

Cretaceous climate, volcanism, impacts, and biotic effects

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ABSTRACT

Cretaceous volcanic activities (LIPs and CFBPs) appear to have had relatively minor biotic effects, at least at the generic level. Major biotic stress during the Cretaceous was associated with OAEs and related to nutrient availability largely from weathering, greenhouse warming, drowning of platform areas, and volcanism. The biotic effects of OAEs were often dramatic at the species level, causing the extinction of larger specialized and heavily calcified planktonic foraminifera (rotaliporid extinction) and nannoconids (nannoconid crises), the temporary disappearances of other larger species, and the rapid increase in *r*-selected small and thin-walled species, such as the low oxygen tolerant heterohelicids and radially elongated taxa among planktic foraminifera and thin walled nanofossils. Biotic diversity increased during cool climates, particularly during the late Campanian and Maastrichtian, reaching maximum diversity during the middle Maastrichtian. High biotic stress conditions began during greenhouse warming and Deccan volcanism about 400 ky before the K-T boundary; it reduced abundances of large specialized tropical planktic foraminiferal species and endangered their survival. By K-T time, renewed Deccan volcanism combined with a large impact probably triggered the demise of this already extinction prone species group.

Evidence from NE Mexico, Texas, and the Chicxulub crater itself indicates that this 170 km-diameter crater predates the K-T boundary by ~300,000 years and caused no species extinctions. The Chicxulub impact, therefore, can no longer be considered a direct cause for the K-T mass extinction. However, the K-T mass extinction is closely associated with a global Ir anomaly, which is considered too large, too widespread, and too concentrated in a thin layer to have originated from volcanic activity, leaving another large impact as the most likely source. This suggests that a second still unknown larger impact may have triggered the K-T mass extinction.

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1. Introduction

The Cretaceous is a complex time period in Earth's history that saw episodes of abrupt greenhouse warming and cooling (Lini et al., 1992; Föllmi, 1995; Huber et al., 1995; Stoll and Schrag, 1996; Larson and Erba, 1999; Jenkyns, 2003; Wilson et al., 2002), repeated platform drowning, major excursions in the carbon isotope record, episodic oceanic anoxic events, and increased volcanic activity forming large igneous provinces (LIPs) (Föllmi et al., 1994; Weissert et al., 1998; Jenkyns et al., 1994; Hochuli et al., 1999; Clarke and Jenkyns, 1999; Wissler et al., 2003; Bodin et al., 2006). By the late Cenomanian to early Turonian, sea levels peaked at about 300 m above present, inundating continental shelf areas and forming extensive inland seas that favored deposition and preservation of chalk deposits (Hallam and Wignall, 1997). Widespread oceanic anoxic events accompanied the sea level transgressions and climate changes (e.g., Valanginian, Hauterivian, Barremian/Aptian, Aptian/

Albian, Albian/Cenomanian, Cenomanian-Turonian, and Coniacian) leading to high biotic stress conditions largely due to high nutrient levels associated with greenhouse warming, increased weathering, and LIPs (e.g., Erba, 1994, 2004; Galeotti, 1998; Kerr, 1998; Cobianchi et al., 1999; Riebesell et al., 2000; Keller et al., 2001, 2004a,b; Wignall, 2001; Hart et al., 2002; Courtillot and Renne, 2003; Cocconi and Luciani, 2004). But the expansive shallow seas also provided new habitats for shallow water marine and terrestrial life, leading to diversification of coccoliths, diatoms, foraminifera, ammonoids, mollusks, angiosperms, gymnosperms, dinosaurs, plesiosaurs, pterosaurs, and ichthyosaurs. By the end of the Cretaceous, the coincidence of volcanism, impacts, and climate change likely caused one of Earth's major mass extinctions.

This report focuses on Cretaceous biodiversity and extinction trends and examines their potential cause-and-effect relationships between volcanism (LIPs and CFBPs), impacts, climate, ocean anoxia, and the carbon cycle. Specific topics addressed include the Cretaceous-Tertiary (K-T) boundary mass extinction, the age of the Chicxulub impact, biotic effects of volcanism, and the biotic response to oceanic anoxic events. The report is based on the

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keynote lecture delivered in 2005 at the 7th International Symposium on the Cretaceous in Neuchatel, Switzerland. I thank the organizers, Karl Follmi and Thierry Adatte, for a magnificent meeting.

2. What constitutes a mass extinction?

In general we can assume that the disappearance of >40% of generic diversity qualifies as a major mass extinction. By this measure the five Phanerozoic mass extinctions, based on Sepkoski's (1996) genus level database, are easily recognized and corroborated: the end-Cretaceous, end-Permian, mid-Permian (Maokouan), end-Devonian and end-Ordovician (Fig. 1). The middle and late Cambrian extinction events reach just above 40%, but it is still uncertain whether these are regional or global events. Minor extinctions are more problematical in part due to the use of stage-level data, which means that any genera that disappeared in a given stage are grouped as one sudden extinction event, when in reality they may have been spread out over 6–12 m.y. and constitute background or accelerated extinctions. Stage-level extinction data must therefore be interpreted with caution.

The Phanerozoic extinction record suggests that a 10% generic turnover per stage can be considered as overall background extinctions, though the rate is significantly lower during the Cretaceous and even lower during the Cenozoic (Fig. 1). Indeed the Cretaceous marks a very long period (145–65 Ma) of background extinctions (<10%), except for elevated extinctions during the oceanic anoxic events (OAE) at the Aptian/Albian and Cenomanian/Turonian (17%), and of course the end-Cretaceous mass extinction. Elevated extinctions of between 10–20% occurred during all time periods, with the exception of the Cenozoic. Accelerated extinctions, here defined as between 20–30%/stage, commonly precede and follow mass extinctions, particularly at the end of the Triassic and late Devonian. During the middle and late Cambrian and late Triassic stage-level extinctions between 30–40% mark minor mass extinctions. Thus, within the context of Phanerozoic genus-level extinctions, the Cretaceous marks a period of low biotic stress followed by the K-T mass extinction. Genus-level diversity trends (Sepkoski, 1997; MacLeod, 2004) for the Cretaceous and Tertiary support this observation. During the Cretaceous there is a steady increase in diversity, leveling off during the Turonian to Santonian followed by a rapid increase in overall biodiversity

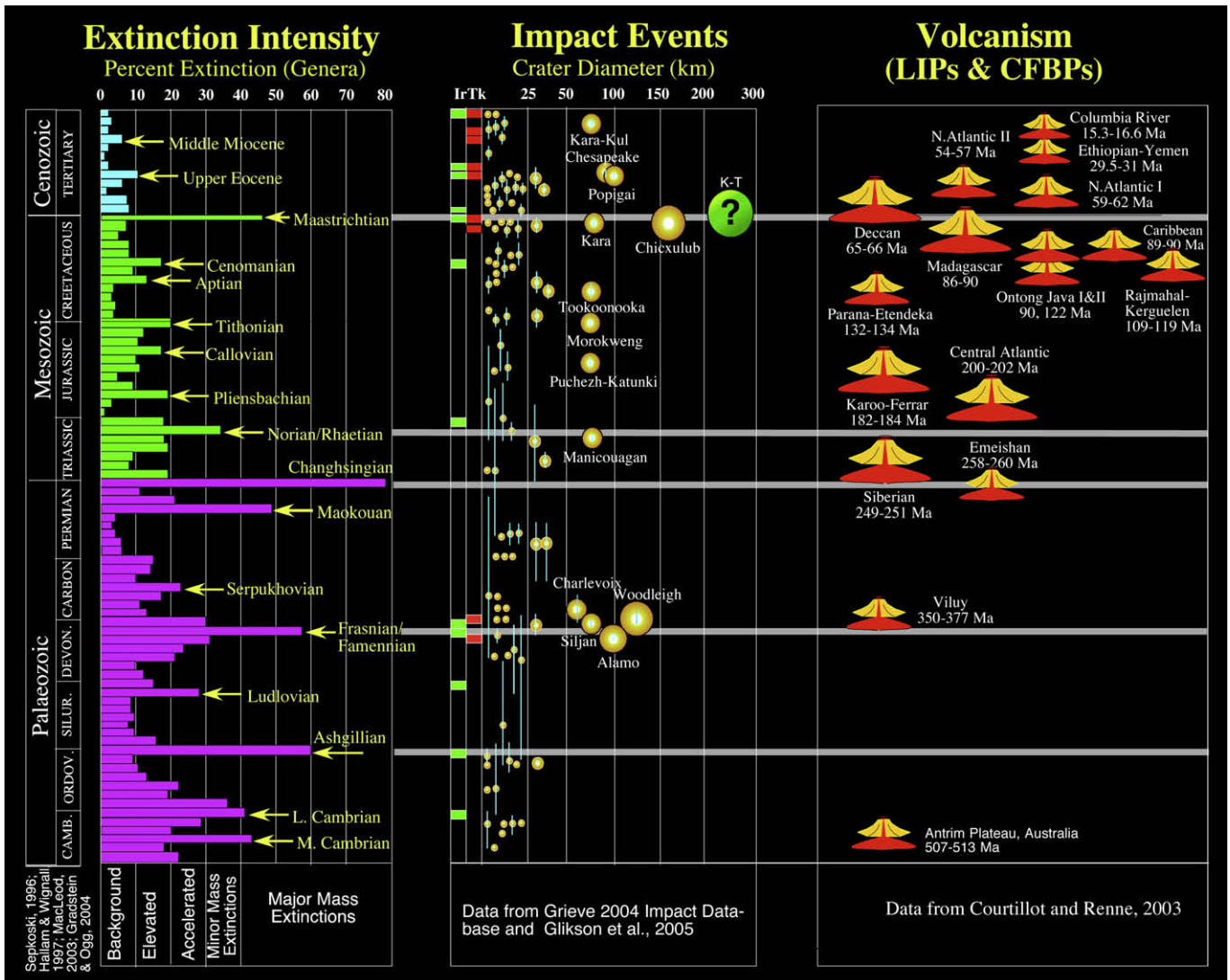


Fig. 1. Mass extinctions, impacts and large igneous provinces during the Phanerozoic. Stratigraphic subdivisions and numerical ages from the 2004 International Stratigraphy Chart (ICS) of Gradstein and Ogg (2004). Note that the Chicxulub impact predates the K-T boundary by 300 ky and a second larger (>200 km) impact crater is postulated for the K-T boundary based on the Ir distribution pattern and mass extinction (Modified after Keller, 2005a).

during the Campanian and Maastrichtian (Fig. 2). This genus-level diversity trend, however, may significantly underestimate the environmental stress associated with volcanism, impacts, climate and sea level changes.

3. Are impacts to blame for mass extinctions?

Common sense tells us that very large impacts could cause catastrophic mass extinctions. But how large must an impact be to cause a mass extinction? The Phanerozoic impact database (Earth Impact Database 2005) notes a large number of craters 5–49 km in diameter, but only 11 between 50–100 km, one possibly 120 km (Glikson et al., 2005) and Chicxulub at 170 km the largest known impact crater (Fig. 1). Thus, during the Phanerozoic only a few impacts are large enough to have the potential to cause significant biotic stress (Table 1). Of course, this database is necessarily incomplete because oceanic impact craters are generally not preserved due to seafloor subduction, whereas preservation of impact craters on continents depends on tectonic activity, erosion and vegetation cover. Nevertheless, the impact record provides sufficient information to draw some conclusions as to the potential kill-effect of large impacts.

A plot of Phanerozoic mass extinction intensity versus impact crater size shows no direct correlation, except for the potential close association of the Chicxulub impact with the K-T mass extinction (Courtilot, 1999; Wignall, 2001; MacLeod, 2003; Keller, 2005a; White and Saunders, 2005). However, to really determine the impact kill-effect we must first determine the precise ages of impacts. Most impact craters are difficult to date even within ± 0.5 to ± 2 million years (error bars in Fig. 1), which eliminates most craters from useful correlation with extinction horizons. To prove the kill-effect of an impact requires close age correlation between the impact and extinction horizons, preferably in the same stratigraphic sequence. Such dated sequences exist for the two large (90–100 km) late Eocene Popigai and Chesapeake impacts (Montanari and Koeberl, 2000; Poag et al., 2002) and the Chicxulub impact (Stinnesbeck et al., 2001; Keller et al., 2003a, 2004a,b).

