

Impacts, volcanism and mass extinction: random coincidence or cause and effect?

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Large impacts are credited with the most devastating mass extinctions in Earth's history and the Cretaceous–Tertiary (K/T) boundary impact is the strongest and sole direct support for this view. A review of the five largest Phanerozoic mass extinctions provides no support that impacts with craters up to 180 km in diameter caused significant species extinctions. This includes the 170 km-diameter Chicxulub impact crater regarded as 0.3 million years older than the K/T mass extinction. A second, larger impact event may have been the ultimate cause of this mass extinction, as suggested by a global iridium anomaly at the K/T boundary, but no crater has been found to date. The current crater database suggests that multiple impacts, for example comet showers, were the norm, rather than the exception, during the Late Eocene, K/T transition, latest Triassic and the Devonian–Carboniferous transition, but did not cause significant species extinctions. Whether multiple impacts substantially contributed to greenhouse warming and associated environmental stresses is yet to be demonstrated. From the current database, it must be concluded that no known Phanerozoic impacts, including the Chicxulub impact (but excluding the K/T impact) caused mass extinctions or even significant species extinctions. The K/T mass extinction may have been caused by the coincidence of a very large impact (> 250 km) upon a highly stressed biotic environment as a result of volcanism. The consistent association of large magmatic provinces (large igneous provinces and continental flood-basalt provinces) with all but one (end-Ordovician) of the five major Phanerozoic mass extinctions suggests that volcanism played a major role. Faunal and geochemical evidence from the end-Permian, end-Devonian, end-Cretaceous and Triassic/Jurassic transition suggests that the biotic stress was due to a lethal combination of tectonically induced hydrothermal and volcanic processes, leading to eutrophication in the oceans, global warming, sea-level transgression and ocean anoxia. It must be concluded that major magmatic events and their long-term environmental consequences are major contributors, though not the sole causes of mass extinctions. Sudden mass extinctions, such as at the K/T boundary, may require the coincidence of major volcanism and a very large impact.

KEY WORDS: impact, mass extinction, Phanerozoic, volcanism

INTRODUCTION

Over the past 500 million years of Earth's history (Phanerozoic), five major extinction episodes resulted in the disappearance of a majority of Earth's biota. These are known as the five great mass extinctions and each occurred at or near the end of the period in the Ordovician (Ashgillian), Devonian (Frasnian/Famennian), Permian (Tatarian), Triassic (Norian) and Cretaceous (end-Maastrichtian). Many lesser and geographically more restricted extinction events also occurred, though these are more difficult to document. On a long-term time-scale of millions of years, all but four of the major mass extinctions occurred during times of major volcanic eruptions (Courtilot & Gaudemer 1996; Courtilot *et al.* 2000; Wignall 2001; Courtilot & Renne 2003), most were accompanied by multiple impact events (Grieve 1997; Earth Impact Database 2005), and all coincide with major changes in climate, eustatic sea-level and oxygenation of the water column (Hallam & Wignall 1997). This consistent association is a first-order

test favoring some direct or indirect causal relationship between mass extinctions, major volcanic eruptions, large impacts and major climate and environmental changes.

With so many potential causes it should be no surprise that the five major mass extinctions have been the subject of heated debates in the scientific community for over 25 years. Many geologists and palaeontologists have long-favoured intrinsic causes, such as plate-tectonic activity (Rich *et al.* 1986; Smith & Pickering 2003; Cañón-Tapia & Walker 2004), climate and sea-level changes, or greenhouse warming accompanied by fluctuating anoxia and/or nutrient dynamics leading to decimation of largely shallow water organisms (Walliser 1996; Hallam & Wignall 1997; House 2002; Racki *et al.* 2002). MacLeod (2003) concluded that plate-tectonic activity is the primary control on extinction and diversification patterns at all time-scales, superimposed by shorter term tectonically influenced environmental perturbations (e.g. sea-level and climate

changes, large igneous provinces). This conclusion echoes the earlier study by Rich *et al.* (1986) based on the correlation of sea-floor spreading rates and marine taxonomic diversity. But not all mass extinctions have the same footprints. McGhee *et al.* (2004) concluded that crises in biodiversity and ecological severity are decoupled and that ecological crises only lead to severe mass extinctions if the dominant or keystone taxa are eliminated—as was the case at the Cretaceous–Tertiary (K/T) boundary. These studies make convincing arguments of the tectonic and environmental effects on biotic diversity on both long and shorter time-scales.

