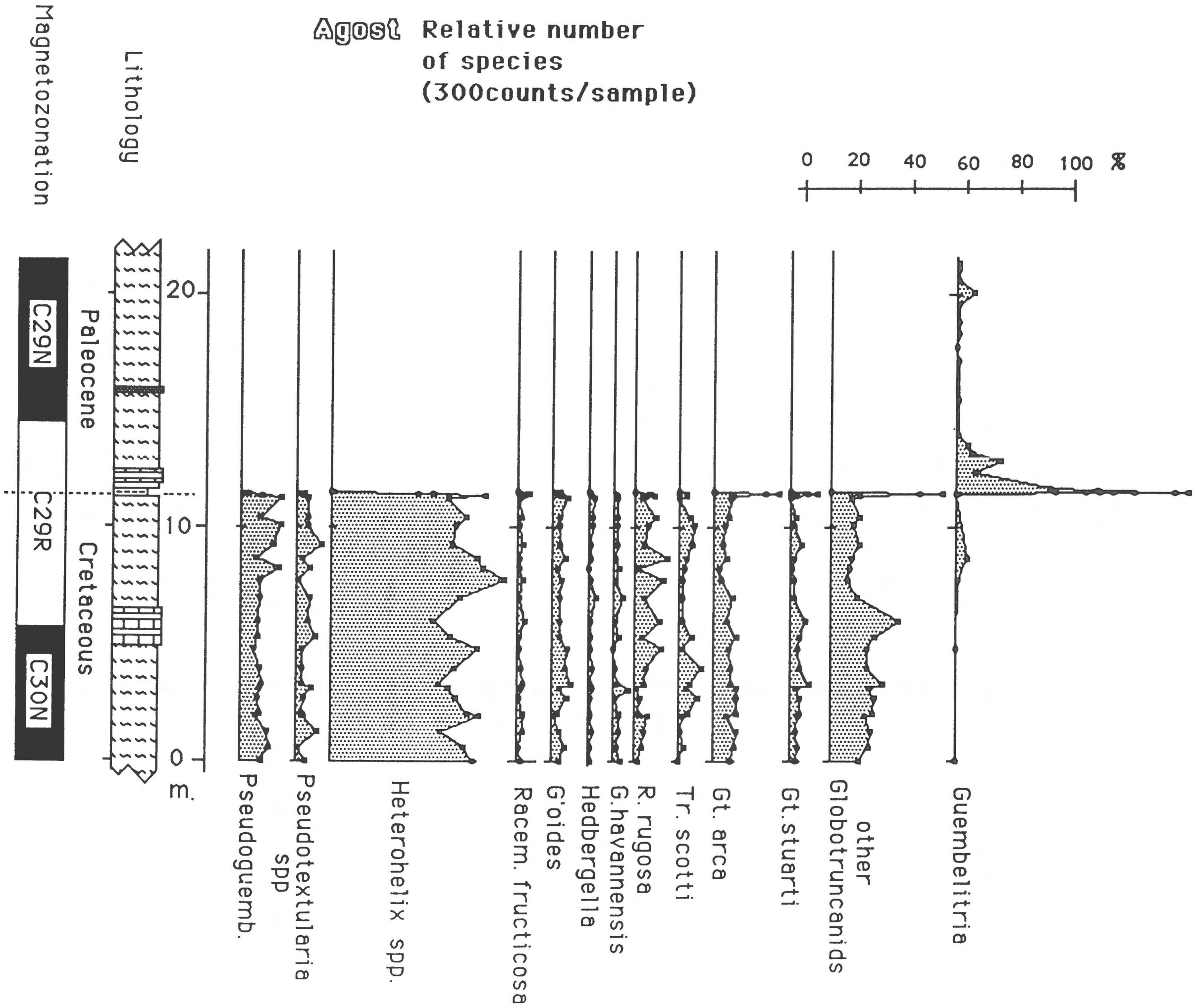

Fig. 3. Relative abundances of planktonic foraminiferal genera and species below the KTB in the Agost section, Spain. There are no systematic trends towards the KTB. The spike of Globotruncanids just below KTB is due to dissolution, which affected preferentially heterohelicids.

upper boundary of the boundary clay is gradual and marks the gradual return of a completely new nannofossil flora and planktonic foraminiferal fauna. Geochemically and isotopically the lower Paleocene sediments are almost indistinguishable from the upper Maastrichtian marls, although sedimentation rates are still reduced to half the Maastrichtian rate (20 mm/kyr). Primary production is also lower as in the upper Maastrichtian as confirmed by carbon isotope analysis (Zachos et al., 1985; Fig. 2).

In 1989, members of the working group on the Cretaceous-Paleogene boundary, (which is a working group of the subcommissions of the Cretaceous and Paleogene, of the International Commission of Stratigraphy) have voted the 'golden spike' for the Cretaceous-Tertiary boundary to be at the base of the boundary clay, in the Kef section in Tunisia. This is at the iridium anomaly (fallout lamina), at the major extinction event. Biostratigraphers usually prefer to place zonal boundaries at the first appearance of new species. This practice is not followed by the working group members, because the first appearance of new species of different groups (nannofossils, foraminifers, dinoflagellates) at the base of the Paleocene is at different levels. The base of the boundary clay is the least problematical.

Planktonic foraminifers

Paleontologists already knew for a long time that almost all planktonic foraminifers became extinct at KTB (Luterbacher & Premoli Silva, 1964) to be replaced by entirely new foraminiferal taxa. Because foraminifers are diverse, occur in large numbers and because they all become extinct at KTB, planktonic foraminifers are an ideal subject of study to understand the underlying extinction mechanisms. Brummer & Kroon (1988) and Troelstra (1989) have shown that planktonic foraminiferal populations are very sensitive indicators of environmental changes. Shifts e.g. in relative abundances of species should be expected when environmental parameters as temperature, salinity, waterdepth or available nutrients change. However, quantitative analysis of uppermost Cretaceous planktic populations of the Agost and Caravaca sections do not show any significant changes, even when approaching the KTB within centimetres (Fig. 3). Planktonic foraminifers suddenly became extinct at KTB without leaving a trace in the uppermost Cretaceous geological record. From information gained from recent living planktonic foraminifers we may thus conclude that there was little or no change in climate before the actual KTB event, as marked by the clay layer.

A sudden extinction assumes that the sections are complete and that no significant hiatus exists. The best method to assess the extent of a possible hiatus at KTB is by paleomagnetic analysis. The magnetostratigraphy (Fig. 4) of the KTB interval of the Caravaca and Agost sections is complete. Chrons C3ON, C29R and C29N are present in both sections. The KTB occurs in C29R, which has a duration of about 500.000 years. The deposition of normal pelagic marls in C29R takes already this much time. At most, given the errors, the equivalent of about 30.000 years of less of sediment could be missing at KTB in both sections. But whatever small hiatus exists the events which have lead to extinction, the extinction itself, the repopulation and diversification by new faunas, all have to take place in the oceans at KTB in less than 30.000 years.

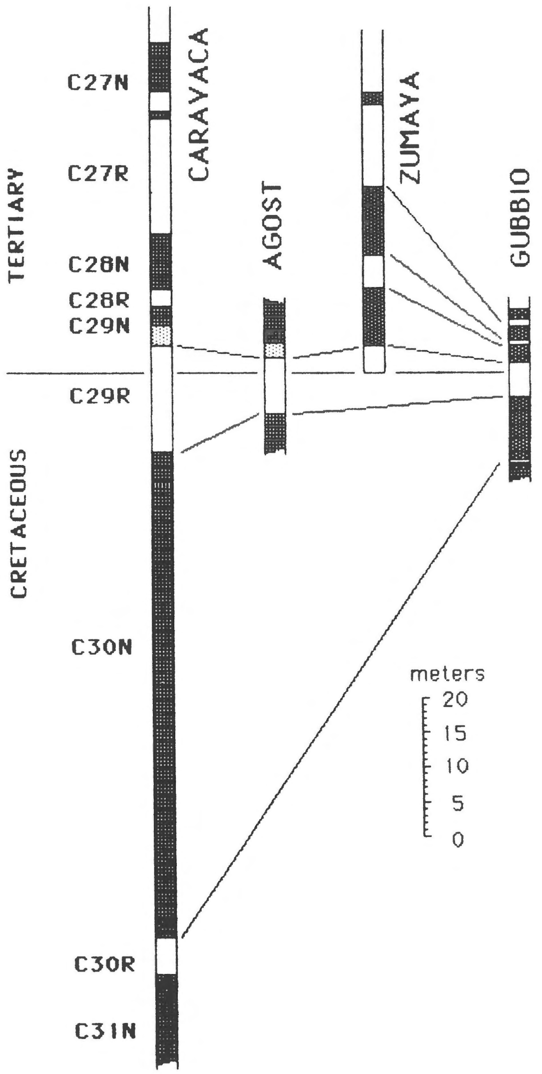

Fig. 4. Comparisons of the magnetic reversal pattern across KTB in four complete KTB sections.