The late Eocene impact craters are not associated with any species extinctions (Keller et al., 1983; Keller, 1986; Montanari and Koeberl, 2000; Monechi et al., 2000). Nevertheless, Poag (1997) and Alvarez (2003) argued that an impact-cause is “abundantly confirmed” by the minor and gradual extinctions over the following 3 m.y. interval of the late Eocene, which they attributed to the delayed adverse effects of the impacts causing long-term global warming. These interpretations are in direct contradiction with climate studies, which show neither short-term nor long-term changes associated with the impact horizons and global cooling, not warming, during the late Eocene (Montanari and Koeberl, 2000; Bodiselitsch et al., 2004; Keller, 2005a). Moreover, generic diversity between middle and late Eocene rapidly increased, whereas extinction intensity increased only from 6% to 10%, remaining within the level of background extinctions (Fig. 2). We can therefore conclude that impacts with crater sizes up to 100 km cause no significant biotic or environmental effects (Fig. 3). This leaves only the Chicxulub impact as a potential impact large enough to cause a major mass extinction, but a K-T age for this impact is in question.

4. Chicxulub impact predates K-T by ~300 ky

The Chicxulub impact is commonly believed to have caused the K-T mass extinction. But recent evidence puts this conventional wisdom increasingly in doubt in the most continuous and expanded K-T sequences with impact ejecta layers in NE Mexico, Texas and the Chicxulub impact crater (Fig. 4). In these sequences the Chicxulub ejecta spherule and breccia layers are stratigraphically well below the K-T boundary in late Maastrichtian pelagic sediments that predate the boundary by about 300,000 years (Keller et al., 2002, 2003a, 2004a,b). The evidence is briefly summarized here.

4.1. Northeastern Mexico

The commonly cited K-T age for the 170 km Chicxulub impact crater on Yucatan, Mexico, is based on: (1) $^{39}\text{Ar}/^{40}\text{Ar}$ ages of impact

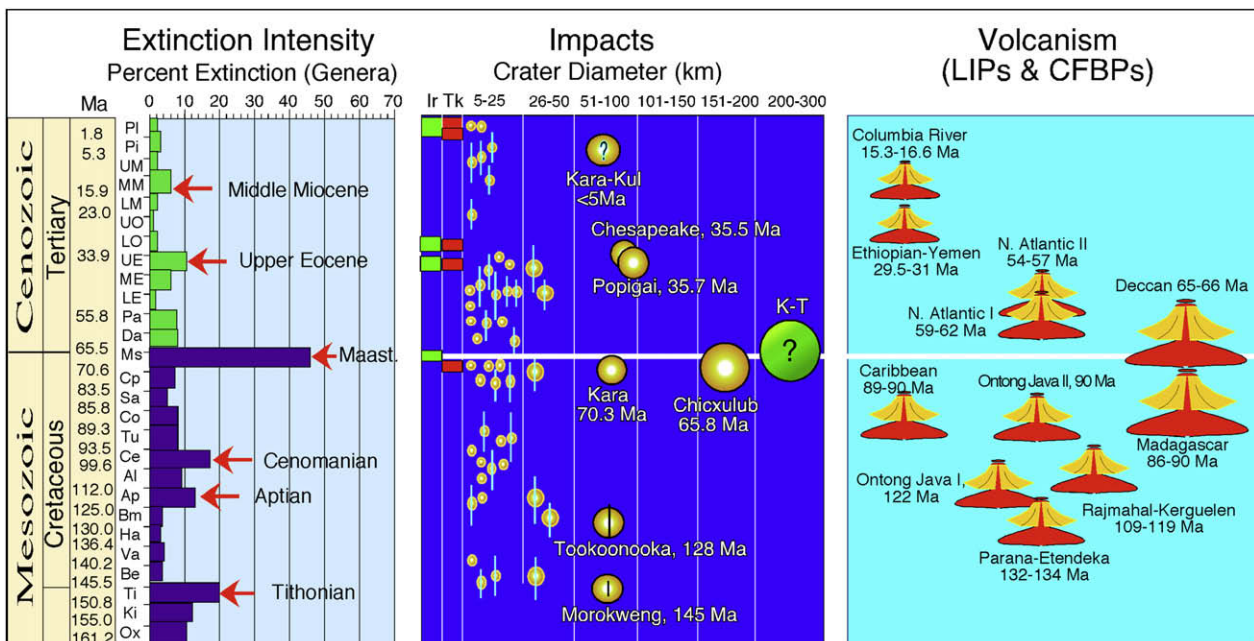


Fig. 2. Mass extinctions, impacts and large igneous provinces during the Cretaceous and Cenozoic. Diversity and extinction data (genera) from Sepkoski (1996, 1997). Time scale after Gradstein and Ogg (2004); impacts from Grieve database and volcanic provinces after Courtilot and Renne (2003). Note that the Chicxulub impact predates the K-T boundary by 300 ky and a second larger (>200 km) impact crater is postulated for the K-T boundary based on the Ir distribution pattern and mass extinction.

Table 1

Jurassic, Cretaceous, and late Eocene impact craters. Chicxulub at 170 km-in diameter is the largest known Phanerozoic impact crater, followed by the late Devonian Woodleigh (120 km) and late Eocene Popigai and Chesapeake craters (90–100 km). The Chicxulub impact predates the K-T by ~300 ky and caused no extinctions. The global K-T Ir anomaly suggests that a second larger impact may have triggered the end-Cretaceous mass extinction

Crater	Location	Latitude	Longitude	Diameter (Km)	Age (Ma)	Period	Extinct
Popigai	Russia	N 71° 39'	E 111° 11'	100	35.7 ± 0.2	Late Eocene	~10%
Chesapeake	Virginia, USA	N 37° 17'	W 76° 1'	90	35.3 ± 0.3	Eocene	~10%
Mistastin	NF, Canada	N 55° 53'	W 63° 18'	28	36.4 ± 4		~10%
KT impact	unknown	unknown	unknown	>200?	65.5	K-T	<45%
Boltysh	Ukraine	N 48° 45'	E 32° 10'	24	65.67 ± 64	Late Maast.	<5%
CHICXULUB	Yucatan, Mexico	N 21° 20'	E 89° 30'	170	65.8 ± 0.1	Maast.	<5%
Kara	Russia	N 69° 6'	E 64° 9'	65	70.3 ± 2.2	E. Maast	~8%
Lappajarvi	Finland	N 63° 9'	E 23° 42'	23	73.3 ± 5.3	Late Campanian	~8%
Manson	Iowa, USA	N 42° 35'	W 94° 33'	35	73.8 ± 0.3	Campanian	~8%
Steen River	Canada	N 59° 30'	W 117° 38'	25	91 ± 7	E. Turonian	~8%
Carswell	Canada	N 58° 27'	W 109° 30'	39	115.0 ± 10	Aptian	~12%
Tookoonooka	Australia	S 27° 7'	E 142° 50'	55	128.0 ± 5	Barremian	<5%
Mjolnir	Norway	N 73° 48'	E 29° 40'	40	142.0 ± 2.6	Berriasian	<5%
Gosses Bluff	NT Australia	S 23° 49'	E 132° 19'	22	142.5 ± 0.8	Berriasian	<5%
Morokweng	South Africa	S 26° 28'	E 23° 32'	70	145.0 ± 0.8	Tithonian	~20%
Puchezh-Katunki	Russia	N 56° 58'	E 43° 43'	80	167 ± 3	Jurassic	<10%
Obolon'	Ukraine	N 49° 35'	E 32° 55'	20	169 ± 7	Jurassic	<10%

glass at 65.0, 65.2 or 65.4 ± 0.2 Ma (Izett et al., 1991; Sharpton et al., 1992; Swisher et al., 1992). (Note, the K-T boundary has been revised from 65 Ma to 65.5 Ma by Gradstein and Ogg, 2004.) However, radiometric dating is not precise enough to distinguish whether the impact occurred at the K-T or within a few hundred thousand years of it. (2) The presence of impact ejecta spherules at the base of the 2 m to 8 m thick sandstone-siltstone units that infill submarine channels at paleodepths of 500–1000 m and underlie the K-T boundary mass extinction and Ir anomaly throughout NE Mexico. To explain the widely separated impact glass spherule layers and Ir anomaly, the sandstone-siltstone units have been interpreted as impact-generated tsunami deposit with the Ir settling at the top within hours to a few days (Smit et al., 1992, 1996; Smit, 1999; Soria et al., 2001).

From the very beginning there were many problems with this interpretation (see review in Keller et al., 2003a). Problems include: (1) Presence of repeated burrowed (*Thalassinoides*, *Chondrites*) horizons of fine-grained sediments, truncated by the overlying

sandstones. This indicates that the ocean floor was repeatedly colonized, then buried by slumps or gravity flows, which is incompatible with deposition within hours to days via tsunami (Adatte et al., 1996; Keller et al., 1997; Ekdale and Stinnesbeck, 1998). (2) The impact spherules at the base of the channel deposits are reworked, as indicated by the presence of shallow water benthic foraminifera, wood and plant debris (Keller et al., 1994; Alegret et al., 2001). Thus, spherules were originally deposited in a shallow nearshore area and subsequently eroded and transported into the deep submarine channels. (3) There are two such spherule-erosion and redeposition events separated by a 20–25 cm thick sandy limestone, which contains J-shaped spherule-filled burrows truncated at the top (Fig. 5). This means that after deposition of the limestone layer, animals colonized and burrowed on the seafloor prior to deposition of the second spherule layer. (4) The mass extinction and Ir anomaly coincides with a thin red clay layer above the channel-fill deposits and cannot be temporally tied to the two reworked spherule layers below.

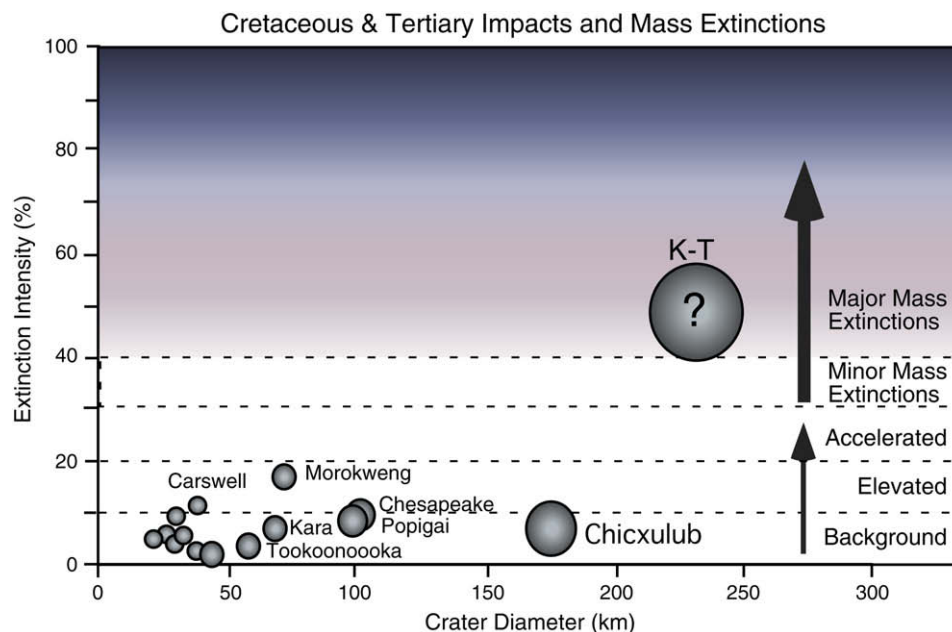


Fig. 3. A plot of extinction intensity versus crater diameter of known impacts reveals no correlation between mass extinctions, including the Chicxulub impact at 170 km-in-diameter. Note that the Chicxulub impact crater predates the K/T boundary and is not associated with the mass extinction. A K/T impact crater of >200 km-in-diameter is inferred based on this plot and the global iridium anomaly at the K/T mass extinction. (Modified after Keller, 2005a).