Volcanologists and palaeontologists have advocated the global devastation by massive volcanic eruptions causing extinctions by poisoning and eutrophication, exacerbated by climate change, with the most recent debates centred on continental flood-basalt provinces and large igneous provinces (McLean 1985; Officer *et al.* 1987; Courtillot *et al.* 1986; Courtillot 1999; Kerr 1998; Racki 1999a, b; Ray & Pande 1999; Robock 2000; Wignall 2001; Courtillot & Renne 2003; Mather *et al.* 2004; Vermeij 2004). But no debate has captured the imagination of scientists and public alike as the theory that an extraterrestrial bolide impact was the cause of the K/T mass extinction (Alvarez *et al.* 1980), and that by extension similar impacts possibly caused all or most major mass extinctions. Hybrid hypotheses have tried to link mass extinctions, volcanism and impacts, with the latter triggering large-scale magmatism (Rampino & Stothers 1988; Stothers & Rampino 1990; Jones *et al.* 2002; Alvarez 2003). Ivanov and Melosh (2003) concluded that large impacts do not initiate volcanic eruptions, although location of such impacts on thin geothermally active oceanic crust may result in major volcanism (Glikson 2004a). Most recently, Morgan *et al.* (2004) proposed a controversial theory that sudden mantle plume-induced gas explosions (Verneshots) may account for the association of flood basalts and mass extinctions and could also explain the various signals currently attributed to impacts, such as microtektites, iridium anomalies, shocked quartz and even large craters—a theory contradicted by field observations and experimental data Glikson (in press).

None of these hypotheses have been entirely convincing, in part because they could not account for some critical aspects of the empirical record, such as the selective nature and variable rates of extinctions, the appearance of stepwise extinctions, and the timing between impacts, volcanism and mass extinctions. A most vexing problem has been that of determining the correspondence between mass extinctions, impacts and volcanism. This is largely due to the fact that generally not all three records can be observed within the same stratigraphic sequence, or the sedimentary record is incomplete. Hence correspondence must be inferred by stratigraphic correlation, which often lacks the necessary time resolution, or by radiometric dates with large error bars. In practice, this has led some workers to claim cause-and-effect between impacts and mass extinctions where the close stratigraphic proximity is merely the result of an incomplete stratigraphic record, or where disparate times-scales suggest overlap. Conversely, a strong belief in the cause–effect scenario (cf.

‘strong expectations syndrome’ of Tsujita 2001), has led some workers to ignore the widely separated mass extinction and impact signals claiming them to be one and the same. Others have implausibly proposed that a lag-time effect of several million years can explain the stratigraphic separation of impacts and mass extinctions, as for example the multiple impacts of the Late Eocene and the stepped extinctions of the Late Devonian (Poag 1997; McGhee 2001).

This review article compares the nature, magnitude and timing of mass extinctions, impacts and massive volcanism (e.g. large igneous provinces and continental flood-basalt provinces) in order to evaluate cause–effect patterns. Because a comprehensive review of the voluminous literature on Phanerozoic mass extinctions is beyond the scope of this study (see reviews in Hallam & Wignall 1997), this review is largely built around critical studies and review articles of the last ten years. Specifically addressed are: (i) the currently available databases on mass extinctions, impacts, and volcanism; these are summarised in a chart to provide an overview of temporal correspondence; (ii) impact-crater size and extinction intensity are compared to derive a biotic catastrophe index to help evaluate the likelihood of impact-caused mass extinctions in the geological record; and (iii) the five major mass extinctions are examined on a case-by-case basis with special emphasis on the stratigraphy and timing of mass extinctions, impacts and volcanic events.

DATABASE

Evaluation of the correspondence between mass extinctions, impacts and large igneous provinces remains a problem despite improved dating (Figure 1). Similar data compilations were earlier published by MacLeod (1998, 2003). Although such evaluations are necessarily limited by the available databases with all their inherent imperfections, this is an opportune time for an update as there have been significant advances in refining the stratigraphy and numerical ages at the stage level [Geologic Time Scale 2004 (GST2004): Gradstein & Ogg 2004]. In addition, many new impact craters have been discovered and the ages and sizes of impact craters more narrowly defined (Earth Impact Database 2005; Glikson & Haines 2005; Haines 2005). Large igneous provinces have also been more narrowly defined with significantly improved dating (Courtillot & Renne 2003). Together these updated sources permit a much more improved correlation between impacts, volcanism and mass extinctions than has been previously possible.