A few dissenters remain. Keller (1988) recently reported foraminiferal extinctions somewhat before the KT boundary at El Kef, Tunisia. She regarded the actual extinctions at KTB to be just one step, be it the largest, in a series of stepwise extinction events around KTB. We examined the same

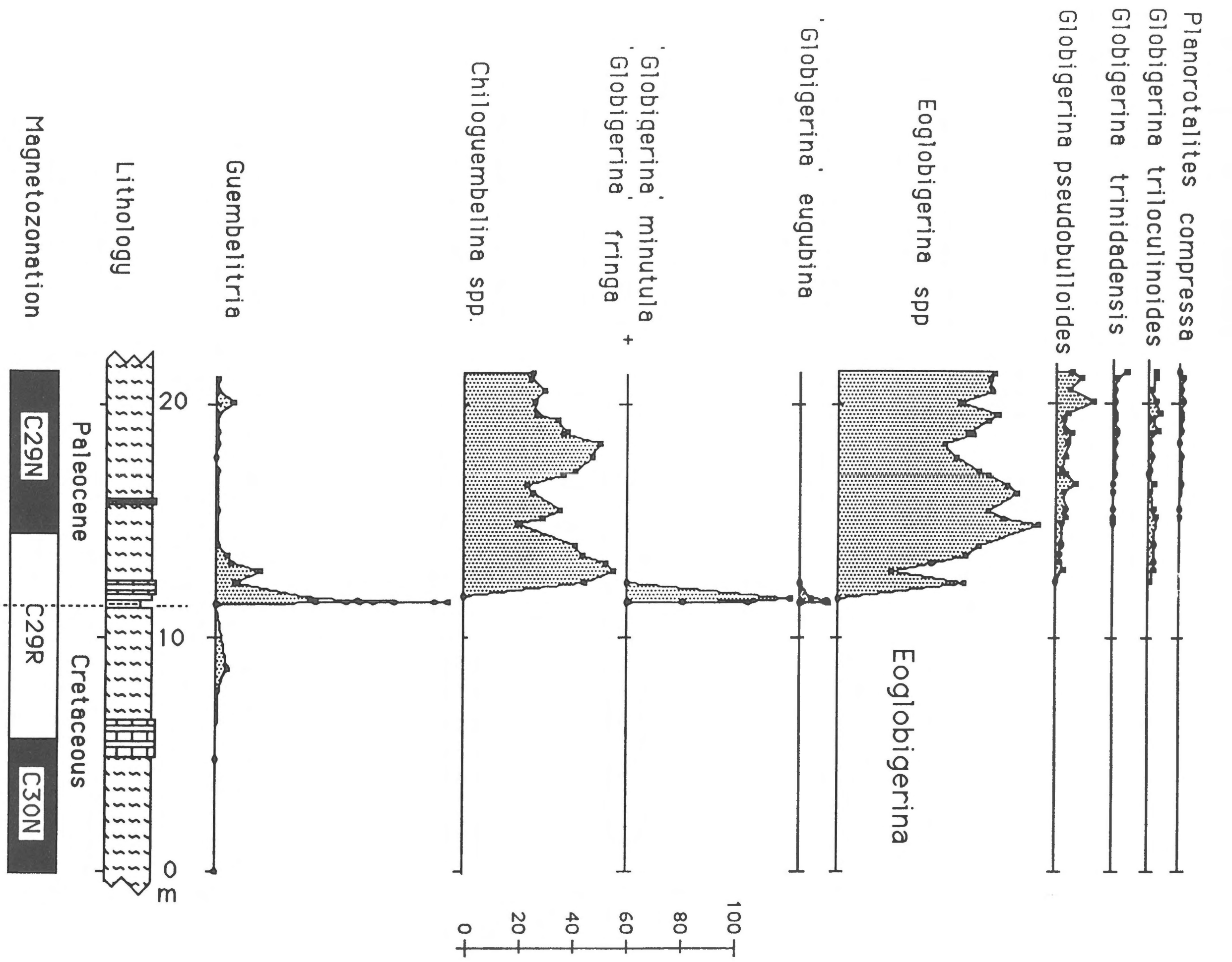

Fig. 5. Relative abundances of Paleocene planktonic foraminifers in the Agost, Spain section. Only *Guembelitria cretacea* is a clear survivor, has a high relative abundance in the boundary clay, but seems to have few descendants. The first 50 cms (50 kyr after the KTB extinctions) of the Paleocene are characterized by rapidly changing abundances of species with short geological lifetimes.

samples but found the same species to range up to the boundary level. Some of these species (*Ventilabrella multicamerata, Racemiguembelina fructicosa*) are fairly easy to recognize. We do not know what caused the discrepancy between our and Keller's analysis, but a new sampling trip (1991) under auspicium of the Global Sedimentary Program may straighten out the differences. The samples will there be carefully sampled and coded by a neutral referee, like the 1989 'blind sampling' trip to the Bottaccione gorge near Gubbio in the Italian Appennines, where Robert Ginsburg supervised a carefull re-sampling of clay layers above and below the KTB, where shocked minerals and spherules were reported (Naslund et al., 1986).

The fallout lamina does not contain foraminifers. The 6.5 cm boundary clay contains a poor, dwarfed fauna. A few reworked larger Cretaceous planktonic foraminifers occur, but small species like *Guembelitria cretacea, Hedbergella monmouthensis* and *Globigerinelloides messinae* are somewhat more abundant relatively. At the gradual upper boundary of the boundary clay, new Paleocene foraminifers appear in the sections, such as *'Globigerina' eugubina, G. fringa.* The earliest faunas are pioneering faunas, characterized by rapid turnover of species and rapidly changing species abundances (Fig. 5). Apparently these earliest Paleocene taxa filled the ecological space left by the upper Maastrichtian fauna. But since they were the first to do so, they were obviously not the best adapted. More stable faunas, characterized by longer living species like *M. pseudobulloides* and more equable relative species abundances, soon replaced these pioneers.

I will focus here on the 'fallout' lamina and overlying boundary clay, because these layers contain the evidence for meteorite impact or evidence for widespread volcanism. The boundary clay documents the environmental changes leading to mass extinction.

Table 1. Instrumental neutron activation analysis of the fallout layers at KTB (μg/g)

	Zumaya	Caravaca	Agost
Na	1830	1490	2738
K	12370	13900	30270
Ca	116000	–	48000
Sc	7.08	9.46	16.51
Cr	451	801	747.5
Fe	119500	52200	110000
Co	491	537	61.05
Ni	1430	1990	357.5
As	14130	456	442.3
Br	3.54	4.3	8.91
Rb	60.27	–	113
Sr	2385	–	443.9
Sb	54.62	13.7	6.65
Cs	3.88	–	1.5
Ba	601	–	308
La	29.08	3.78	5.78
Ce	49.33	9.19	15.55
Nd	38	–	–
Sm	3.59	0.735	1.09
Eu	0.767	0.204	0.315
Tb	0.191	–	–
Yb	0.743	0.557	0.415
Lu	–	–	–
Hf	2.714	4.66	7.4
Ta	1.5	–	1.4
Au	–	23.3 ng/g	33.9 ng/g
Ir	26.3 ng/g	26.6 ng/g	24.6 ng/g
Th	13.59	8.59	8.75
U	47.7	11.7	26.9

The fallout lamina

Trace element analysis of the fallout lamina of Agost and Caravaca, shows this lamina to be entirely different from all other sediment in these sections (Fig. 1). There is, on the other hand, a remarkable similarity in composition between the fallout laminae of the Agost, Caravaca and Zumaya sections (Table 1). Alvarez et al. (1980) and Smit & Hertogen (1980) concluded on the strength of the iridium anomaly that an extraterrestrial object must have hit the earth at KT time. Alvarez et al. (1980) suggested that supression of sunlight by the upthrown dust cloud of a single impact could have led to a cessation of photosynthesis and collapse of the food-chain, with resulting extinctions. Of course this picture was much too simplistic, but it still works as the major trigger mechanism for the KT mass-extinction. It did on the other hand not take into account other environmental changes that may have operated at the end of the Maastrichtian, and some groups as dinosaurs, ammonites, rudists, inoceramids (Ward, 1988) may have suffered species diversity reductions before the actual KT boundary. Others contend that extinctions extended beyond the KT boundary (Rigby et al., 1987; Keller, 1988). The magnitude of these extinctions, however, are almost negligeable (expressed as species extinct/yr) against the mass-extinctions which take place precisely at the KT boundary.