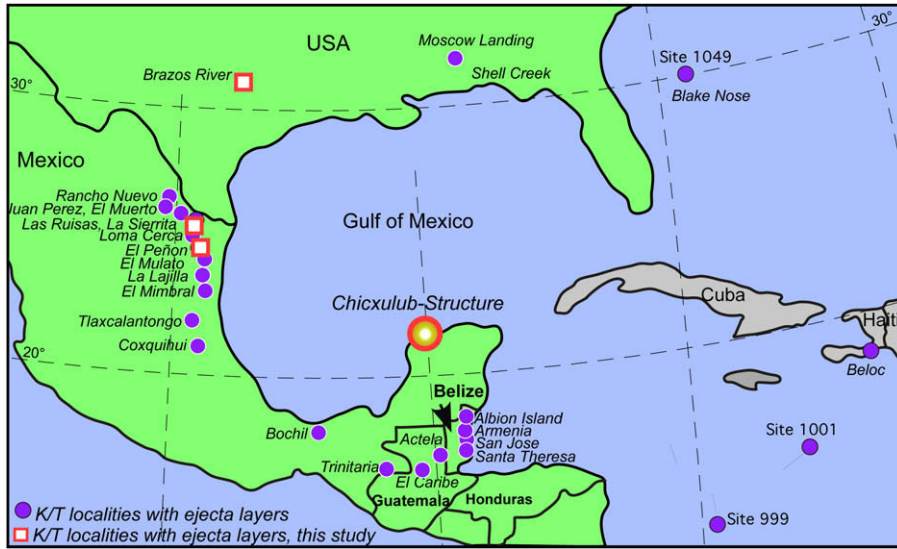


Fig. 4. Localities with Cretaceous-Tertiary boundary sequences in Central America and the southern USA that contain Chicxulub impact glass spherule ejecta.

If the spherule layers at the base of the submarine channel deposits are reworked, then finding the original spherule ejecta layer in its stratigraphic position would provide proof of the real age of the Chicxulub impact. In the late 1990's a systematic search was undertaken by a group of Masters and graduate students from Neuchatel, Karlsruhe and Princeton Universities and their advisors Thierry Adatte, Wolfgang Stinnesbeck, Doris Stueben and myself. Over an area spanning 80 km² and over 40 outcrops in NE Mexico, the late Maastrichtian pelagic sediments below the channelized sandstone-siltstone units were systematically examined (e.g.,

Lindenmaier, 1999; Schulte, 1999; Affolter, 2000; Schilli, 2000). Most outcrops revealed additional reworked spherule layers interbedded in late Maastrichtian sediments. At Loma Cerca, the oldest spherule layer was found more than 5 m below the channel deposits. During a fieldtrip in 2002, a group of Princeton undergraduate students discovered a 1.8 m thick spherule layer nearly 5 m below the reworked spherule layers at the base of the channel-fill deposits at El Penon (Fig. 6). As at Loma Cerca 30 km to the northwest, this spherule layer contains no shallow water debris and the basal layer consists of almost pure impact glass spherules

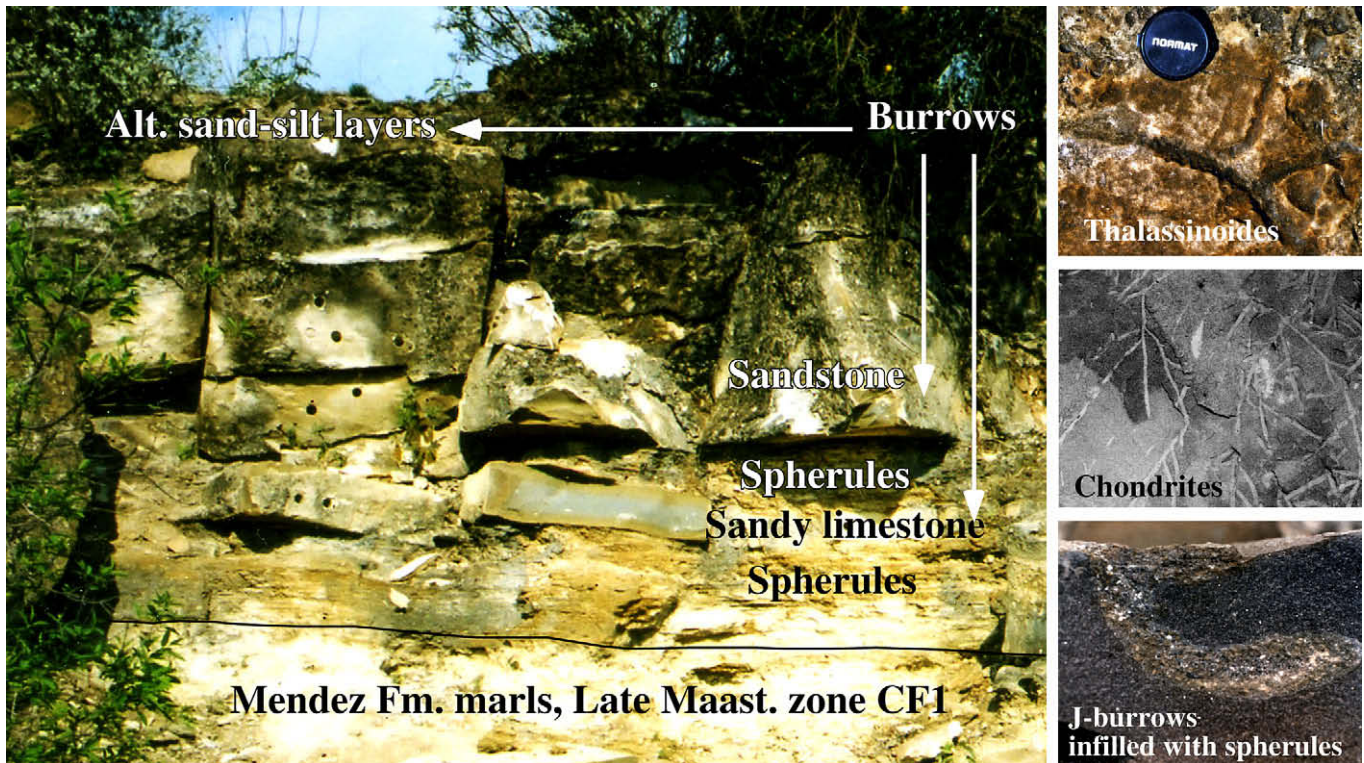


Fig. 5. The sandstone-siltstone complex at El Penon, NE Mexico, overlies two reworked spherule layers that are separated by a 20 cm thick sandy limestone. J-shaped burrows infilled with spherules and truncated at the top are present in the limestone and overlying sandstone. Siltstone and marly layers above the sandstone are intensely burrowed by *Chondrites* and *Thalassinoides*. These are not characteristics that would be found in a tsunami deposit; they indicate deposition over an extended time period during which the ocean floor was repeatedly colonized by invertebrates. The K-T boundary is eroded at the top of the section.

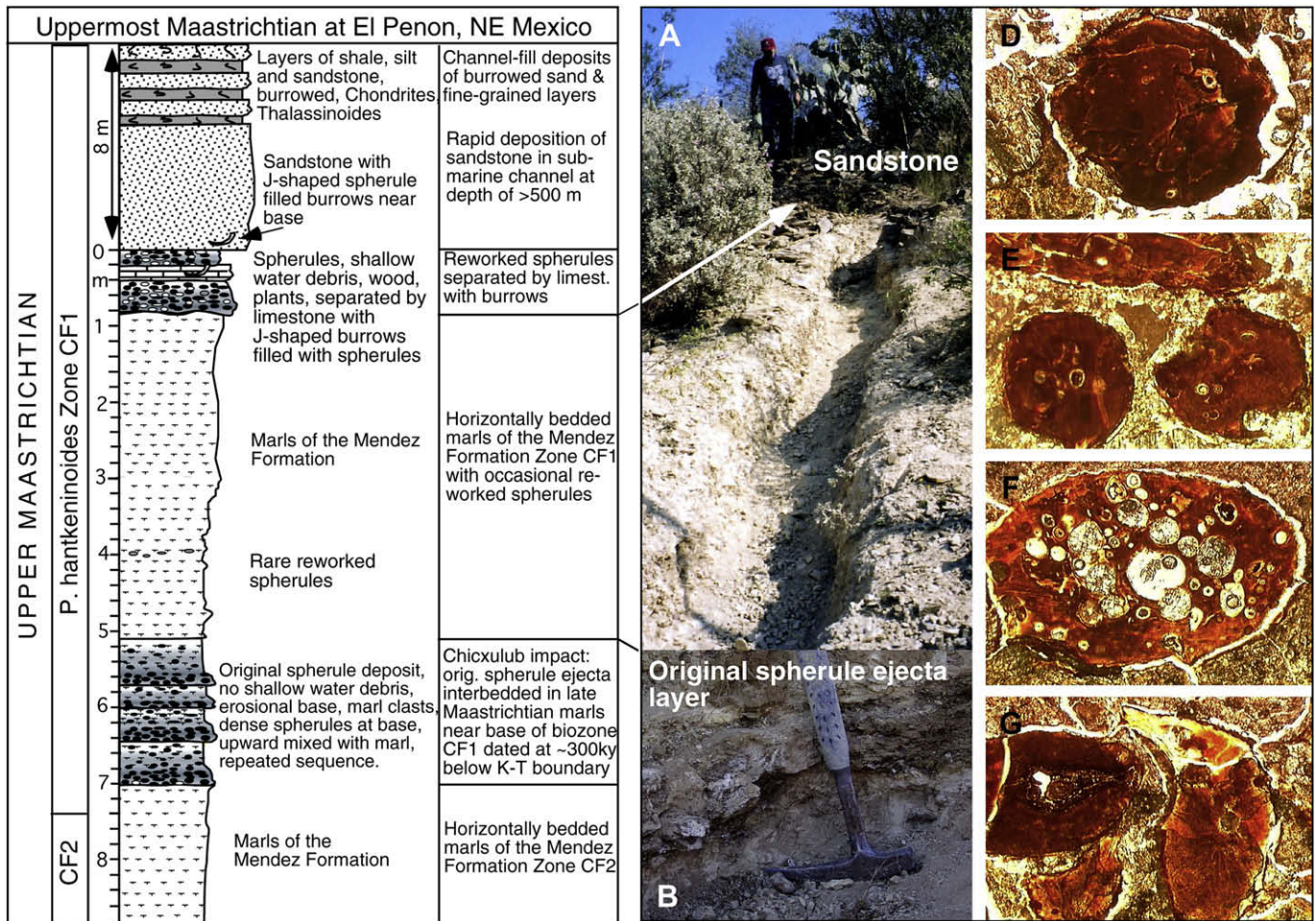


Fig. 6. Late Maastrichtian sequence at El Penon, NE Mexico, with the original Chicxulub impact spherule deposit near the base of zone CF1, 4 m below the two reworked spherules layers and sandstone-siltstone complex. A, Trenched sequence from sandstone complex to original spherule ejecta layer shows undisturbed pelagic marls. B, Base of the 1.8 m thick original spherule ejecta layer. D–G) Spherules from original spherule ejecta layer range from 2–5 mm in size.

with interstitial calcite cement, which strongly indicates that this is the original spherule ejecta deposit that was subsequently reworked and redeposited at the base of the submarine channel, probably during the latest Maastrichtian sea level lowstand (Adatte et al., 1996). Biostratigraphy places this spherule layer at Loma Cerca and El Penon in the lower part of Chron 29r, the lower half of the nannofossil *Micula prinsii* zone, and at the base of planktic foraminiferal zone CF1 (*Plummerita hantkeninoides*), which spans the last 300 ky of the Maastrichtian. This data indicates that the Chicxulub impact predates the K-T boundary by about 300 ky (Keller et al., 2002, 2003a,b).

Critics countered that any spherule layer below the “tsunami” deposit must be due to slumps induced by the Chicxulub impact (Soria et al., 2001; Smit et al., 2004). However, all across NE Mexico there is no significant disturbance of the pelagic late Maastrichtian sedimentary layers and there is no evidence of significant slumping. In over 40 outcrops only two minor (<2 m in dimension) folds related to internal local deformation have been documented in one area (Mesa Juan) (e.g., Soria et al., 2001; Schulte et al., 2003; Harting, 2004).