Mass extinctions

Figure 1 summarises the extinction intensity record for the past 600 million years with the five major, as well as minor, mass extinctions clearly marked. In this summary, MacLeod’s (1998, 2003) extinction intensity record, which is based on genus level data by Sepkoski (1994, 1996), is used with some modifications due to recent changes and additions in geological stage levels. For example, the chart in Gradstein and Ogg (2004) has

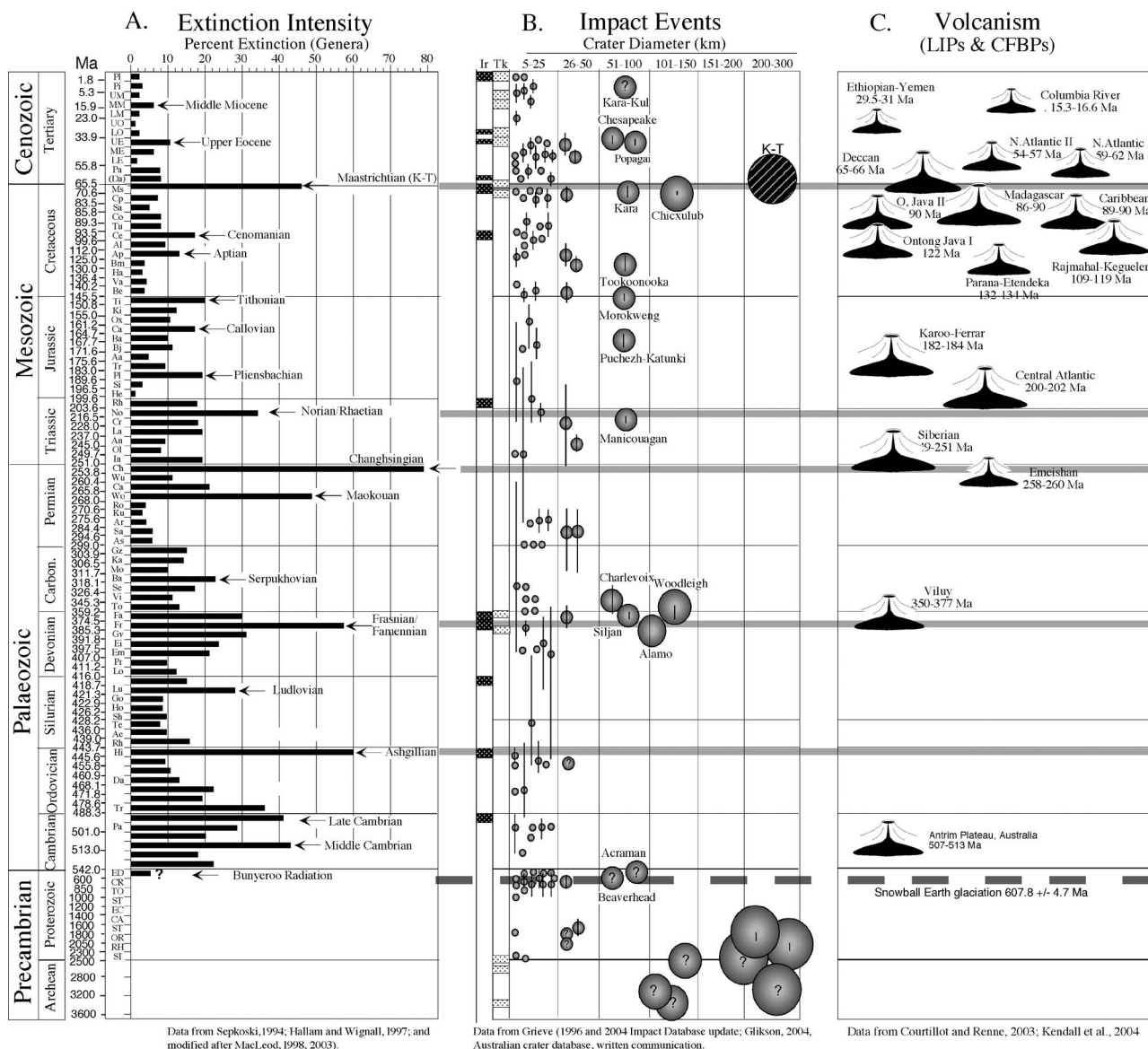


Figure 1 Mass extinctions, impacts and large igneous provinces during the Phanerozoic. Stratigraphic subdivisions and numerical ages from GST2004 (Gradstein & Ogg 2004). The extinction record is based on genus-level data by Sepkoski (1996) and follows an earlier compilation by MacLeod (1998, 2003), with some modification based on Hallam & Wignall (1997). The number of impact events, size and age of craters follows largely the Earth Impact Database (2005).

added a number of new stages, primarily by subdividing old stages. In non-extinction intervals (< 20% generic extinctions), where no updated data exists, the Sepkoski (1994) extinction data has been equally divided among the new stages. For the mass extinction intervals the data were updated based on Hallam and Wignall (1997), Racki (1999a) and Wignall (2001). Here the end-Triassic mass extinction is provisionally divided as one-third Rhaetian and two-thirds Norian (P. Wignall pers. comm. 2004), which reflects the uncertainty in dating and correlation of this mass extinction.