Hypotheses as impact related extinctions, or volcanic eruption related extinctions should be tested. Predictions which come forth from either of the two rival hypotheses will, if they are true, (hopefully) support or reject one of the rival hypotheses, although scientists seldom give up on their favourite hypothesis.

Predictions of the impact hypothesis include the finding of a large impact crater, other remains of the bolide, and ejecta products from the excavation of the impact crater.

A definitive crater has up till now not been found. Although the Manson crater in Iowa (Fig. 6) has the appropriate age (Kunk et al., 1989), it is too small (diameter 32 km). If the impact, or one of the impacts, has taken place in the ocean, it may

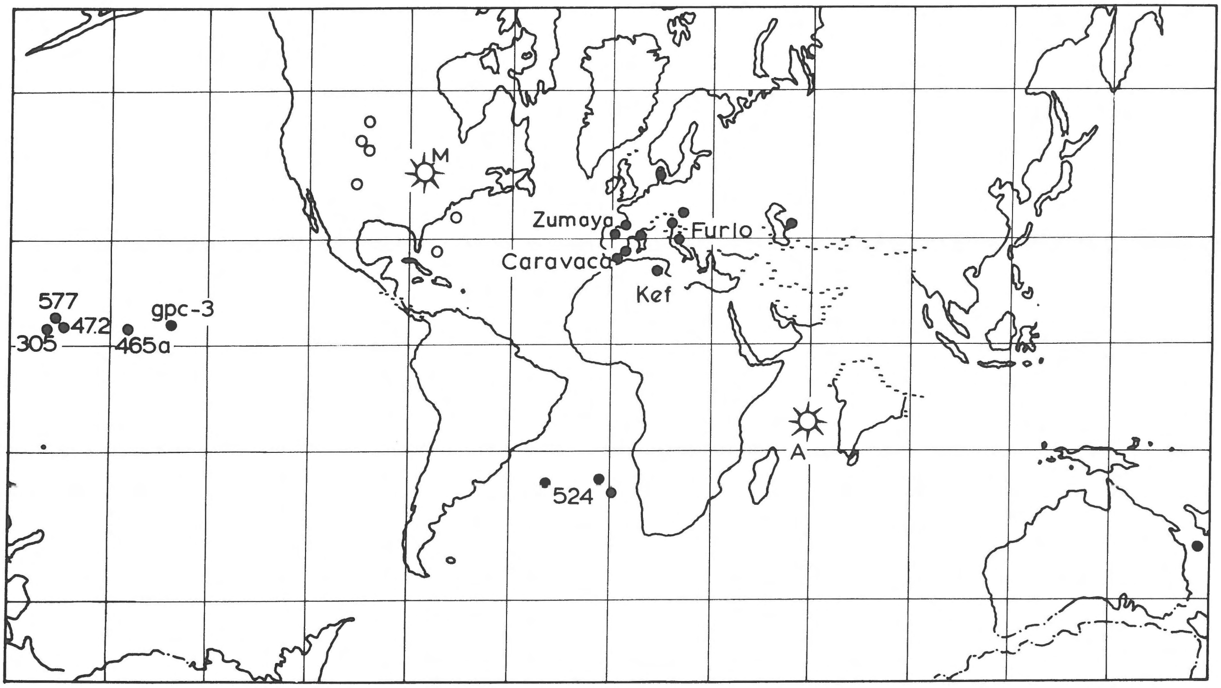

Fig. 6. Paleogeographic map (early Paleocene time) modified from Smith & Briden (1977). Open circles are KTB sites with larger non-crystalline spherules and droplets, closed circles KTB sites with crystalline spherules. The possible KT bolide impact craters Manson and Amirante basin are indicated.

have escaped detection, or it may have been subducted. Over 51% of the ocean seafloor has been subducted since KT times. An interesting speculation (Hartnady, 1986) is that the Amirante basin south of the Seychelles islands is an impact site. An impact there may have triggered the formation of the Deccan-Reunion hotspot and the eruption of the Deccan traps, considered by some the largest outpouring of plateau lavas in the Phanerozoic. Recent dating of the Deccan traps (Courtillot et al., 1988) suggests, however, that the Deccan traps began erupting in Chron 30 N, predating the KTB some 0.5 Ma.

Schmidt & Holsapple (1982) estimate that the amount of target rock melted or vaporized by the impact is 100 times the mass of the bolide. Over 1000 times the mass of the bolide of target rock will show evidence of shock or even shock metamorphosis (shatter, cones, shock lamellae, stishovite).

Approximately 15 times the bolide mass of vaporized, molten and shocked crater material will be ejected in the atmosphere. Some of the material, in forms such as tektites, is ejected out of the atmosphere at enourmous speed (up to 9 km/s) along ballistic trajectories. The largest amount will be deposited, as estimated from Martian craters or the Miocene Ries crater in Germany, within the crater and up to 3 or 4 crater radii away from the crater as a thick proximal ejecta blanket. The glassy material in the proximal ejecta is known as suevite.

Some high speed impact ejecta are known as tektites and microtektites. These ejecta products are made of pure glass, extremely low in volatile constituents, such as water, which makes them different from all known terrestrial volcanic glasses. There is a bimodal size distribution in tektites. The larger tektites are 1–25 cm in diameter, and have many typical forms. The smaller microtektites are 0.05–1 mm in diameter. Most microtektites are spherical. In total mass microtektites are more abundant than tektites, and can be distributed globally, and were expected to be found at KTB (Glass, 1982).