4.2. Chicxulub impact crater

In 2001–2002 the International Continental Scientific Drilling Program (ICDP) supported the drilling of a new core within the Chicxulub crater with a major objective to prove once and for all

that the Chicxulub impact is the long-sought K-T impact and cause for the mass extinction (Dressler et al., 2003). Instead, it caused renewed controversy. The new Yaxcopoil-1 core recovered a 100 m thick impact breccia, which is separated from the K-T boundary by a 50 cm thick micritic limestone. Thus, the Yaxcopoil-1 core reveals the same temporally separated impact deposits from the K-T boundary as observed in NE Mexico. This limestone was immediately interpreted as high-energy backwash and crater infill after the impact (Dressler et al., 2003; Smit et al., 2004). As with the tsunami interpretation in NE Mexico sections, there are many obvious problems with this backwash-crater fill interpretation.

- (1) The micritic limestone is finely layered with patches of anhedral dolomite crystals, which indicates deposition under low-energy, quiet-water conditions where dolomite formed by diagenetic replacement of the precursor limestone (Keller et al., 2004a,b). Only two 1 cm layers of oblique bedding suggest occasionally slightly agitated water (Fig. 7). There is no evidence of breccia clasts from the underlying impact breccia or surrounding crater walls, as would be expected if this were a high-energy backwash and crater infill deposit.
- (2) The presence of five thin green glauconite layers (Fig. 7) effectively rules out backwash and crater infill (Keller et al., 2004a,b). Glauconite forms at the sediment-water interface in environments with very low detritus accumulation (Chamley, 1989). Each layer represents tens of thousands of years of

deposition. Each layer is bioturbated, which indicates colonization of the ocean floor during deposition. Thus, sediment deposition occurred over a very long time period. No evidence of altered impact glass (Smit et al., 2004) was found in the glauconite clay layers.

- (3) Further evidence of normal sedimentation is found in the stable carbon isotope record, which shows characteristic late Maastrichtian values above the impact breccia and the K-T negative excursion at the boundary (Fig. 8). These geochemical signals provide further support for normal sedimentation.
- (4) Paleomagnetic data also reveal late Maastrichtian sediments of Chron 29r (Keller et al., 2004a).

- (5) The presence of late Maastrichtian planktic foraminifera, including the index species *Plummerita hantkeninoides*, indicates deposition occurred during the last 300 ky of the Maastrichtian (Fig. 8) (Keller et al., 2004a,b).

Thus there are five independent lines of evidence that support normal sedimentation during the latest Maastrichtian after deposition of the impact breccia. The likelihood of an older pre-K-T age for the Chicxulub impact was previously noted in a study of older cores across Yucatan and the impact crater (Ward et al., 1995; Cantu-Chapa and Landeros-Flores, 2001; Cantu-Chapa, 2004). Moreover, the evidence from the Yaxcopoil-1 crater core

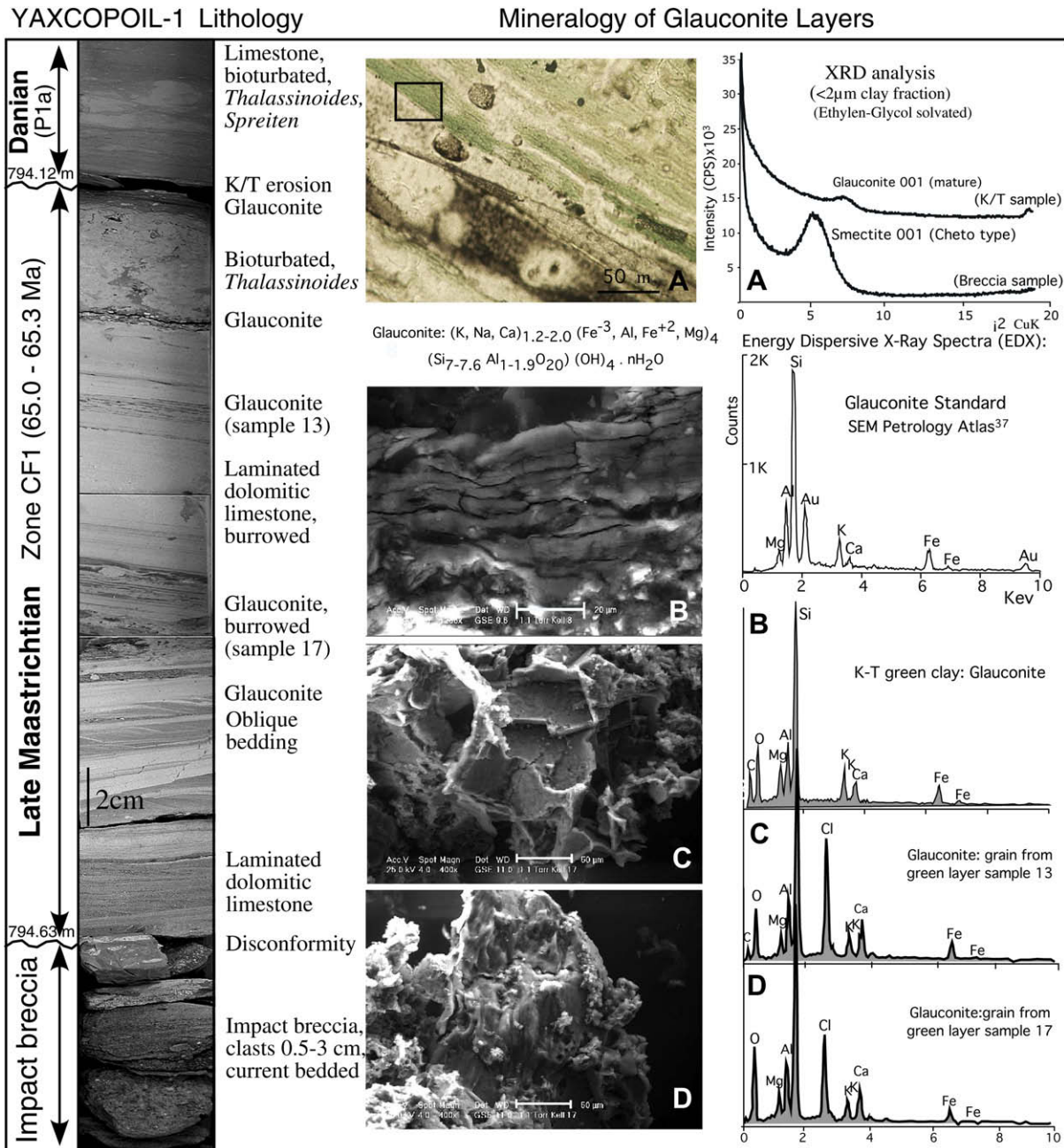


Fig. 7. Chicxulub impact crater core Yaxcopoil-1 litholog showing the finely layered micritic limestone and five glauconite clay layers in the critical 50 cm interval between the impact breccia and the K-T boundary. A. Thin-section micrograph of the green K-T clay layer with insert marking location of analysis. The XRD diffractogram indicates mature glauconite; in contrast, XRD of breccia clay shows typical Cheto smectite, the alteration product of impact glass. B, C, D. Environmental scanning electron microscope micrographs of the K-T green clay and green clay layers samples 13 and 17 also show mature glauconite. The glauconite reference standard from the SEM Petrology Atlas is shown for comparison. (Modified from Keller et al., 2004a). The presence of these glauconite layers indicates deposition occurred over a very long time period (tens of thousands of years), which is incompatible with the backwash and crater-fill interpretation.

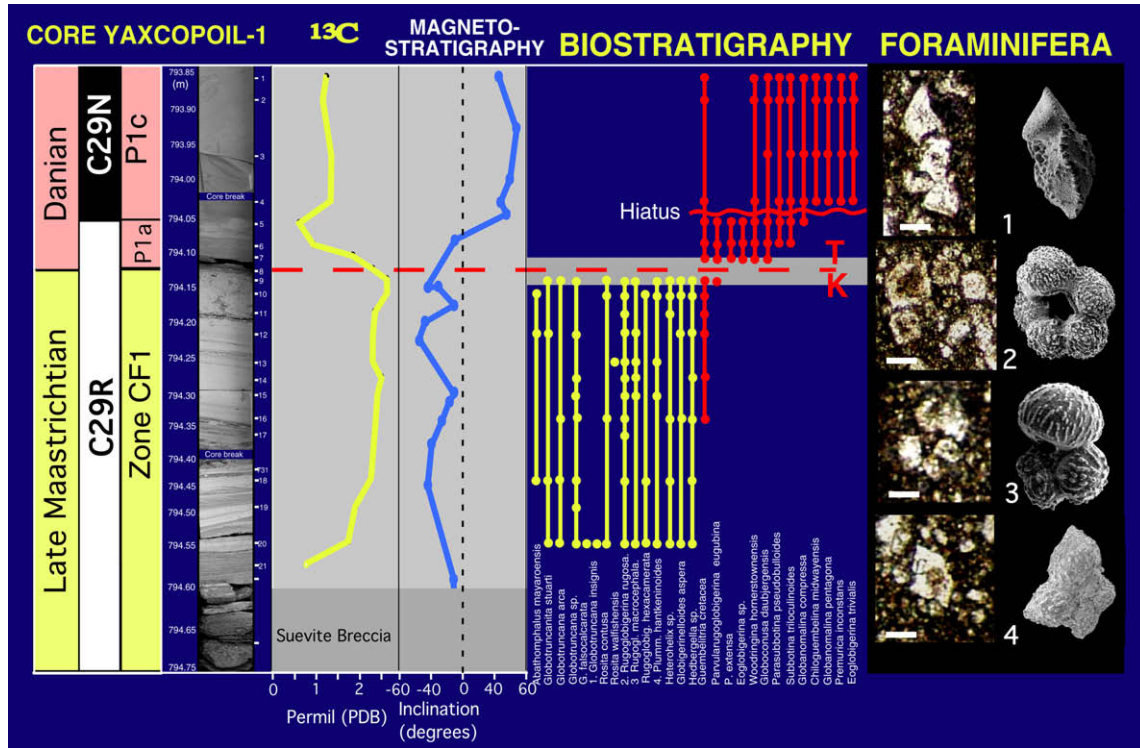


Fig. 8. In Chicxulub crater core Yaxcopoil-1 the pelagic limestone between the impact (suevite) breccia and the K-T boundary contains: (1) typical late Maastrichtian carbon-13 isotope values followed by the characteristic negative $\delta^{13}\text{C}$ at the K-T boundary, (2) the latest Maastrichtian chron 29r magnetic signature, (3) late Maastrichtian planktic foraminifera, including the zone CF1 index species *Plummerita hantkeninoides*, which spans the last 300 ky of the Maastrichtian. These age related proxies indicate that the Chicxulub impact predates the K-T boundary by about 300 ky (Modified from Keller et al., 2004a.).

shows great stratigraphic similarity with sections from NE Mexico.

Critics have largely countered this multi-disciplinary evidence with claims that (a) foraminifera in the limestone of the crater core Yaxcopoil-1 don't exist and are simply dolomite crystals (Smit et al., 2004), or (b) that foraminifera exist, but are all reworked (Arz et al., 2004). Naturally, one would expect some reworked specimens from the surrounding crater walls, but most species Arz identified ranged through the Maastrichtian and can't be differentiated from reworked specimens. No data contradicting our evidences 1 to 5 have been presented to date.

4.3. Brazos River, Falls County, Texas

One way of solving the controversy over the age of the Chicxulub impact is to drill in a location farther from the impact crater and in shallow water sequences where slumps can be excluded. The K-T sequences along the Brazos River are ideal. Located on a shallow Cretaceous inland sea (<100 m), 1000 km from the Chicxulub crater, they show continuous sedimentation, excellent fossil preservation and the presence of Chicxulub ejecta spherules. Earlier studies of the Brazos-1 reference section exposed along the river bank revealed the K-T boundary and Ir anomaly about 20 cm above the "event deposit" with its reworked impact spherules at the base (Fig. 9) (Jiang and Gartner, 1986; Keller, 1989; Rocchia et al., 1996). However, the position of the K-T boundary is currently under review. The frequently lenticular "event deposits" infill submarine channels with the basal unconformity overlain by clasts, glauconite, shellhash and reworked Chicxulub spherules. Above it the laminated and hummocky cross-stratified sandstones are frequently burrowed (Yancey, 1996; Gale, 2006). These deposits have been variously interpreted as impact tsunami (Bourgeois et al., 1988;

Smit et al., 1996; Heymann et al., 1998), sea level lowstand (Keller, 1989; Yancey, 1996), or seasonal storm deposits (Gale, 2006).