The Permian mass extinction is recognised as two major distinct mass extinctions. The earlier, smaller mass extinction in the middle Permian is generally known as Maokouan and occurred within the Wordian Stage of the end-Guadalupian Epoch (Figure 1) (Hallam & Wignall 1997; Jin & Shang 2000; Wignall 2001). The

largest mass extinction in Earth history occurred at the end of the Permian in the Changhsingian Stage. The two mass extinctions are separated by the Capitanian and Wuchiapingian Stages (265.8–253.8 Ma).

The Precambrian Ediacaran is marked by the Snowball Earth glaciation dated 607 ± 4.7 Ma (Kendall *et al.* 2004) and the Acraman impact crater dated between 590 and 580 Ma (Grey *et al.* 2003). These ecological crises are followed by the Bunyeroo radiation, but the adverse biotic effects of the Acraman impact and glaciation are still poorly understood and therefore not reviewed here.

How well suited is the Sepkoski database to evaluate the correspondence between mass extinctions, major volcanic episodes and impacts? The database is based on genera and stratigraphic stage-level data. This means that all genera that disappeared in a given stage are grouped into one event, regardless of whether they went

extinct at the beginning, middle or late part of the stage. For this reason, no information regarding the nature of the mass extinction, whether gradual or sudden, can be obtained. Moreover, since stages vary in age, and average about 6 million years, the time resolution is very low. At best, this database can identify times of biotic crises intervals and their association with impacts and volcanism, but no cause–effect can be determined. Higher resolution extinction data are available for some areas based on biozonations for all of the major mass extinctions and will be discussed on a case-by-case basis where applicable. McGhee *et al.* (2004) have tried to circumvent this taxonomic problem by estimating the range of mass extinctions based on the ecological ranking of biodiversity crises. But, any cause–effect evaluation between mass extinctions, impacts and volcanism requires high-resolution biostratigraphic data and/or faunal and impact signals within the same stratigraphic sequence; this type of data is currently only available on a global scale for the K/T mass extinction and the Late Eocene impacts.

Numerical ages for geological stages are based on Gradstein and Ogg. (2004), although this new geological time-scale still requires testing. The recent changes in numerical ages and the addition of new stages have introduced some uncertainties in the correlation with the older systems of the literature. As a result, well-known ages of some stage boundaries have changed, including those in the Late Devonian, the Valanginian/Hauterivian boundary and the K/T boundary (the latter from 65.0 to 65.5 Ma), though for the most part these discrepancies are minor on the scale of the correlation chart (Figure 1). To facilitate correlation with the literature we keep the 65.0 Ma age for the K/T boundary, but adjust the larger discrepancies of the older time-scales.

Impact events

The impact database has been growing rapidly with the discovery of many new impact structures, impact ejecta and iridium anomalies. A continuously updated impact record is available online (Earth Impact Database 2005) and for Australia in Haines (2005). These two databases were used with some modifications as discussed below. In Figure 1, the crater sizes are listed based on size categories and the symbols scaled accordingly. Error bars are given where available; crater ages without error bars are marked by question marks because the margin of error is unknown. Most large impacts are relatively well dated with error bars within ± 0.5 to ± 2 million years, though error bars for smaller (< 50 km) and older impacts range from ± 2 to ± 50 million years (Figure 1). Large error bars (> 10 million years) eliminate impact craters from any useful correlation. Just because impact error bars overlap the time of a mass extinction does not prove a cause–effect relationship, it simply means that there is a high probability that the impact occurred sometime within the span of the error bars. To prove cause-and-effect will require close correlation of the time of the impact with the extinction horizon within the same stratigraphic

sequence. Only relatively large impacts (> 50 km in diameter) are considered in this analysis of the biotic effects of impacts, based on the assumption that if large impacts cause no significant extinctions, then the environmental effects from several smaller impacts are not likely to affect species survival rates.

For the Archaean to Palaeoproterozoic impacts the question marks refer primarily to uncertainty in the estimate of the crater size, which is based on mass-balance calculations of iridium and $^{53}\text{Cr}/^{52}\text{Cr}$ isotopes, whereas the age is relatively well-known within limits of error bars (Lowe *et al.* 2003; Kyte *et al.* 2003; Glikson *et al.* 2004). No evaluation of the biotic effects of these impacts is attempted because the biotic record for the Archaean is insufficiently known and possibly unknowable due to lack of shelly faunas preserved in the sedimentary record.