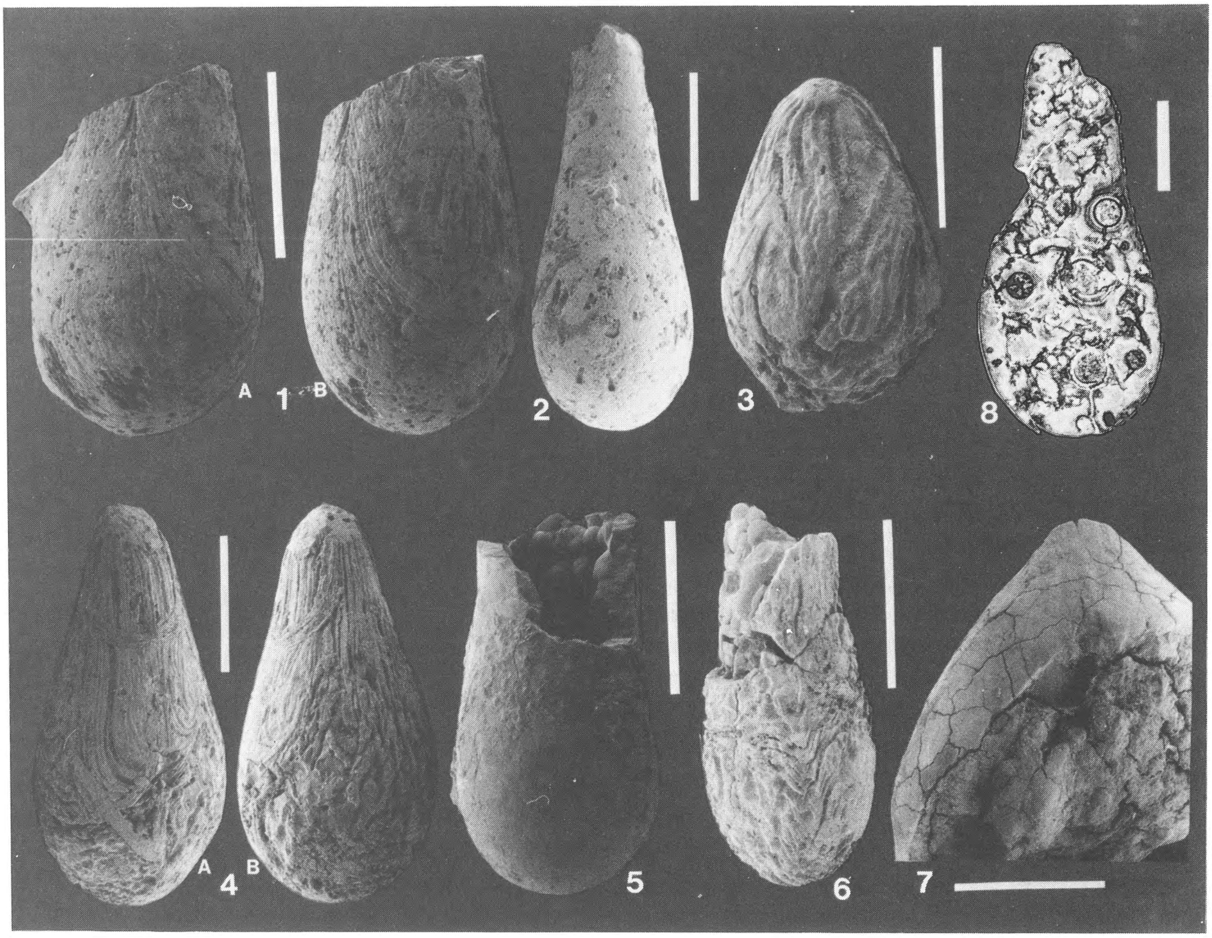

Fig. 7. Droplets of the KTB American microtektite strewnfield. Especially the forms from Dogie Creek, Wyoming (Bohor et al., 1987) show flowlines on the surface (1–6). Such flowlines probably originate because areas with tektite glass of slightly different composition are etched differently by terrestrial weathering. Similar flowlines are well known from moldavite and australasian tektites. 7: droplet from DSDP Hole 390a. 8: thin section of a droplet from DSDP Hole 603 (from Klaver et al., 1987). The original gas-bubbles are filled with authigenic smectite. Bars of 1–6, 8 are 0.5 mm, of 7: 0.25 mm.

Microtektites

Larger tektites have not been found at KTB. Microtektites s.s., which are made of glass by definition, have not yet been found at KTB, but diagenetically altered analogs – of the same size and form – to microtektites have been found globally at KTB. They are provisionally referred to as 'microtektite-like spherules' (Smit & Romein, 1985). In Agost and Caravaca these spherules are essentially confined to the fallout lamina. The density distribution of the spherules in most KTB sites is about 200–400 cm^2, a density comparable to that of true microtektites in e.g. the australasian tektite strewn field found in deep sea piston cores (Glass, 1982).

'Microtektite-like' spherules at KTB occur in two distinct populations. One group is confined to North America (Fig. 6). The KTB layers in the US western interior and DSDP sites 390A and 603 (Klaver et al., 1987) near the east-coast of North America contain non-crystalline spherules and droplets with exquisite flow-lines on the surface (Fig. 7). All are completely altered and are distinctly larger (Fig. 8) than the crystalline spherules

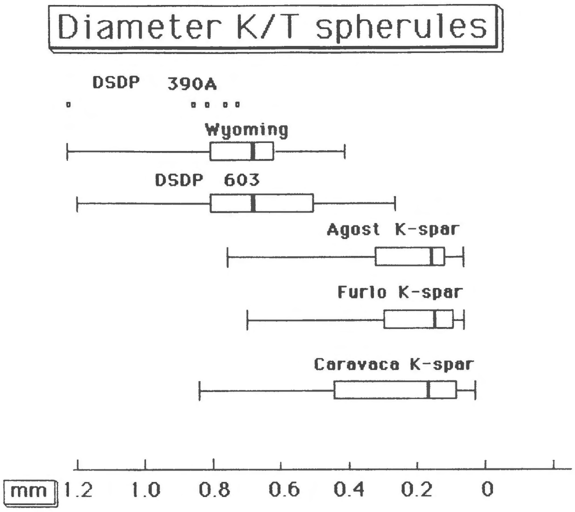

Fig. 8. Box and whisker plots of the diameter of KTB microtektite-like spherules. The non-crystalline spherules from North America and DSDP holes near the east coast of North America are clearly larger than the crystalline (mafic) spherules found elsewhere at KTB.

of the other population at KTB. The majority of the KTB spherules contain a relict dendritic crystal texture of potassium feldspar (Figs 9A, B) or smectite (Figs 9C, D). These crystalline dendrites have formed pseudomorphically after a precursor mineral, which must have quench-crystallized in a mafic liquid, analogous to quench crystals in submarine pillow lava's (Klaver, 1988). The original glass phase has never been found, but recently at one place, at KTB in DSDP hole 577 on the Shatsky Rise in the west Pacific, we have found a few spherules made up of clinopyroxene crystals in a typical dendritic texture (Figs 9E, F). These cpx dendrites are identical in form to the K-spar (Fig. 9B) and smectite (Fig. 9C) dendrites found elsewhere at KTB, and are presumably the precursor mineral of these altered spherules at KTB. In Hole 577 transitions from cpx to smectite and K-spar are found within one and the same spherule (Fig. 10A). Other mafic minerals as plagioclase or olivine form morphologically different quench crystals (Lofgren, 1980)

In dark smectitic spherules (Fig. 10B) from KTB in the Italian Appennines, South Atlantic (DSDP Hole 524), and Pacific (DSDP Hole 577, GPC-3), magnesioferrite-hercynite spinels, with a distinct skeletal morphology, also have survived diagenetic alteration (Smit & Kyte, 1983). Solid-solution of Fe, Al, Mg and Cr in the spinels, point to a high crystallization temperature, because otherwise the spinels would exsolve to pure end-member compositions. Also these skeletal spinels originated in a mafic liquid and occur together with cpx in DSDP Hole 577. Glass et al. (1985) reported highly similar cpx and spinel skeletal crystals within mafic spherules from the late Eocene (Fig. 10C).

Alternative explanations have been brought forward for the origin of the KTB spherules. Hansen et al. (1986) explained the K-spar spherules as infilling of prasinophyte algae, because after dissolution in strong acids of K-spar spherules he has found fatty-acid membranes in the residue. We repeated the experiment of Hansen carefully with over 600 cleaned and handpicked spherules from Caravaca, but could not find a single membrane relict. High temperature minerals are also hard to explain as low temperature infillings.

Izett (1987) explained the K-spar spherules as purely authigenic, because he found mostly irregular K-spar fragments in his microprobe x-ray maps of the Caravaca fallout lamina. Izett, however, overlooked the incomplete alteration to K-spar; most spherules at Caravaca are mixtures of more fragile smectite and patches of K-spar, and these spherules are all compressed by compaction. Round spherules at Caravaca or elsewhere at KTB which escaped compaction consist entirely of K-spar.

Naslund et al. (1986) believed the spherules to be of volcanic origin, and reported 'similar spherules' also from other clay layers above and below the KTB in the Gubbio section in Italy. They failed, however, to demonstrate a relict crystalline texture in non-KTB spherules. Volcanic spherules are known as 'Pelea's tears' originating in lava fountains in Hawaii, and from some DSDP sites close to mid-ocean ridges (Fig. 10D). All these 'tears' have a low viscosity mafic composition. The quench crystal morphology in the 'tears' is similar to the KTB spherules.

Volcanic spherules are spherical because of the very low gas-content of the mafic lavas and all