In March 2005, the U.S. National Science Foundation supported the drilling of three cores, which was successfully completed by DOSECC (Drilling, Observation and Sampling of Earths Continental Crust) with nearly 100% recovery. Our objectives were to determine the stratigraphic position of the K-T boundary with respect to the event deposit and investigate late Maastrichtian sediments for evidence of the original Chicxulub ejecta layer. Apart from the drilling, we conducted fieldwork along the tributaries of the Brazos River and collected various outcrops that span the K-T boundary. These studies are still in progress, but studies of the most expanded K-T transition in core Mullinax-1 reveal: (a) 80 cm of late Maastrichtian dark shale with burrows between the K-T boundary and the top of the event deposit (Fig. 10), (b) an event deposit that consists of separate seasonal storm events with invertebrates repeatedly colonizing the sea floor between storm deposits (Gale, 2006), (c) two to three reworked spherule depositional events with upward fining sandstone at the base, and (d) the discovery of the original Chicxulub impact layer 40 cm below the base of the event deposit (Keller et al., 2007).

4.3.1. Lithology of the K-T boundary and event deposit

The lithology of the event deposit is best illustrated by the new core Mullinax-1 (Fig. 11). Erosion marks the base of the event deposit and is interpreted as a sea level lowstand or sequence boundary (SB). Above it are three units of spherule-rich coarse sandstones (SCS) with the lower parts consisting of poorly sorted, upward fining sandstones with abundant shell fragments, glauconite, mudstone and phosphatic clasts and Chicxulub impact spherules. The upper parts of units 1 and 2 grade into thin layers of rippled or cross-bedded sandstones. These units are diagnostic of storms with high-energy debris flows followed by upward waning

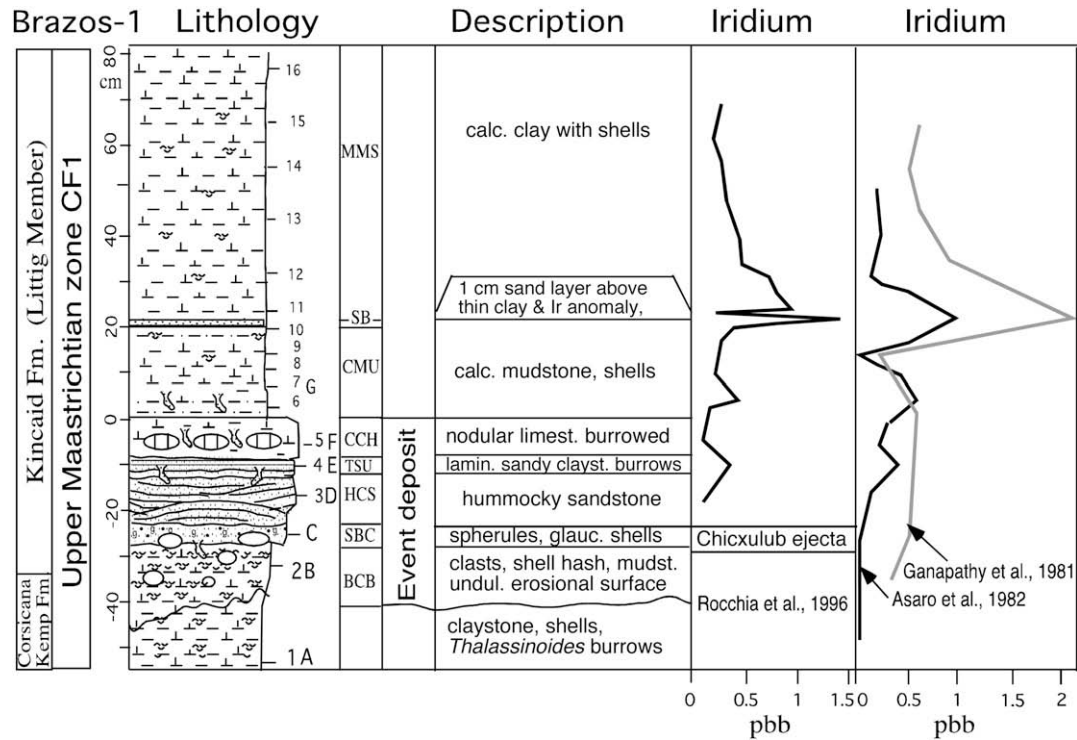


Fig. 9. Stratigraphy of the classical Brazos-1 section near the Route 413 bridge across the Brazos River, Falls County, Texas. Three Ir studies recorded the anomaly in a thin red-brown layer and 1 cm sandstone 17–19 cm above the calcareous claystone (CCM) that forms the top of the event deposit. The two minor Ir enrichments just above the CCM layer and within a laminated sandy claystone of the event bed are likely due to postdepositional processes.

energy depositing the rippled sandstones. An erosional surface separates the spherule-rich units from the overlying hummocky cross-bedded sandstone (HCS) of unit 4, which is strongly burrowed (Fig. 11). The burrows are infilled with darker mudstone and truncated by erosion at the top. Burrows within this unit were previously observed as *Ophiomorpha nodosa*, *Thalassinoides* sp., and *Planolites* sp. (Gale, 2006).

The erosional surface above the HCS is draped by fine gray sandstone and silty mudstone with swaley and horizontal laminations that indicate comparable energy to the HCS. Small burrows are present and have also been observed in outcrops of the Brazos River bed and identified as *Thalassinoides* sp. and *Planolites* sp. (Gale, 2006). The top of the event deposit is marked by a 16 cm thick calcareous silty claystone (CCH, Fig. 11) with upward-fining grain size and grading from light grey to darker grey at the top. Above this interval, normal shallow shelf sedimentation resumed with dark grey laminated mudstones (Fig. 10). Discrete burrows infilled by pyrite framboids mark typical dysaerobic biofacies (Wignall, 2001). Deposition occurred in a very shallow (20–30 m depth) inner neritic environment, which gradually deepened upwards towards the K-T boundary and above. No lithological change marks the K-T boundary, which is identified at 80 cm above unit CCH by (1) the global negative $\delta^{13}\text{C}$ excursion, (2) the extinction of tropical and subtropical Cretaceous planktic foraminifera, and (3) the first appearances of Danian planktic foraminifera (*Woodringina hornerstownensis*, *Parvularugoglobigerina extensa*, *Globigerina daubjergensis*), which are usually found within a few cm of the iridium anomaly and $\delta^{13}\text{C}$ shift worldwide.

4.3.2. Original Chicxulub impact ejecta layer

We discovered what appears to be the undisturbed original impact spherule ejecta layer 40 cm below the event deposit in outcrops of the Cottonmouth Creek tributary of the Brazos River (Fig. 12). This spherule ejecta layer is now altered to a 3 cm thick

yellow clay, which consists of 100% Mg-smectite, or Cheto smectite. Cheto-smectite is derived from weathering of glass, as indicated by the 001 reflection and low crystallinity index (0.5 to $0.8 \cdot 2\theta$) with the 060 reflection around 61° indicating a composition corresponding to nontronite, or cheto Mg-smectite (Debrabant et al., 1999; Keller et al., 2003b). In contrast, smectite in Maastrichtian and Danian claystones is a common montmorillonite derived from weathering of soils, as indicated by the 060 reflection between 61.7° and 62.3° (Moore and Reynolds, 1997). The same cheto-smectite clay phase is also present in the altered Chicxulub glass spherules of the event deposit (Keller et al., 2007). This strongly suggests a common impact origin for the altered glass spherules of the event deposit and the yellow clay 40 cm below. Both altered spherule layers reveal typical Mg enriched cheto type-smectite high in SiO_2 (66–71%), Al_2O_3 (19–20%), FeO (4.4–4.8%), MgO (2.8–3.3%) with minor K_2O (1–1.1%) and NaO (<0.5%). This composition is very similar to the altered smectite rims from Haiti glass spherules and to NE Mexico glass spherules (Schulte et al., 2003; Harting, 2004; Stüben et al., 2005).

Additional evidence of the pre-K-T age of the Chicxulub impact spherule ejecta is evident from clasts at the base of the event deposit that infills the submarine channels (Keller et al., 2007). The lithified mudstone clasts derived from the underlying hemipelagic facies, as indicated by late Maastrichtian planktic foraminifera. They also reveal a complex depositional history of the Chicxulub impact spherules. For example, some clasts contain lenses of impact spherules, whereas others contain morphologically well-preserved spherules (Fig. 13C, D), which were incorporated into the sediments prior to lithification, erosion, transport and re-deposition. In some clasts there are fractures, or cracks, infilled with spherules and the fractures rimmed by several generations of sparry calcite, then truncated and followed by normal sedimentation (Fig. 13E, F). This suggests complex diagenetic processes and possible emergence prior to erosion and transport. These clasts with impact spherules provide very strong evidence of the

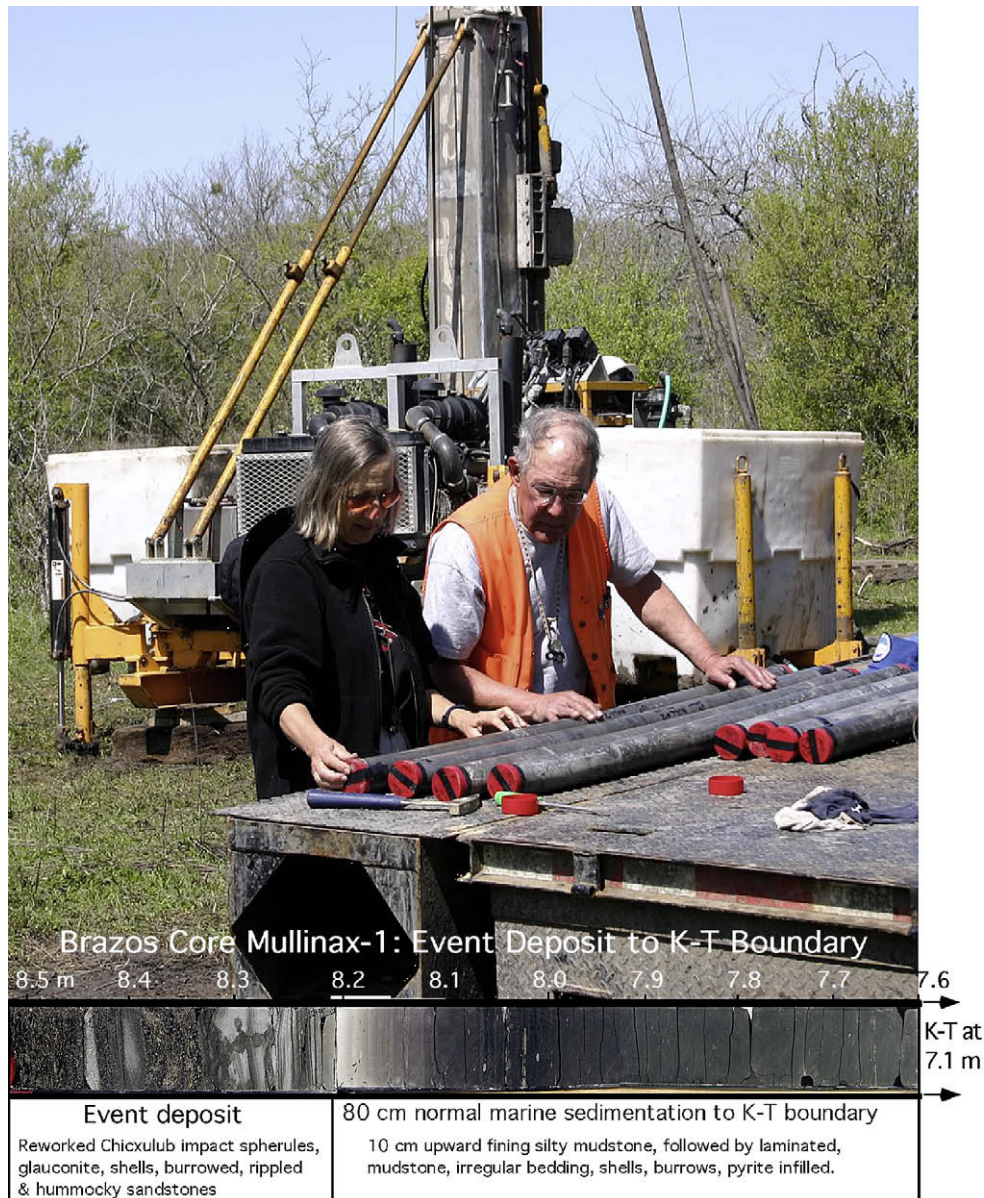


Fig. 10. New drilling of Brazos K-T sequences in March 2005 (by DOSECC; Gerta Keller and Jerry Baum) recovered an expanded sequence with the K-T boundary 80 cm above the top of the event deposit. The 80 cm interval consists of bedded, burrowed, shelly shale and mudstone with late Maastrichtian microfossils assemblages of planktic foraminifera, calcareous nannofossils, palynomorphs and late Maastrichtian carbon isotopes.

existence of an older spherule-ejecta layer, which was lithified and subsequently eroded, transported and redeposited at the base of the event deposit.