Modifications made to the existing impact crater database are based on more recent studies and prudent assessments of data in the literature. Since impact structures are generally underrepresented with respect to the true impact bombardment of the Earth, the addition of well-documented, though still controversial, craters seems justified. For example, the Alamo breccia of southeastern Nevada is recognised as an impact structure with a predicted maximum 100 km-diameter crater and the age is within the middle Frasnian *Palmatolepis punctata* conodont zone (Sandberg *et al.* 1997, 2002; Warme & Kuehner 1998). The age of the Alamo breccia was thus estimated at about 367 Ma; based on the new time-scale in Gradstein and Ogg (2004) the age is between 385.3 and 382.8 Ma.

There is some controversy regarding the size and age of the Woodleigh impact. The Earth Impact Database (2005) lists Woodleigh as a 40 km-diameter crater with an age of 364 ± 8 Ma, though without direct age measurements. In view of recent studies this listing seems outdated. Detailed studies by Mory *et al.* (2000a, b), Uysal *et al.* (2001) and Glikson *et al.* (2005) determined this as a 359 ± 4 Ma and 120 km-diameter structure based on seismic reflection profiles and the Bouguer anomaly data that indicate a sharp truncation at 120 km. This age and crater size are accepted in this summary. Though contested at first, Reimold *et al.* (2003) acknowledged an impact crater size of about 60–70 km, and Bevan and De Laeter (2002 p. 185) acknowledged a crater size to > 90 km, which is in closer agreement with the 120 km original suggestion of Mory *et al.* (2000a,b).

The recent publication and media hype asserting that the Bedout structure of Western Australia is a large impact crater that may have been responsible for the end-Permian mass extinction (Becker *et al.* 2004) is omitted in this study for lack of corroborating evidence (see below).

The Chicxulub impact structure is listed in the Earth Impact Database (2005) with a crater rim diameter of 170 km. However, as noted by J. Morgan (written communication, March 2005) ‘‘The location of the ‘crater rim’ is irrelevant with respect to the damage the impact may have caused to the environment. What is important is the size of the transient (and excavation) cavity. This is very clear in the seismic data – the diameter is 90 km ± 10 km. This parameter tells you about the energy of

impact, the volume of material excavated, and the volume of pollutants released into the atmosphere.”

The age of 64.98 ± 0.05 Ma, or precisely K/T age, listed in the Impact Database (2005) is the younger of several published $^{39}\text{Ar}/^{40}\text{Ar}$ ages of 65.0, 65.2 and 65.4 Ma based on impact melt glass spherules from the Chicxulub impact (Izett *et al.* 1991; Sharpton *et al.* 1992; Swisher *et al.* 1992; Dalrymple *et al.* 1993) and may have been picked because it is closest to the age of the K/T boundary mass extinction.

It is critical that the age and size of the Chicxulub impact crater is estimated as accurately as possible because this is believed to have been the impact that caused the K/T mass extinction. New stratigraphic data has shown that Chicxulub pre-dates the K/T boundary by about 300 000 years both in sections in northeastern Mexico (Keller *et al.* 2002, 2003) and in the new Chicxulub crater corehole Yaxcopoil 1 (Keller *et al.* 2004a, b). Although there is still some dispute, the Chicxulub impact is considered as pre-K/T age in this summary (see discussion below). A second larger impact at the K/T boundary is inferred from the global iridium distribution associated with the mass extinction, though an impact structure has yet to be identified.

Volcanism: large igneous provinces

The database for large igneous provinces and continental flood-basalt provinces has been recently updated and ages refined by Ernst and Buchan (2001), Abbott and Isley (2002) and Courtillot and Renne (2003). Relatively narrow age limits have been determined for the major eruption episodes, and some major events have been separated into two different phases (e.g. Ontong Java I and II, North Atlantic I and II: Courtillot & Renne 2003). It is important to note that these new dates generally encompass the most intensive phases of the eruptions. In Figure 1 this database is used with the symbols scaled for size and age duration of each province given. It is particularly difficult to evaluate the biotic effects of volcanism (see discussion below) and this chart can only serve to indicate times of major volcanic activity, many of which coincide or span across mass extinction episodes.