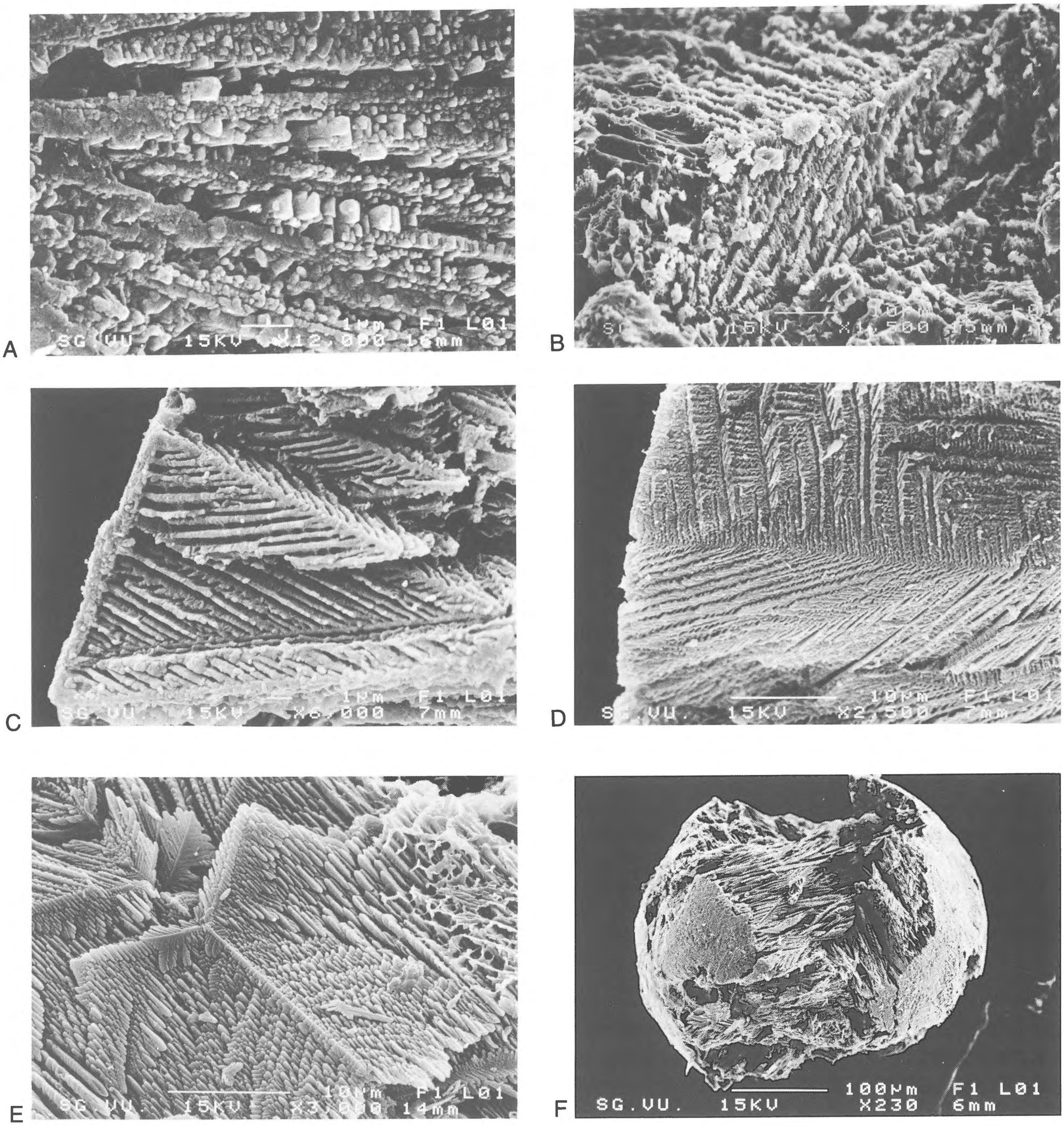

Fig. 9. A: SEM graph of a K-spar dendrite in a microtektite-like spherule from the KTB at Furlo, Italian Appennines. The dendrite is a pseudomorph after clinopyroxene (see Fig. 9F). B: Detail of a K-spar dendrite from within a KTB spherule from Caravaca. SEM graph. C: Dendrite, composed of clayminerals, from the interior of a spherule of KTB of DSDP Hole 577a, core 12 section 4, 70 cm. Note the similarity to the cpx dendrites in Fig. 9F and to the K-spar in Fig. 9B. The curved rim of the spherule is visible to the left. SEM graph. D: Typical aspect of a smectite dendrite from KTB spherule of Hole 577A-12-4-70 cm. SEM graph. E: Broken spherule filled with a dense mesh-work of clinopyroxene (cpx) quench crystals of DSDP Hole 577a-12-4-68 cm. F: Detail of a cpx dendrite in a broken spherule from KTB in DSDP Hole 577a-12-4-70 cm. The cpx dendrites nucleate probably on a tiny fragment in the middle (Spinel?) (Klaver, 1988). The cpx dendrites form probably fractal series, and therefore one can observe identical dendrites at different scales. SEM graph.

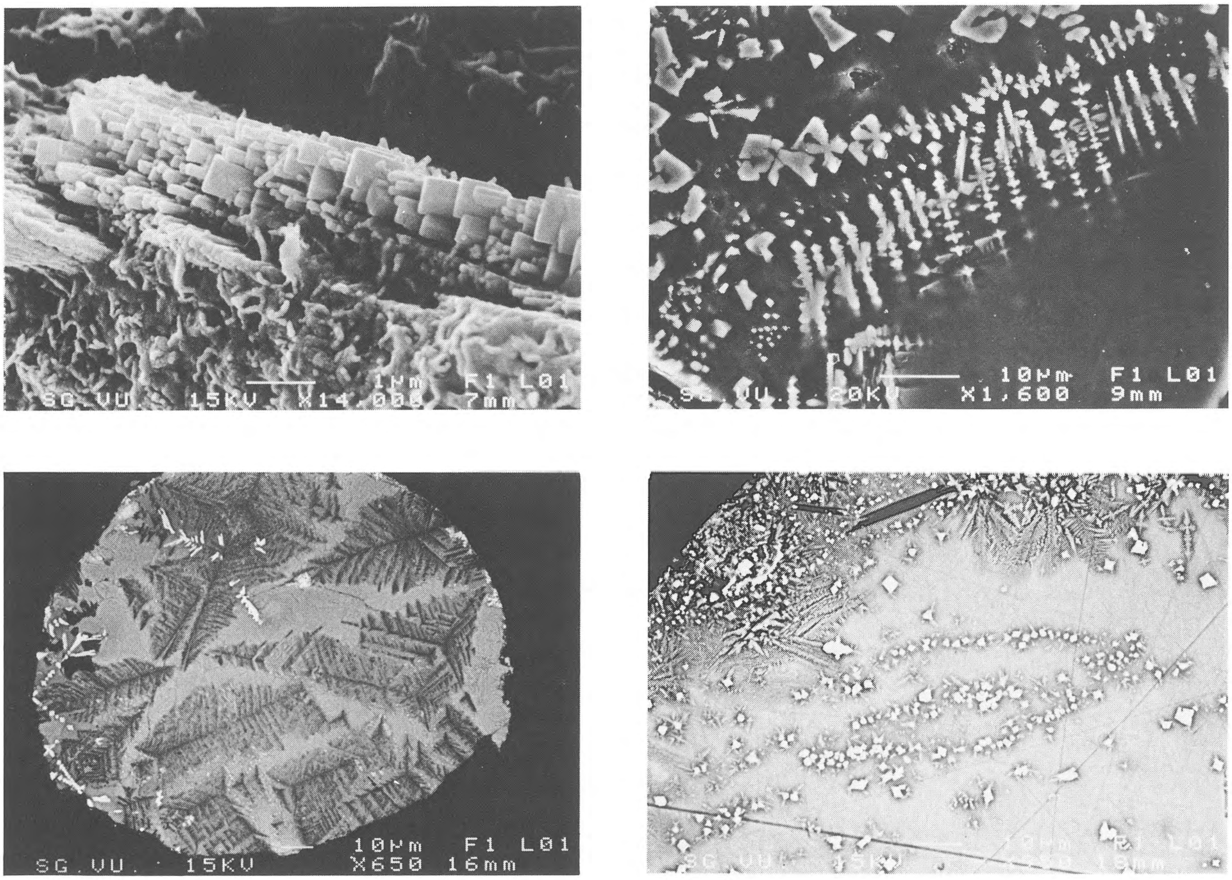

Fig. 10. A: Dendrite of tiny K-spar crystals in the interior of a KTB spherule, enbedded in smectite. Compare with Fig. 9B. B: Backscatter SEM graph of skeletal spinel crystals within a dark spherule from the KTB of Furlo, Italian Appennines. The spinels are embedded in a claymineral matrix, and are identical in form to skeletal titano-magnetite from pillow-lavas, and mafic spherules of volcanic origin (see Fig. 10D) or from the Eocene cpx-bearing microtektite strewn field. However, the KTB spinels have more Al, Mg and Cr in solid solution with Fe in the spinel. C: Backscatter SEM graph of an upper Eocene microtektite from DSDP Hole 166. The cpx crystals are partially etched out from the glass matrix. Skeletal spinels (magnetite) grows in between the cpx dendrites. D: Backscatter SEM graph of a mafic volcanic spherule from DSDP Leg 5. Some magnetite skeletal grains form the nucleus for cpx quench crystal growth. Towards the edge of the spherule (upper left) more spinel and cpx crystals occur.

occur very close to the eruptive vents. Mafic spherules from KTB occur up to 2000 km or more from the nearest volcanic source, and are thus difficult to explain by such non-violent volcanic eruptions. Although highly improbable, an extremely violent volcanic eruption may be capable to transport such spherules over large distances. Such violent eruptions are driven by high gas pressures, and therefore the erupted tephra are not spherical but angular shard- and blade-like fragments of more felsic composition. The Toba-lake eruption of 75.000 years ago is probably the largest eruption in the last 5 million years. Tephra occur as far west as Sri Lanka, 2000 km from the caldera (Ninkovitch et al., 1978). ODP Site 758 penetrated the Toba tephra at Ninetyeast Ridge, but we found not a single spherule in any of the samples of the 25 cm thick tephra layer.