Planktic foraminiferal biostratigraphy places the age of this ejecta layer near the base of zone CF1, or 300 ky below the K-T boundary, similar to NE Mexico and the Chicxulub crater core Yaxcopoil-1. Thus, the Brazos sections provide additional evidence that the Chicxulub impact predates the K-T boundary by about 300,000 years. The Brazos and NE Mexico sections further show that the Chicxulub impact occurred prior to the sea level lowstand near the end of the Maastrichtian, which caused erosion, transport and infilling of submarine channels. During this sea level lowstand, the original Chicxulub ejecta deposits were eroded in nearshore areas, transported and redeposited in the submarine channels. Most early studies of Chicxulub impact ejecta concentrated on these submarine canyon deposits as impact-generated tsunami deposits and failed to observe the original impact ejecta layer in the underlying Maastrichtian sediments.

5. Impacts and the K-T mass extinction

Stratigraphic evidence from NE Mexico, Texas and the Chicxulub crater show that the Chicxulub impact predated the K-T boundary by 300 ky and therefore could not have caused the mass extinction. No sudden species extinctions occurred during the last 300 ky of the Maastrichtian in any fossil group, though rudistids disappeared around this time (Johnson and Kauffman, 1996) and there is a gradual decrease in species diversity and increased biotic stress during the last 500 ky that is generally attributed to climate changes (e.g., cooling followed by greenhouse warming due to Deccan volcanism, Kucera and Malmgren, 1998; Li and Keller, 1998a,b,c; Olsson et al., 2001; Abramovich and Keller, 2003; Keller and Pardo, 2004a). We can therefore conclude that impacts forming craters as large as 170 km in diameter do not cause mass extinctions (Fig. 3).

However, the K-T mass extinction is closely associated with a global Ir anomaly, which is considered too large, too widespread,

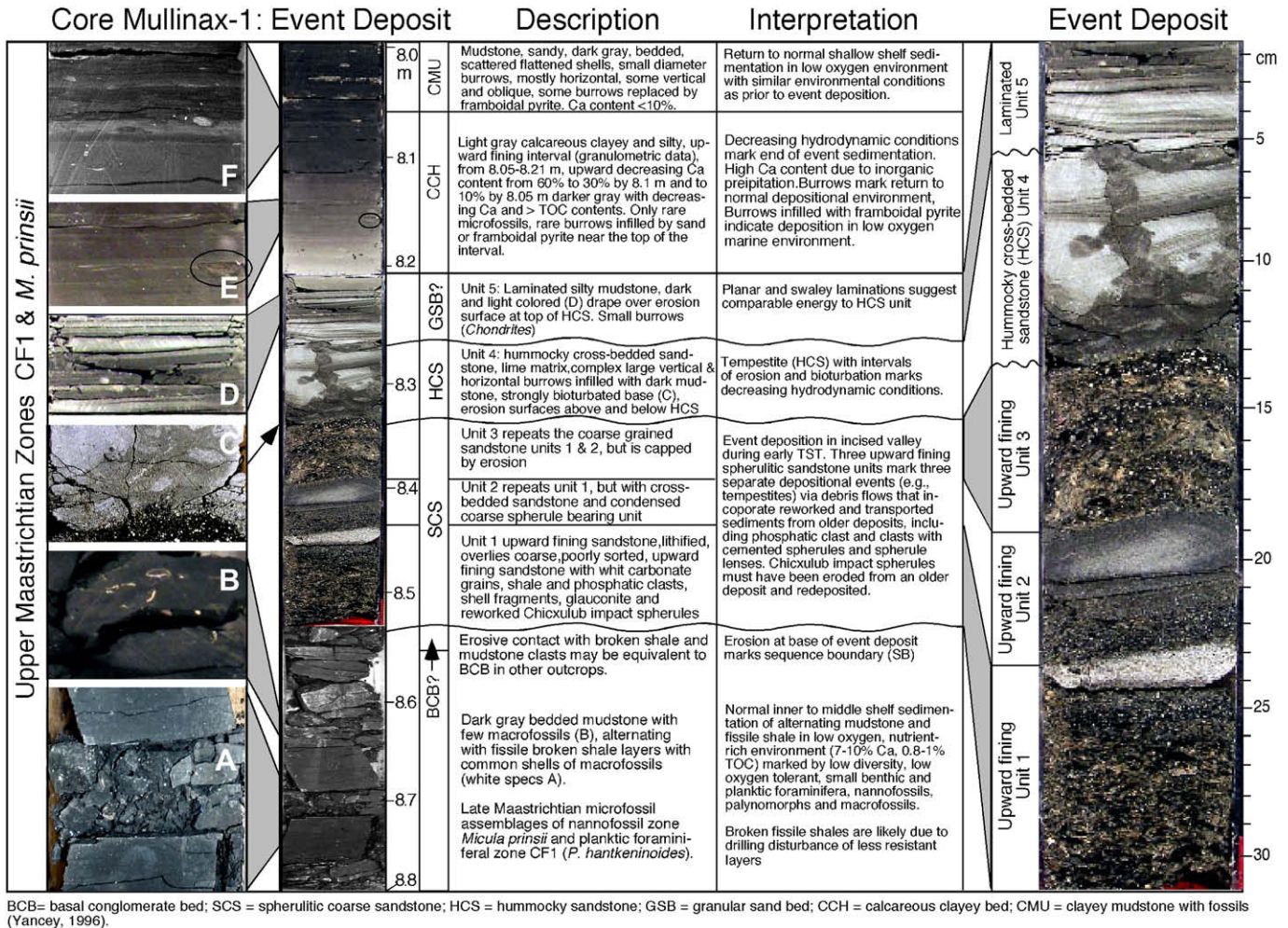


Fig. 11. Lithology, description and interpretation of the event deposit in core Mullinax-1. Note the three upward fining glauconitic spherule-rich coarse sandstone units topped by fine sandstone layers (unit SCS) and the truncated burrows in unit HCS. These are diagnostic features of multiple storm deposits. (Modified after Keller et al., 2007.)

and too concentrated in a thin layer to have originated from anything other than a large extraterrestrial impact (Sawlowicz, 1993; Rocchia et al., 1996). Thus, we can infer that, in addition to the late Maastrichtian Chicxulub impact, another larger impact occurred at the K-T boundary with more grave biotic consequences. Though no crater is known to date, the global Ir distribution and biotic effects suggest that this K-T impact crater should exceed 200 km in diameter (see Fig. 3). The biotic effects of the hypothesized K-T impact, however, must be considered speculative. We cannot assume that the mass extinction was solely caused by a large impact, because the biotic effects of the Deccan volcanism are largely unknown and those of climate change still inadequately studied.

Nevertheless, the climate warming, probably due to Deccan volcanism between 65.2–65.4 Ma preceding the K-T mass extinction, and its detrimental biotic effects are relatively well known (Fig. 14). Biotic stress is observed in the decreased species diversity and the dwarfing of species generally reaching half or less than normal adult size. This is interpreted as a result of adaptation to biotic stress with species maturing earlier and reproducing at a faster rate as a survival tactic (Abramovich and Keller, 2003). During the late Maastrichtian warming, this adaptation technique is observed in cosmopolitan species as well as large specialized tropical and subtropical species. Most of the cosmopolitan species survived the K-T event, but soon became extinct, probably due to competition from the evolving Danian fauna (Keller, 2001). For the

more specialized tropical and subtropical species, this survival technique failed, species populations dwindled and all species became extinct when another environmental disturbance hit at the K-T boundary. It thus appears that the K-T boundary mass extinction may have been due to the culmination of long-term biotic stress conditions due to volcanism and climate changes that led to decreased abundance of large specialized tropical and subtropical species leaving them prone to extinction (Keller, 2001). In this environment the K-T impact may have been the final trigger, rather than the ultimate cause of the mass extinction.

6. Does volcanism cause mass extinctions?

Courillot (1999) and Wignall (2001) made a convincing case for a potential cause-and-effect relationship between volcanism and mass extinctions during the Phanerozoic where four of the five major mass extinctions are associated with LIPs, but only one, the K-T boundary, with a large impact (Fig. 1). Demonstrating a rough correlation between LIPs and mass extinctions is relatively simple because volcanic eruptions generally occurred over long time periods. But determining volcanisms' specific adverse biotic effects is difficult because of the confluence of many environmental changes (e.g., climate, sea level, ocean anoxia, impacts) over the long course of volcanic activities. For the Cretaceous, Deccan volcanism correlates with the K-T mass extinction, as well as the Chicxulub and K-T

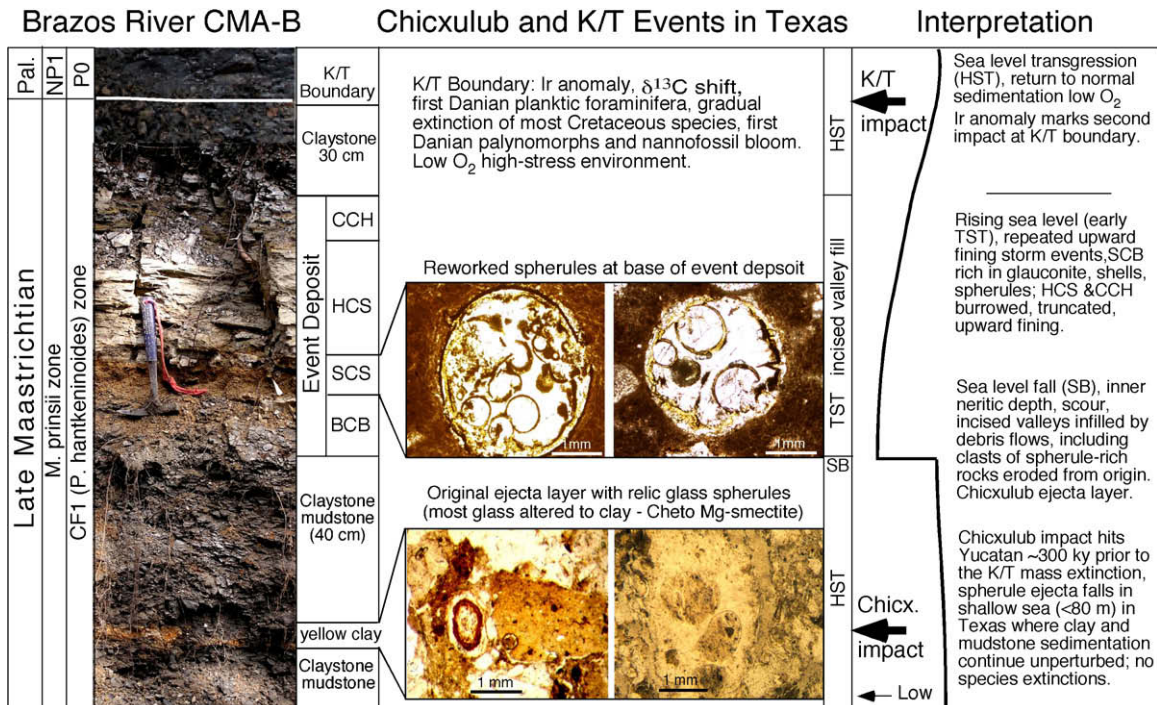


Fig. 12. Sequence of three events identified at the CMA-B outcrops: (1) The yellow clay marks the Chicxulub spherule ejecta layer now altered to cheto smectite with relic glass; this impact occurred near the base of zone CF1, about 300 ky before the K-T boundary. (2) The event deposit marks the latest Maastrichtian sea level fall, channel scour and subsequently infilling by storms during the early transgression. (3) The K-T boundary occurs during the subsequent sea level high and is marked by the global $\delta^{13}\text{C}$ shift and first appearance of Danian species (Modified after Keller et al., 2007).

impacts (Fig. 15). No increased extinction intensity or OAEs are associated with the Madagascar, Caribbean and Ontong Java I LIPs (Fig. 2). No LIP is associated with OAE2 and its elevated extinctions (17%) (reviews in Premoli Silva et al., 1999; Hart et al., 2002). Moreover, Smith et al. (2001) discussed the C-T (OAE2) extinctions in macrofossils as possibly an artefact of preservation. The late Aptian OAE coincided with the Rajmahal-Kerguelen volcanic eruptions, but caused only minor elevated extinctions (Luciani et al., 2001). Other LIPs appear to have had no significant biotic effects when evaluated on the basis of generic extinctions (Figs. 2, 15). This begs the question, are generic extinctions sufficiently sensitive indicators of environmental and biotic effects? Species level data suggest that this is not the case.