TEMPORAL CORRESPONDENCE

At the gross multi-million year time-scale of Figure 1, the mass extinction – impact–volcanism database shows an overall correspondence between large impacts and three of the five major mass extinctions [end-Cretaceous, end-Triassic, mid-Late Devonian Frasnian–Famennian (F/F)]. Four of the five major mass extinctions (all except the end-Ordovician) also show a correspondence with major volcanic eruptions, and as many as five of the minor extinction episodes. But only two intervals (Devonian F/F and end-Cretaceous) show a correspondence between large impacts and major volcanic eruptions, indicating that impacts are not likely triggers of large igneous provinces. A plot of ages of major volcanic eruptions versus ages of most major and minor mass extinctions and oceanic anoxic events

by Courtillot and Renne (2003) showed this correlation well. When impact events are added to this plot it becomes obvious that environmental upheavals associated with major volcanic eruptions and impacts mark almost all major and minor mass extinctions, at least at this gross geological time-scale (shaded intervals in Figure 2). But it would be premature to infer a cause–effect relationship based on this dataset because the age resolution is inadequate. For this reason, the five major mass extinctions are examined on a case-by-case basis with special emphasis on the stratigraphy and timing of events.

A graph of mass extinction intensity versus the major volcanic events shows that although most major and minor mass extinctions occurred at times of major volcanic eruptions, the magnitude of extinctions at the genus level is relatively low, except for the end-Cretaceous, end-Permian and Devonian F/F mass extinctions (Figure 3). This suggests at least two possibilities: (i) only the largest volcanic eruptions caused sufficient biotic stress leading to major (> 40% genera) extinctions (e.g. end-Permian and end-Cretaceous); or (ii) large volcanic eruptions coupled with major impacts lead to mass extinctions (e.g. K/T boundary). The likelihood of either possibility can be assessed based on higher age resolution. Nevertheless, determination of the biotic effects of large igneous provinces remains a problem despite improved dating, though it has narrowed down the age range in which biotic effects related to volcanism can be evaluated. In addition to age constraints, various other factors should be taken into consideration in determining causal links between volcanism and biotic effects, including the nature of volcanism, whether subaerial or the less-detrimental submarine eruptions (Coffin & Eldholm 1994; Courtillot 1999), whether in high or low latitudes, and the type of volcanism (e.g. carbonatite: Ray & Pande 1999). Some type of volcanic eruptions (e.g. sea-floor eruptions) may have a stimulatory effect on the oceanic biosphere due to nutrient release (Vermeij 1995; Racki 1999a), and only equatorial eruptions of sulfate aerosols may affect atmospheric circulation on a global scale (Robock 2000; Zielinski 2000).

A plot of mass extinction intensity versus crater size shows no direct correlation between impacts and major mass extinctions, and a very poor correlation with minor extinctions (Figure 4). Craters up to 170 km in size do not appear to cause significant extinctions (e.g. > 20%). Only the 52 km-wide Siljan (361 ± 1.1 Ma) crater at the end of the Devonian appears to coincide with elevated extinctions, although this is misleading as it coincides with the aftermath of the F/F mass extinction, which cannot be attributed to this small impact crater. The 120 km Woodleigh crater (359 ± 4 Ma) also spans the last of three extinction crises at the end of the Devonian, whereas the Alamo breccia pre-dates the F/F extinction by at least 5 million years (Figure 1). None of the other impact craters in the 50–100 km size range are associated with significant extinctions. This includes the 90 km and 100 km size Chesapeake and Popigai craters of the Late Eocene. The 145 km Chicxulub impact crater of Yucatan, Mexico, which is now determined to pre-date the K/T mass extinction by about 300 000 years