In conclusion, it is unlikely that the mafic KTB spherules can be explained as volcanic products, prasinophyte algal infilling, or as purely authigenic forms.

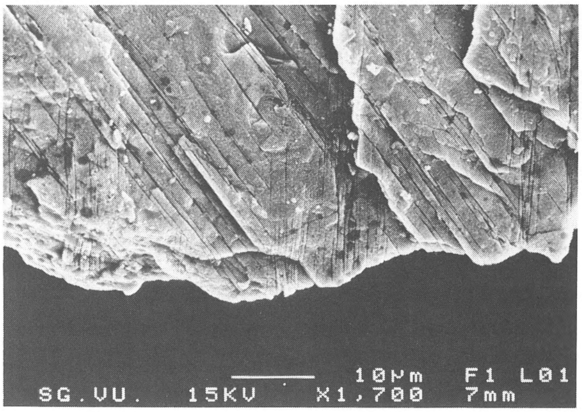

Fig. 11. SEM graph of a quartz crystal with 3 sets of crossing shack lamellae from the KTB of Brownie Butte, Montana (Bohor et al., 1984).

Shocked minerals

Bohor et al. (1984) reported felsic minerals, quartz, feldspar and quartzite with shock lamellae from the KTB clay in Montana (Fig. 11). Later these shocked minerals have been found in other sites in North America, but also in the fallout lamina at KTB in marine deposits, such as Caravaca, Agost and the Italian Appennines. The largest crystals occur in highest abundance in the western interior of the USA. Recently McHone et al. (1989) found stishovite at KTB, a high density polymorph of quartz which is stable only at high pressures (>90 kb) at low temperatures. Transport in a hot volcanic eruption column will result in re-alteration of the stishovite to quartz, and if the finding at KTB is confirmed, it forms compelling evidence for impact at KTB.

Multiple impacts

Shocked quartz and quartzite grains are difficult to reconcile with a mafic (= ocean floor) target site for the KTB impact which is indicated by the crystalline spherules, and the low-Rare Earth abundances (Fig. 1) in the fallout lamina (Smit & Romein, 1985). These felsic minerals must come from a continental crustal source. True microtektites

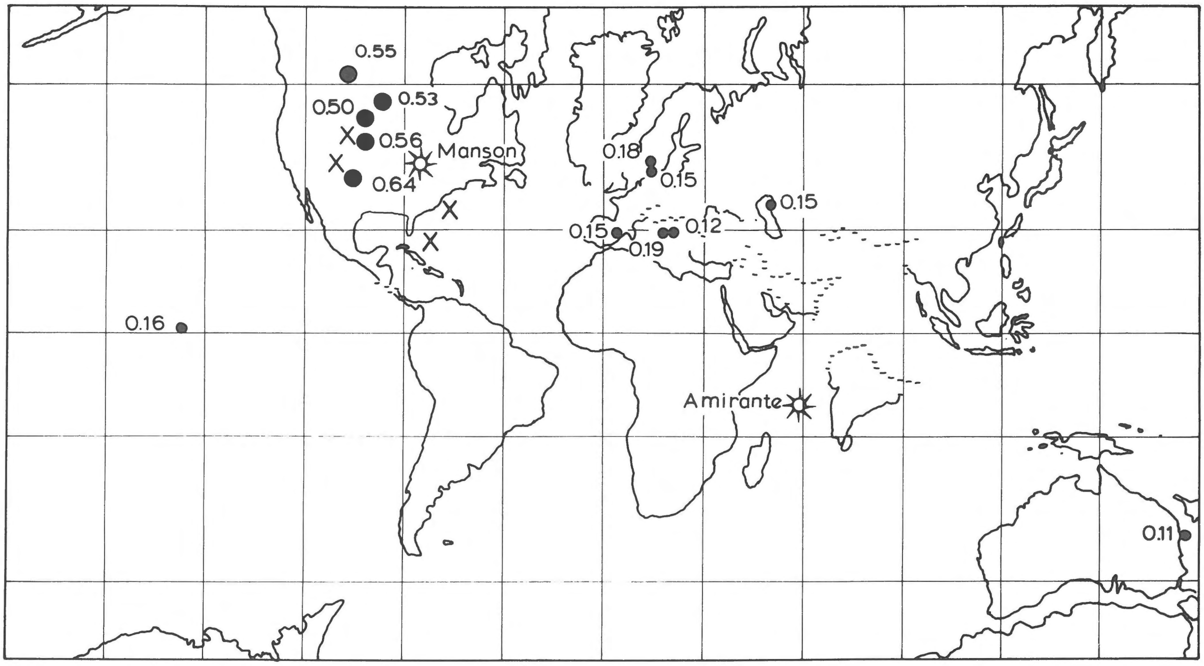

Fig. 12. Paleogeographic map (Early Paleocene) adapted from Izett (1986). The size of the dots reflects the abundance of shocked minerals. The maximum size of the shocked quartz is given in mm. Crosses indicate sites with large KTB microtektites.

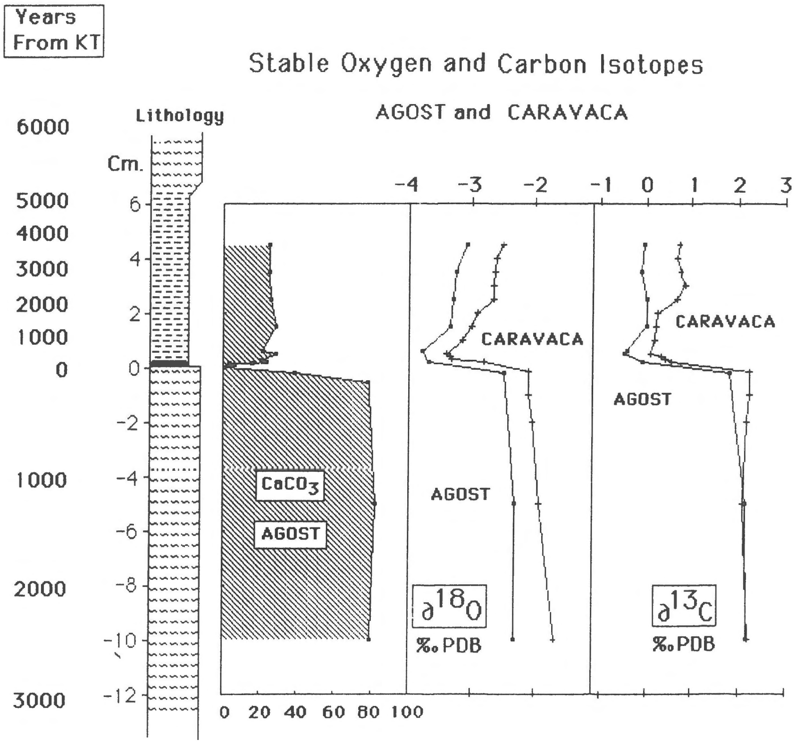

Fig. 13. Comparison of stable isotope values from bulk carbonate of the Agost and Caravaca sections. The vertical axis is linear with time. The consistent 0.5 per mil shift in δ 18 O between Caravaca and Agost may reflect the different paleoceanographic positions of the two sections.

originating from continental impacts are solidified as glassy tektites, without any crystallites within, as is the case in known microtektites from an impact on continental rocks (Glass, 1982). The larger spherules at KTB in North America may be altered glassy microtektites, because in none of these a relict crystalline texture has been found. The Manson crater is centered conveniently in the middle of this microtektite field (Figs 6, 12), and the largest shocked minerals. This crater may be responsible for the continental signature in the fallout lamina. Another impact must have taken place in the ocean, because of the huge amount of mafic spherules (total estimated mass of mafic material at KTB is about 1000 km^3). These impacts must have been almost simultaneous, because in Caravaca and Agost the shocked quartz occurs in the same 2 mm fallout layer at KTB as the mafic spherules.

Volcanic ejecta have still to be demonstrated in KTB sections. The Deccan trap eruptions bracket the KTB from the top of Chron 30N to the base of C29N, a time span of 0.5 to 1 million years. Their eruption rates are comparable to the present day outpouring of mafic lava at the mid-oceanic ridges combined, and as such are not sufficient to explain the mass-extinctions. Modelling suggest that the eruptions may have occurred in distinct pulses (Jaeger et al., 1989), but the dispersion of spherules and shocked minerals at KTB cannot be explained this way.