A study of the biotic effects of volcanism during the late Maastrichtian on Ninetyeast Ridge DSDP Site 216 revealed that at the species level the adverse biotic effects of volcanism are identical to the K-T mass extinction in planktic foraminifera (Keller, 2003, 2005b). The disaster opportunists (*r*-selected) *Guembeltria* are the dominant survivors (80–90%) with minor abundances of (*r*-selected) low oxygen tolerant small heterohelicids. Recovery of the ecosystem results first in increased abundance of heterohelicids, eventual disappearance of the disaster opportunists, followed by the return of small surface dwellers and finally the (*k*-selected) larger specialized species (see recovery model in Pardo and Keller, this issue). The actual cause of biotic stress appears to be a large influx of biolimiting minerals and nutrients from mantle plume volcanism and increased weathering leading to eutrophication of surface waters, as evident by the phosphate-rich volcanic sediments. Similar observations were made for the late Albian and late Cenomanian OAEs (Jarvis et al., 1988; Galeotti, 1998; Hart et al., 2002; Keller et al., 2001, 2004c; Keller and Pardo, 2004b). Trace element composition of foraminiferal tests reveals the periodic influx of metal-bearing hydrothermal fluids (Co, Cu, Zn, Pb) and micronutrients (Ni and Cd) related to intensive volcanic activity on the Ninetyeast Ridge.

7. Cretaceous climate and carbon cycles

Composite Cretaceous oxygen and carbon isotope curves were generated based on the early Cretaceous bulk rock isotopes from the Tethys Ocean compiled by Weissert and Erba (2004) and for the middle and late Cretaceous from planktic and benthic foraminifera of the southern oceans (Huber et al., 1995; Li and Keller, 1998b,c). In order to correlate these high-resolution original datasets with the genus-level data at equal stage lengths, the isotope data were necessarily compressed into equidistant stage intervals regardless of age duration. Some resolution is therefore lost, though the overall trends are preserved. Actual temperature trends for the middle and Late Cretaceous are based on well-preserved benthic and planktic foraminifera (Huber et al., 1995; Li and Keller, 1998b,c). For the early Cretaceous the $\delta^{18}\text{O}$ bulk rock values cannot be easily translated into absolute temperatures because of diagenetic alteration (Weissert et al., 1998), though the trends serve as proxies for past climate changes.

7.1. Climate changes - $\delta^{18}\text{O}$ record

The generalized composite $\delta^{18}\text{O}$ curve reveals a complex and highly variable Cretaceous climate regime (Fig. 11). During the early Cretaceous climate cooled from the Beriasian through the Valanginian with the coolest temperatures in the early Hauterivian. Subsequent warming culminated in the early Barremian and may have been associated with the Parana-Etendeka volcanic province (132–134 Ma). (Note that the new time scale by Gradstein and Ogg (2004) places this LIP in the Hauterivian, as compared with the Valanginian in the old time scale). Increased warming during the late Barremian reached maximum greenhouse conditions during the early Aptian, possibly associated with the Ontong Java I volcanic province (122 Ma). During the late Aptian and into the early Albian climate cooled significantly.

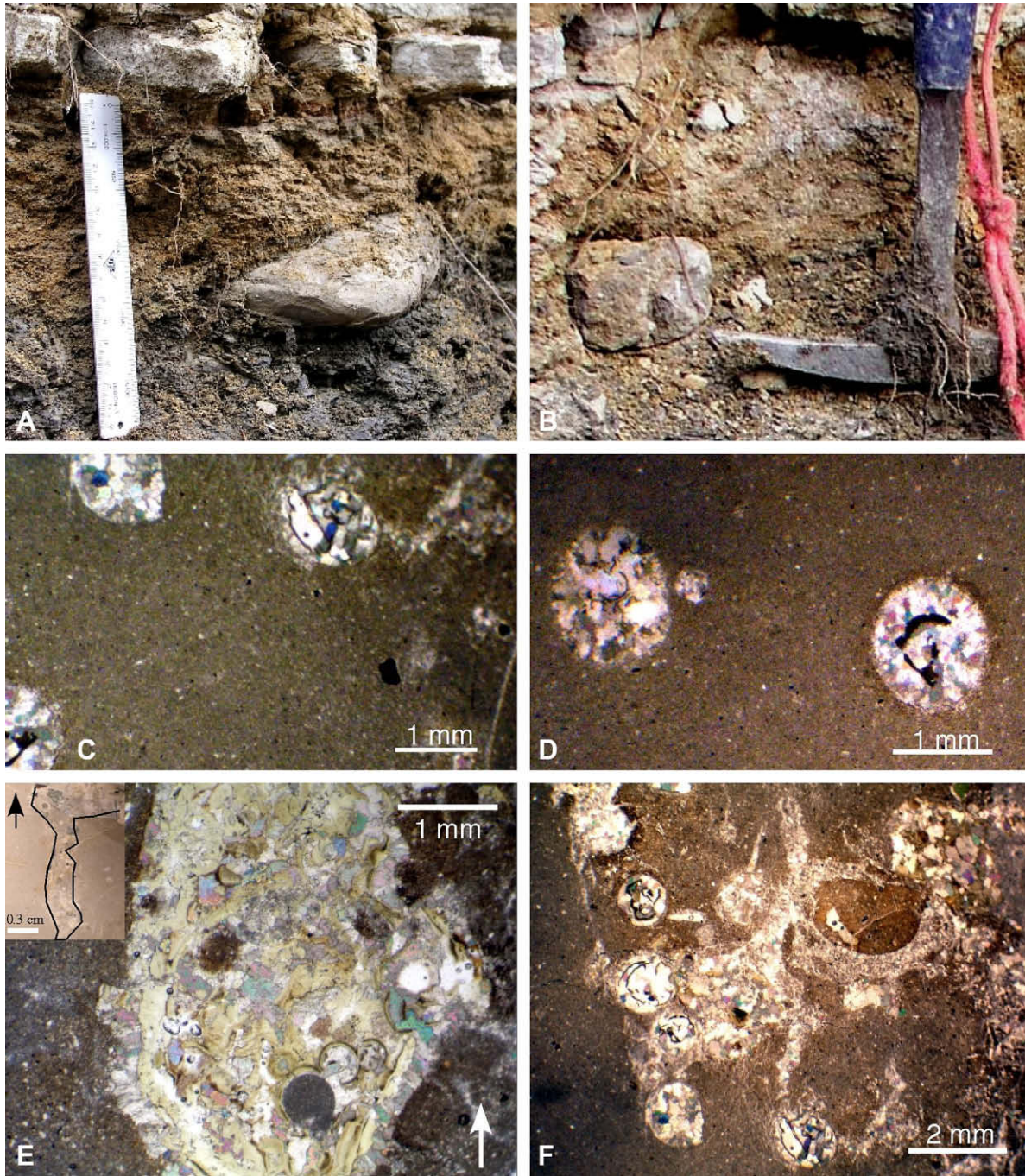


Fig. 13. A, B: Clasts from the basal conglomerate (BCB) of the event deposit contain Chicxulub impact spherules. C, D: spherules in mudstone clasts. E, F: spherules within cracks of mudstone clasts. E: cracks rimmed by sparry calcite. Insert shows morphology of crack and total length of ~2 cm. F: clast with cracks infilled with spherules and sparry calcite, then truncated by erosion and followed by normal sedimentation. These clasts reveal a history of Chicxulub ejecta fallout and lithification well prior to exposure to erosion, transport and redeposition at the base of the event deposit.

Climate warming during the Cenomanian reached maximum marine surface temperatures (28 °C) for the Cretaceous in high southern latitudes during the early Turonian (Huber et al., 1995). Warm, though reduced temperatures (24 °C) prevailed through the Turonian and Coniacian, a time interval associated with LIPs in the Caribbean (89–90 Ma), Ontong Java II (90 Ma) and Madagascar (86–90 Ma). During the Santonian through early Campanian warm but fluctuating climate conditions prevailed, followed by rapid cooling of surface waters in the late early Campanian (Jarvis et al., 2006). Gradual cooling continued through most of the Maastrichtian and ended with a short-term abrupt greenhouse warming

200–400 ky before the end of the Maastrichtian followed by gradual cooling across the Cretaceous-Tertiary (K-T) boundary (Li and Keller, 1998b,c; Kucera and Malmgren, 1998; Olsson et al., 2001). The end-Cretaceous greenhouse warming was associated with Deccan volcanism (65–66 Ma), the Chicxulub impact, and severe biotic stress conditions among planktic foraminifera that led to the K-T mass extinction.

Diversity trends suggest that generic diversity increased during cooler climates from the Aptian through Cenomanian and again during the Campanian and Maastrichtian. This may be related to increased weathering and runoff, upwelling and nutrient cycling. In

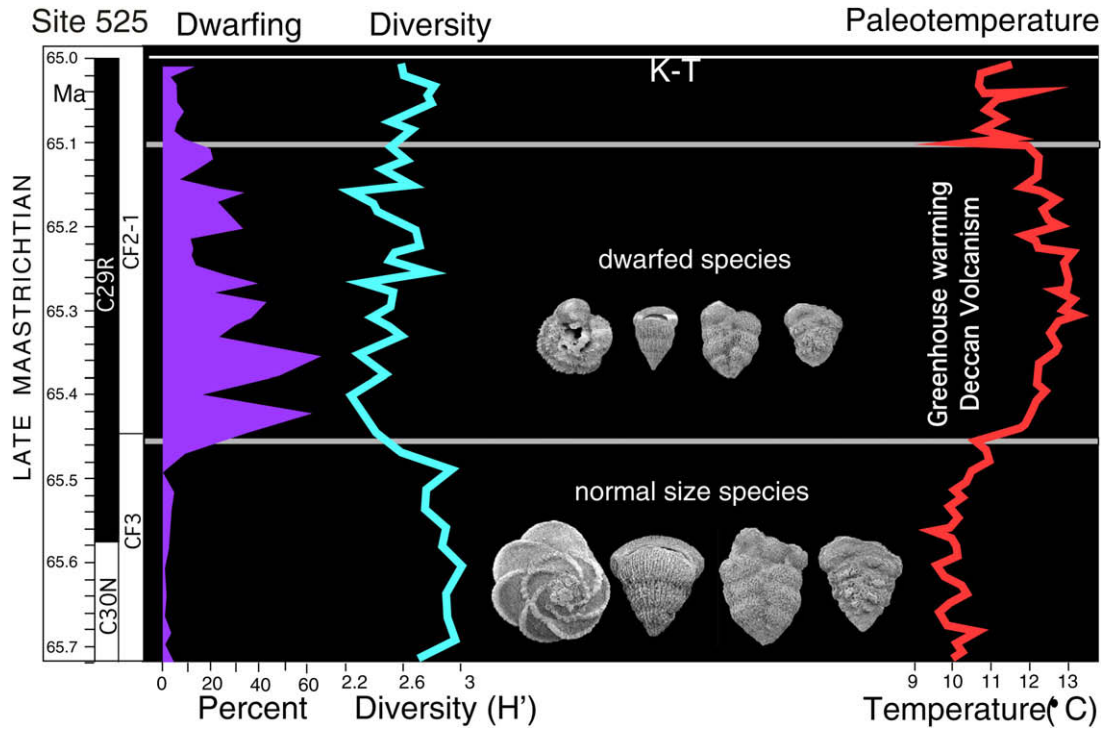


Fig. 14. Biotic effects of latest Maastrichtian climate warming at mid-latitude South Atlantic DSDP Site 525 include decreased species diversity, dwarfing of species to less than half their normal adult size, and decreased abundance of species. These biotic effects are present in the small ecologically tolerant species as well as the large specialized species. Modified from Abramovich and Keller, 2003.