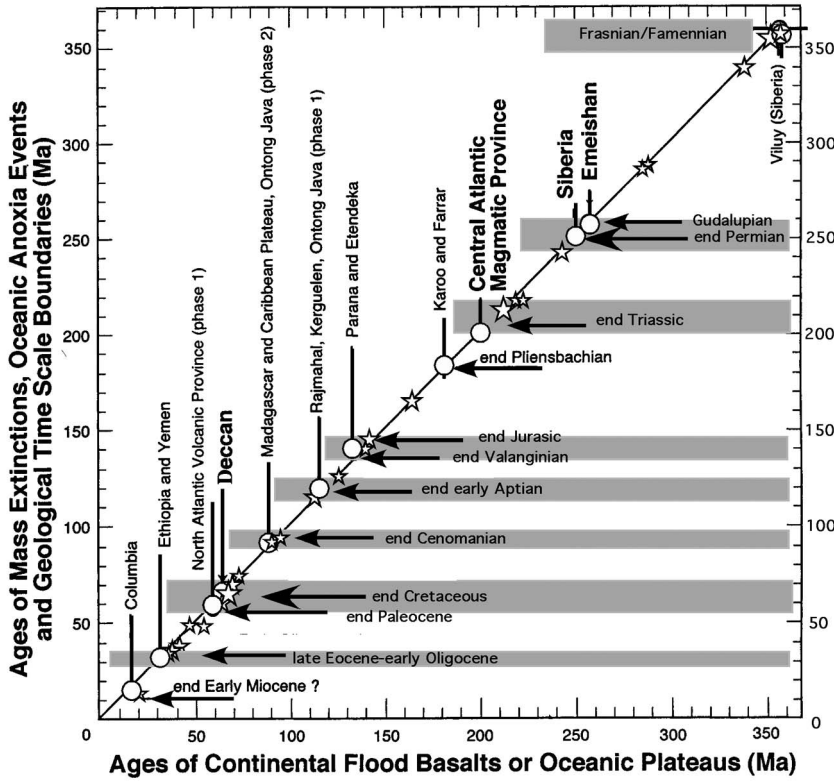


Figure 2 Ages of continental flood basalts and oceanic plateaus and impact craters > 25 km in diameter (modified from Courtillot & Renne 2003). Grey shaded intervals mark time of mass extinctions, circles mark volcanism, stars mark impacts. Note that at this low-resolution (million year) time-scale, several mass extinctions coincide with both impacts and volcanism, but this is not the case when viewed at higher resolution.

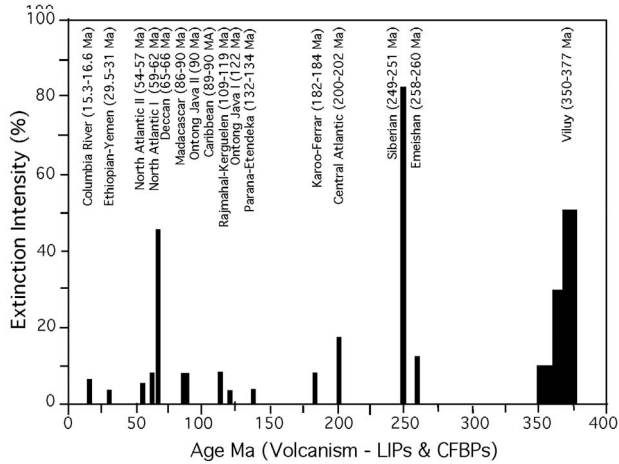


Figure 3 Volcanism vs extinction intensity suggesting that mass extinctions at the Late Devonian, end-Permian and end-Cretaceous may have been significantly influenced by the environmental effects of large igneous provinces (LIPs) and continental flood-basalt provinces (CFBPs).

(Keller *et al.* 2003, 2004a, b), is not associated with significant extinctions even in the vicinity of the impact locality (Keller *et al.* 2002). This indicates that impacts up to 10 km-diameter and creating craters up to at least 170 km-diameter caused no significant global extinctions (Figure 4).

What size impact is required to cause or trigger a major mass extinction? The K/T boundary mass extinction is associated with both volcanism and a major

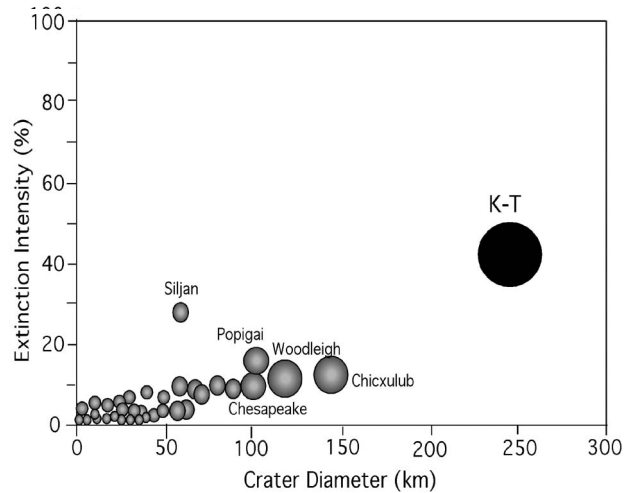


Figure 4 Extinction intensity vs crater diameter of known impacts showing no correlation between mass extinctions (> 30% diversity decline) and impact craters up to 150 km in diameter. The 52 km-wide Siljan crater occurred at the end of the Famennian. Note that the Chicxulub impact crater pre-dates the K/T boundary and is not associated with the mass extinction. A K/T impact crater of > 250 km-diameter is inferred based on this plot and the global iridium anomaly at the K/T mass extinction.

impact. This impact had to be much larger than Chicxulub, as indicated by the global iridium anomaly and the mass extinction. In contrast, no iridium anomaly has been detected in association with the

Chicxulub impact, though the K/T iridium is frequently and erroneously attributed to it (Keller *et al.* 2003). If the K/T boundary mass extinction was caused by a large impact, then the size of the crater must be at least 250–300 km in diameter. Such a large crater of K/T age has yet to be found, which suggests it may have been an oceanic impact. The Amirante Basin of the Indian Ocean and/or the Shiva crater in India (Chatterjee 1997) have been suggested as potential candidates, but hard evidence remains elusive.