The KT boundary clay

The 6.5 cm boundary clay on top of the fallout lamina in both the Agost and Caravaca sections represents the almost empty oceans directly after the mass-extinctions. From sedimentation rates calculated for the Agost section (Groot et al., 1989) the duration of the deposition of the clay is estimat-

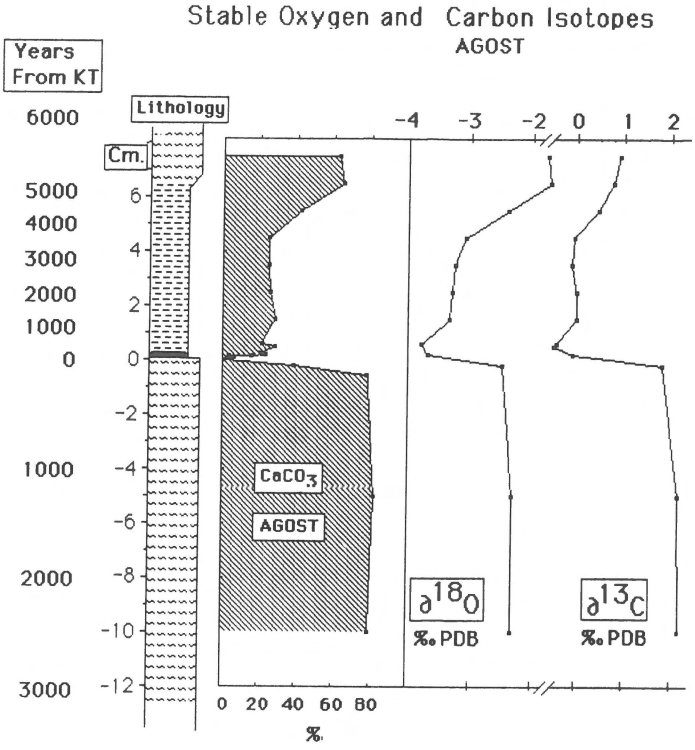

Fig. 14. Detailed δ13C and δ 18 O profiles from the KTB interval of the Agost section, Spain.

ed to be about 5–6 kyr. The boundary clay contains reworked and relict elements of the upper Maastrichtian planktonic fauna, and the new Paleocene planktonic forms appear all together in the gradual top of the clay in the Agost section. The new Paleocene planktonics are all new and show little resemblance to any of the Maastrichtian forms.

We may infer some of the paleoceanographic changes which occurred just after the impact and the mass-extinctions from stable isotope analysis of the carbonate of the boundary clay of Agost and Caravaca. Analyses were performed on the fine fraction and bulk carbonate, but the differences between those two are not significant. The isotopic curves for Caravaca and Agost are highly similar (Fig. 13) which gives us some confidence that isotopic variations are not caused by diagenesis.

Carbon isotopes

Stable carbon isotopic composition of pelagic carbonate reflects the productivity in the oceans. 12C is preferentally picked up from CO_2 of ocean surface water by photosynthesis and incorporated in organic tissue. Therefore CO_2 in surface water is systematically enriched in 13C opposite deeper water, with a healthy and producing photic layer. There is thus a gradient in the ocean from bottom to top which is reflected in isotopic measurements of planktic and benthic organisms, as foraminifers. Zachos et al. (1985) have shown that this gradient does not exist just after the KTB. The isotopic differences between the benthic *Aragonia* and planktonic foraminifers have disappeared, the ocean is homogenized with respect to δ 13C. This implicates that active photosynthesis in the photic zone was almost nil just after the impact event. This also means that we can safely conclude that *all* pelagic photosynthetic organisms, not just the coccolithophorids, were severely affected by the mass extinctions. Hsü characterizes such a non-productive ‘dead’ ocean as a ‘Strangelove ocean’ (Hsü, 1985).

Strangelove ocean conditions persisted later in the Paleocene, and we observe a return to a normal productive ocean on two timescales. In most KTB sections only the longer term return to full pre-KTB conditions is documented, on a timescale of 0.5 to 2 million years, depending probably on local conditions. The longer term lower productivity is reflected in the accumulation rates of pelagic carbonate in the Caravaca and Agost sections. These rates remain lower by a factor of 2 for about 2 million years. The short term productivity crisis has been demonstrated only in the Agost and Caravaca sections. Figure 14 shows a sudden decrease of up to 2.5 per mil in δ 13C in the first sediment on the fallout lamina. A first increase in δ 13C occurs some 0.5-1 kyr after the KTB, but during most of the deposition of the boundary clay (~5000 years) δ 13C (productivity) levels are still low. A second increase occurs near the top of the boundary clay, and is coincident with an increase in carbonate production. However, pre-extinction δ 13C levels are not yet regained.

Oxygen isotopes

Oxygen isotope fluctuations in pelagic carbonate at KTB probably reflect mainly temperature fluctuations. Ice-volume changes are probably not important, in the absence of large ice caps at the end of the Cretaceous. Vital effects are probably minor, because in species composition of nannofossils (the main carrier of the δ 18 O signal) the boundary clay is hardly different from the upper Maastrichtian. O'Keefe & Ahrens (1982) predicted on computer models of an oceanic impact that in the first few days after impact a significant cooling would occur, followed shortly by a temperature rise due to greenhouse effects. The cooling period is too short to be observed even in the high resolution sections of Agost and Caravaca.

The large shift of up to 2‰ in δ 18 O, also in the first sediment on the fallout layer, may indicate a temperature rise of 8° C of oceanic surface waters. Benthic forms do not seem to be affected. The low δ 18 O values (high temperatures) increase gradually through the boundary clay, to 'cool' down at the top of the boundary clay even more than in the top of the Cretaceous. This last 'cooling' may be masked by vital effects, because it is measured on an entirely new population of coccolithophorids and foraminifera.

Extinction scenario

The upper Maastrichtian was a period of stable climates, and little change. The turnover rate of planktonic foraminifers and other marine animals was slow. Stable isotope and trace element measurements do not indicate significant environmental changes before the KTB.

A number of simultaneous impacts occurred; one on land and maybe several in the ocean, leaving globally a thin ejecta layer with shocked minerals and microtektites. The resulting mass-extinctions define the end of the Mesozoic era. Dust clouds, some from extensive forest fires (Wolbach et al., 1986) may have suppressed sunlight for 1–3 months, reducing photosynthesis temporarily on a global scale. Mass-mortality, decay of dead organisms and the destruction of floral biomass by the extensive forest fires led to a large increase of CO_2 content in the atmosphere and the resulting greenhouse increased ocean surface temperatures by 8° C. These high values, at least at low latitudes, persist for 5000 years. When the excess of atmospheric CO_2 was reduced, we observe a cooling of surface temperatures, a partial restoration of photosynthesis in the photic layer of the oceans and restoration of production of pelagic carbonate. Only when these conditions were returned to pre-KTB values, a new pioneering fauna quickly colonized empty niches in the world oceans. In this pioneering period speciations occur frequently, the turnover rate of planktonic species was high and species abundances fluctuate highly. Soon afterwards, some 40.000 years later, the pioneering faunas were replaced by stable, well known planktonic faunas, characterized once again by slow turnover rates.

On the dinosaur front we observe similar patterns. Smit et al. (1987) demonstrated that the gradual extinction affair of dinosaurs well before

KTB is questionable, and that most of the speciation events of new eutherian mammals occurs not before, but just after the KT boundary.

For us, it is important to remember that the speciation event of the primitive primates occurs within the pioneering period just after the dinosaur extinctions.

Acknowledgements

I wish to thank G. Klaver, Th.B. Roep, W.T. ten Kate, T.M. van Kempen, S. Kars, A. Montanari, W. Alvarez, F.T. Kyte, J.T. Wasson, J.E. Bos, H.K.H. Holst, S. van der Kaars, J.J. Hermes, J.E. van Hinte, G. Brunsmann, F. Kievits, A. Prinsen, J.J. Groot, R. de Jonge and many others for their support for my research into the reasons behind the terminal Cretaceous mass-extinctions.