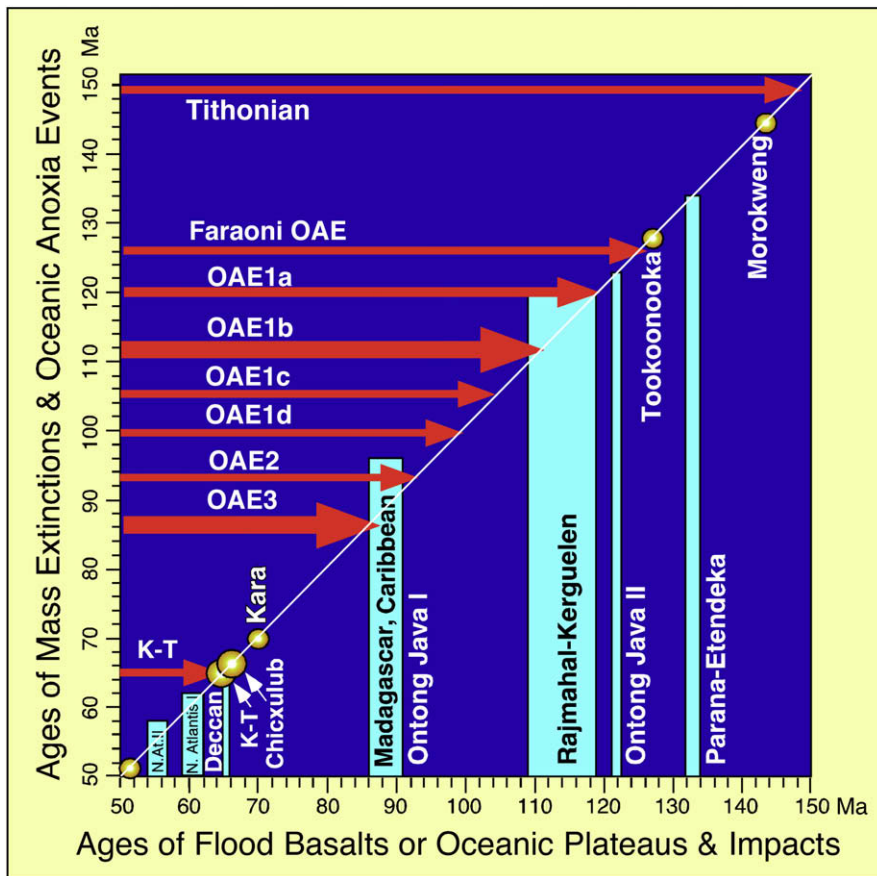


Fig. 15. A plot of Cretaceous volcanism and impacts versus extinction intensity and oceanic anoxic events (OAEs) reveals a good correlation for the K-T boundary and potential correlation between LIPs and OAE1a, 1b and OAE3, though precise age control is lacking. LIPs may contribute to OAEs, but apparently are not the primary cause; there is no correlation with impact other than at the K-T boundary. Modified after Courtillot, 1999.

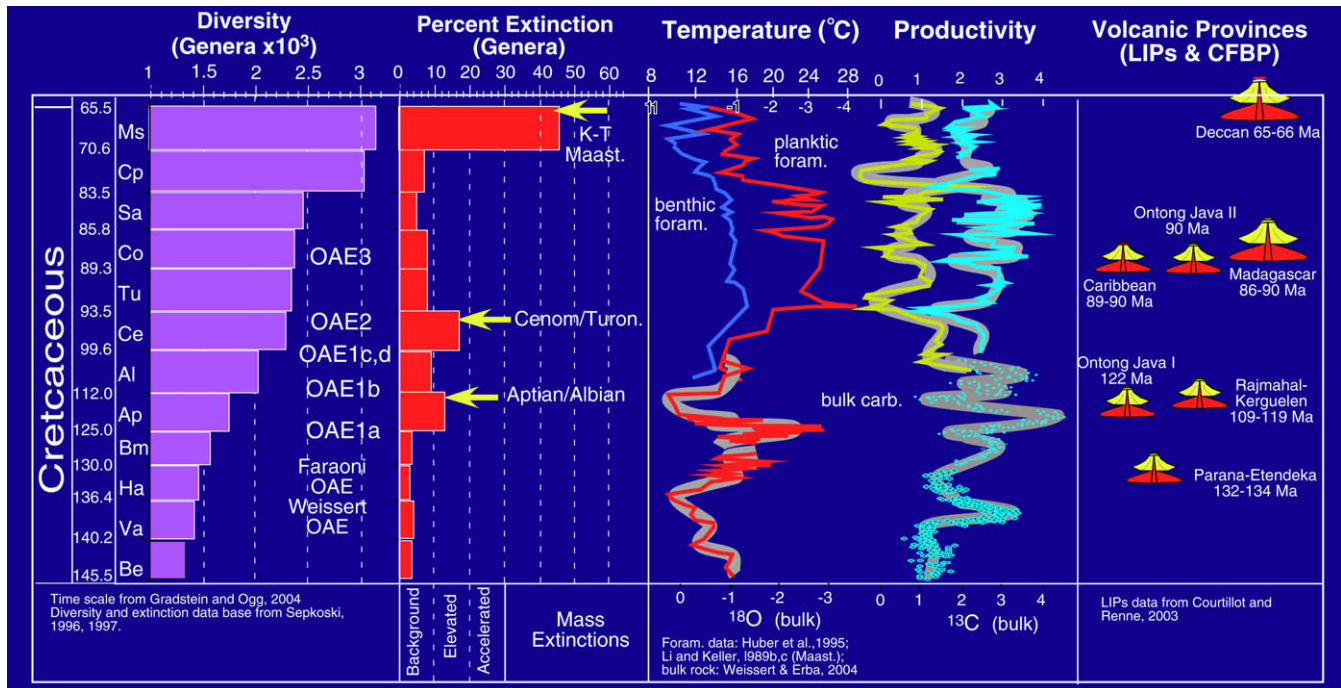


Fig. 16. Cretaceous temperature and carbon isotope (productivity) trends compared with LIPs, diversity and extinction trends. Note that the warmest Cretaceous climate coincides with the most stable species diversity, whereas cool climates correlate with increased diversity related to increased nutrient cycling, watermass stratification and niche creation (e.g., Campanian–Maastrichtian, Aptian–Albian). LIPs occur during warm and cool climates and do not appear to dramatically alter climate at this scale.

contrast, diversity remained static during the warmest Cretaceous temperatures (Turonian–Santonian, Fig. 16), which were also associated with high sea levels. OAEs generally correlate with warmer climates, but at least one (Aptian/Albian) occurred during climate cooling and may have been associated with LIPs. There appears to be no direct correlation of OAEs with other LIPs, though this may be partly an artifact of the new Gradstein and Ogg (2004) time scale.

7.2. $\delta^{13}\text{C}$ isotope record

The composite $\delta^{13}\text{C}$ curve for the Cretaceous shows highly variable conditions with large positive and negative anomalies. In the early to middle Cretaceous bulk rock data show major positive $\delta^{13}\text{C}$ excursions during the Valanginian, Barremian, Aptian, Albian, latest Cenomanian and Santonian, and negative excursions in the early Valanginian, latest Barremian, early Aptian (OAE1a) and late Cenomanian (Fig. 16). The negative anomalies are associated with platform drowning and reduced carbonate production (Föllmi et al., 1994; Wissler et al., 2003). Weissert and Erba (2004) linked these drowning episodes to major biocalcification crises in nannoconids, leading to decreased abundances of heavily calcified species and increase in the rapidly producing (*r*-selected) nanofossils, though without net loss in diversity (Erba, 1994, 2004; Bralower et al., 1999). The suggested cause is eutrophication of surface waters due to volcanism or increased weathering and high terrestrial nutrient influx, $p\text{CO}_2$ levels and temperature of the surface ocean. The negative $\delta^{13}\text{C}$ excursions would thus reflect decreased primary productivity.

Similar *r*-selected species abundance changes are also observed in planktic foraminifera during OAEs and platform drowning events when low oxygen tolerant biserial species dominate (e.g., Hart, 1980; Leckie, 1987; Leckie et al., 2002; Coccioni and Luciani, 2004; Keller et al., 2001, 2004c) and radially elongated taxa are common (see review in Coccioni et al., 2006). Prior to and during the gradual OAE2 positive excursion all heavily calcified and larger

species (i.e. rotaliporids, large whiteinellids) decreased in abundances and disappeared by the time of the maximum positive $\delta^{13}\text{C}$ excursion and onset of OAE2 (Keller and Pardo, 2004b). Mort et al. (2007) observed increased accumulation of phosphorus preceding the positive $\delta^{13}\text{C}$ excursion and returning to pre-excursion values just prior to the first $\delta^{13}\text{C}$ peak. They suggest a global decoupling in the carbon and phosphorus cycles occurred within some tens of thousands of years after the event started.

Between the lower and upper Campanian the negative $\delta^{13}\text{C}$ excursion is associated with the onset of climate cooling that continues into the late Maastrichtian followed by a positive excursion associated with climate warming and Deccan volcanism (Fig. 16). Huber et al. (1995) show increased $^{87}\text{Sr}/^{86}\text{Sr}$ ratios during this time, which suggest increased continental weathering rates and possibly decreasing atmospheric $p\text{CO}_2$ levels. During this time marine bioproductivity significantly increased and diversity reached maximum Cretaceous levels.

8. Discussion and conclusions

Current causal mechanisms for Cretaceous OAEs favor the high productivity model (e.g., Weissert et al., 1985; Föllmi et al., 1994; Föllmi, 1995), enhanced preservation model due to increased thermohaline stratification (Wilson and Norris, 2001; Erbacher et al., 2001), or a combination of high productivity and enhanced preservation, greenhouse warming and submarine volcanic activity (Jenkyns, 2003; Bodin et al., 2006; Mort et al., 2007). The various models largely reflect the variation in environmental conditions associated with OAEs, which cannot be accounted for by a single causal mechanism.

Cretaceous volcanic activities (LIPs and CFBPs) appear to have had relatively minor biotic effects at least at the generic level. Nevertheless, the major biotic stress levels during the Cretaceous were associated with OAEs and related to nutrient availability largely from weathering, greenhouse warming, drowning of

platform areas and volcanism. The biotic effects of OAEs were often dramatic at the species level, causing the extinction of larger specialized and heavily calcified planktonic foraminifera (rotalipora extinction) and nannoconids (nannoconid crises), the temporary disappearances of other species, and the rapid increase in *r*-selected species, such as the low oxygen tolerant heterohelicids and radially elongated taxa among planktic foraminifera (Leckie, 1987; Luciani et al., 2001; Keller and Pardo, 2004b; Coccioni and Luciani, 2004; Coccioni et al., 2006) and thin walled nannofossils (Erba, 1994, 2004). No mass extinction is associated with any OAE, though elevated extinctions are recorded for the late Cenomanian OAE2. At the species level in planktic foraminifera of OAE2, the extinction is masked largely by the one-for-one replacement of outgoing species by new species.

There is general agreement that with the exception of the K-T boundary there is a lack of evidence that impacts caused major mass extinctions during the Phanerozoic (Courtilot, 1999; Wignall, 2001; MacLeod, 2003; Keller, 2005b; White and Saunders, 2005). With evidence from NE Mexico, Texas and the Chicxulub crater itself that this 170 km diameter crater predates the K-T boundary by 300,000 years, this impact can no longer be considered a direct cause for the mass extinction (Keller et al., 2003a, 2004a,b, 2007). However, the presence of the global K-T Ir anomaly is difficult to explain by volcanic origin and suggests that a second still unknown larger impact may have occurred. Deccan volcanism and climate changes during the last 500 ky of the Maastrichtian lead to increasing biotic stress, reduced abundances of large specialized tropical planktic foraminiferal species and endangering their survival. By K-T time a large impact may have triggered the demise of this already extinction prone species group.

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