References

Alvarez, L.W., W. Alvarez, F. Asaro & H.V. Michel 1980 Extraterrestrial cause for the Cretaceous Tertiary extinction – Science 208: 1095–1108

Birkelund, T. & R.G. Bromley 1979 Maastrichtian and Danian of Denmark – Proc. Symp. Cretaceous-Tertiary Boundary events, Copenhagen. 1: 210 pp

Bohor, B.F., E.E. Foord, P.J. Modreski & D. Triplehorn 1984 Mineralogic evidence for an impact event at the Cretaceous Tertiary boundary – Science 224: 867–869

Bohor, B.F., D. Triplehorn, D.J. Nichols & H.T. Millard 1987 Dinosaurs, spherules and the 'magic' layer: A new K-T boundary site in Wyoming – Geology 15: 896–899

Brummer, G.J. & D. Kroon 1988 Planktonic foraminifers as tracers of ocean-climate history – Ph.D. Thesis, free University, Amsterdam: 346 pp

Courtillot, V., G. Feraud, H. Maluski, D. Vandamme, M.G. Moreau & J. Besse 1988 Deccan flood basalts and the Cretaceous Tertiary Boundary – Nature 333: 843–846

Fallot, P.M., M. Durand Delga, R. Busnardo & J. Siagal 1958 El Cretaceo superior del sur de Caravaca (Prov. de Murcia) – Notas. Com. Geol. Min. Espana. 50: 283–299

Glass, B.P. 1982 Possible correlations between tektite events and climate changes? – Geol. Soc. Am. Spec. Pap. 190: 251–256

Glass, B.P., C.A. Burns, J.R. Crosbie & D.L. Dubois 1985 Late Eocene North American microtektites and clinopyroxene-bearing spherules – Proc. 16th Lun. Planet. Conf. – J. Geophys. Res. 90: D175–D196

Groot, J.J., R.B.G. de Jonge, C.G. Langereis, W.G. ten Kate & J. Smit 1989 Magnetostratigraphy of the Cretaceous-Tertiary boundary at Agost Spain – Earth Planet. Sci. Lett. 94: 385–397

Hansen, T., R.B. Farrand, H.A. Montgomery, H.G. Billman & G. Blechschmidt 1985 Sedimentology and extinction patterns across the Cretaceous Tertiary boundary interval in East Texas – Cretaceous Research 8: 229–252

Hansen, H.J., R. Gwozdz, R.G. Bromley, K.L. Rasmussen, E.W. Vogesen & K.R. Pedersen 1986 Cretaceous/Tertiary spherules from Denmark, New Zealand and Spain – Geol. Soc. Denm. Bull. 35: 75–82

Hartnady, C.J.H. 1986 Amirante basin, western Indian Ocean: Possible impact site of the Cretaceous/Tertiary extinction bolide? – Geology 14: 423–426

Hsü, K.J. & J.A. McKenzie 1985 A 'Strangelove' Ocean in the earliest Tertiary – Am. Geoph. Union, Geoph. Monographs 32: 487–492

Izett, G.A. 1987 Authigenic 'spherules' in K-T boundary sediments at Caravaca, Spain and Raton basin, Colorado and New Mexico may not be impact derived – Geol. Soc. Am. Bull. 99: 78–86

Jaeger, J.J., V. Courtillot & P. Tapponnier 1989 A paleontological view on the age of the Deccan traps, of the Cretaceous Tertiary boundary and of the India-Asia collision – Geology 17: 89–93

Keller, G. 1988 Extinction, survivorship and evolution of planktic forminifera across the Cretaceous Tertiary boundary at El Kef, Tunisia – Mar. Micropal. 13: 239–263

Klaver, G.Th. 1987 The Curacao lava formation: an ophiolitic analogue of the anomalous thick layer 2B of the Mid-Cretaceous oceanic plateaus in the western Pacific and central Caribbean – GUA Pap. Geol. Ser. 1, 27: 169 pp

Klaver, G.T., T.M.G. van Kempen, F.R. Bianchi & S.J. van der Gaast 1987 Green spherules as indicators of the Cretaceous oceanic Tertiary boundary in Deep Sea Drilling Project Hole 603b – Init. Rep. Deep Sea Drill. Proj. 92: 1039–1055

Kunk, M.J., G.A. Izett, R.A. Haugerud & J.F. Sutter 1989 40Ar-39Ar dating of the Manson Impact structure: A Cretaceous Tertiary boundary Crater candidate – Science 244: 1565–1568

Lofgren, G.E. 1980 Experimental studies on the dynamic crystallization of silicate melts – In: R.B. Hargraves (ed.): Physics of magmatic processes, Princeton Univ. Press: 487–551

Luterbacher, H.P. & I. Premoli Silva 1964 Biostratigrafia del Limite Cretaceo-Terziario nell Appennino centrale – Riv. Ital. Pal. 72: 67–128

Magaritz, M., S. Moshkovitz, C. Benjamini, H.J. Hansen, E. Hakansson & K.L. Rasmussen 1985 Carbon isotope-, bio- and magnetostratigraphy across the Cretaceous-Tertiary boundary in the Zin Valley, Negev, Israel – Newslett. Strat. 15: 100–113

McHone, J.F., R.A. Nieman, C.F. Lewis & A.M. Yates 1989 Stishovite at the Cretaceous Tertiary boundary Raton, New Mexico – Science 243: 1182–1184

Naslund H.R., C.B. Officer & G.D. Johnson 1986 Microspherules in upper Cretaceous and Lower Tertiary clay layers at Gubbio, Italy – Geology 14: 923–926

Ninkovitch, D., R.S.J. Sparks & M.T. Ledbetter 1978 The exceptional magnitude and intensity of the Toba Eruption, Sumatra: an example of the use of deep-sea tephra layers as a geological tool – Bull. Volcanol. 41: 286–297

Officer, C.B. & D.R. Lynch 1983 Determination of mixing parameters from tracer distributions in deep-sea sediment coves – Mar. Geology 52: 59–74

O'Keefe, J.D. & T.J. Ahrens 1982 The interaction of the Cretaceous Tertiary extinction bolide with the atmosphere – Geol. Soc. Am. Spec. Pap. 190: 103–120

Rigby J. Keith Jr., K.R. Newman, J. Smit, S. van der Kaars, R.E. Sloan & J. Keith Rigby Sr. 1987 Dinosaurs from the Paleocene part of the Hell Creek Formation, McCone county, Montana – Palaios 1: 295–302

Schmidt, R.M. & K.A. Holsapple 1982 Estimates of crater size for large-body impact: gravity-scaling results – Geol. Soc. Am. Spec. Pap. 190: 93–102

Smit, J. & J. Hertogen 1980 An extraterrestrial event at the Cretaceous Tertiary boundary – Nature 285: 198–200

Smit J. & F.T. Kyte 1983 Siderophile-rich magnetic spheroids from the Cretaceous Tertiary boundary in Umbria, Italy – Nature 310: 403–405

Smit, J. & A.J.T. Romein 1985 A sequence of events across the Cretaceous-Tertiary boundary – Earth Planet. Sci. Lett. 74: 155–170

Smit, J. & W.G. ten Kate 1982 Trace element patterns at the Cretaceous Tertiary boundary; consequences of a large impact – Cretaceous Research 3: 307–332

Smit, J., W.A. van der Kaars & J. Rigby Keith Jr. 1987 Stratigraphic aspects of the Cretaceous Tertiary boundary in the Bug Creek area of eastern Montana. Mém. Soc. Géol. Fr. N.S. 150: 53–73

Smith, A.G. & J.C. Briden 1977 Mesozoic and Cenozoic Paleocontinental maps – Cambridge Univ. Press: 102 pp

Stokes, R.B. 1979 An analysis of the ranges of spatangoid echinoid genera and their bearing on the Cretaceous/Tertiary boundary – Proc. Symp. Cretaceous-Tertiary Boundary events, Copenhagen 2: 78–82

Troelstra, S.R. 1989 Studies in late Cenozoic extant and extinct foraminifers – Ph.D. Thesis, Free University, Amsterdam: 275 pp

Van Veen, G.W. 1969 Geological investigations in the region West of Caravaca, South eastern Spain – Ph.D. Thesis, University of Amsterdam: 143 pp

Ward, P. 1988 Maastrichtian ammonite and inoceramid ranges from the Bay of Biscay Cretaceous Tertiary boundary sections – Rev. Esp. Paleontol. Num. Extraord. 119–126 190: 93–102

Wolbach, W.S., I. Gilmour, E. Anders, C.J. Orth & R.R. Brooks 1986 Global wildfire at the cretaceous Tertiary boundary – Nature 334: 665–667

Zachos, J.C., M.A. Arthur, R.C. Thunell. D.F. Williams & E.J. Tappa 1985 Stable isotope and trace element geochemistry of carbonate sediments across the Cretaceous/Tertiary boundary at Deep Sea Drilling Project Hole 577, Leg 86 – Init. Rep. Deep Sea Drill. Proj. 86: 513–523