CALIBRATING BIOTIC EFFECTS OF IMPACTS

It is almost routinely asserted that the K/T mass extinction was the result of the Chicxulub impact and that other mass extinctions were likely caused by large impacts also even though no evidence connects major extinctions to impacts, except for the K/T boundary (Alvarez 2003). Indeed it seems common sense that an impact with the force to wipe out life over an area the size of France could also cause mass extinctions on a global scale. But how large an impact is really required to cause a global mass extinction? Calibration of the biotic effects of impacts requires one or more sedimentary sequences that contain (i) high-resolution age control, (ii) unequivocal impact signals, and (iii) a well-preserved faunal record. These conditions are met for the Late Eocene and K/T boundary transition and are used here to calibrate the biotic effects of impacts with craters 90–100 km and 170 km wide, respectively.

Late Eocene

Ever since the discovery of multiple impacts in the Late Eocene by Keller *et al.* (1983) based on at least three microtektite layers and an iridium anomaly, the Late Eocene has proven a fertile playground for discovering evidence of multiple extraterrestrial events. Two well-dated large (90–100 km wide) impact craters, Popigai and Chesapeake, and three small craters have been identified in addition to the microtektite horizons. All occurred within a period of about 300 000 years. Montanari and Koeberl (2000) dated the 100 km-wide Popigai crater of Russia at 35.7 ± 0.2 Ma based on impact melt rock, and the 95 km-wide Chesapeake Bay (USA) crater at 35.6 ± 0.2 Ma based on biostratigraphic data [35.5 ± 0.3 in Earth Impact Database (2005) and between 36 and 35.7 Ma in Harris *et al.* (2004)]. The smaller 28 km-wide Mistastin crater of Newfoundland/Labrador, Canada, is dated at 36.4 ± 0.4 Ma, the 7.5 km-wide Wanapitei crater of Ontario, Canada, at 37.2 ± 1.2 Ma, and the poorly dated 17 km-wide Logoisk crater of Belarus at 40 ± 5 Ma (Figure 5) (Masaitis *et al.* 1980). These smaller impacts are not likely to have left recognisable signatures in the rocks.

The impact stratigraphy of the Massignano section of Italy has been intensely studied and discussed by Montanari and Koeberl (2000). Two iridium anomalies were identified with peaks of 0.33 ppb and 1 ppb, though a subsequent higher resolution study identified peaks of 1.5 ppb and 2.5 ppb (Bodiselsch *et al.* 2004). The smaller anomaly is associated with flattened microkrystites, as

well as shocked quartz and Ni-rich spinels and interpreted to represent the Popigai impact (Figure 5a). A broad increase in extraterrestrial ^3He , representing an influx of interplanetary dust particles, was detected over an interval of about 2.2 million years (36.2–34.2 Ma) and interpreted as evidence of a comet shower (Farley *et al.* 1998).

One would expect such a comet shower (Hut *et al.* 1987) to have serious biotic consequences, but there are no species extinctions. The Late Eocene has long been known as a time of long-term high biotic stress leading to gradually decreasing biodiversity associated with prolonged global cooling related to tectonic and oceanic circulation changes (Keller 1986; Prothero 1994, in press; Wei 1995; Retallack *et al.* 2004). Marine plankton parallel this global cooling with the successive elimination of warm-water low-latitude species. Short hiatuses associated with intensified circulation and erosion, led to a stepwise extinction pattern in planktic foraminifers (Keller 1986). The evidence conclusively indicates that no sudden catastrophic bioevents are associated with this comet shower, and no species extinctions occurred due to either of the large (90–100 km wide) Popigai or Chesapeake impacts. Despite this lack of evidence, Alvarez (2003) concluded that an impact-cause is 'abundantly confirmed' in the case of the Late Eocene by the minor and gradual species extinctions, and Poag (1997) hypothesised that the minor and gradual extinctions over 3 million years are the result of delayed adverse effects of the impacts causing long-term global warming. These claims have remained unsupported by any evidence; Late Eocene impacts are not associated with either long-term or significant short-term climate changes (Bodiselsch *et al.* 2004).

There are some indications of environmental changes, though they may be part of the long-term climatic trends related to fluctuations in high-latitude glaciation. Monechi *et al.* (2000) reported a warming episode well above the iridium layer in the Massignano section as well as changes in trophic resources, and concluded that these were most likely the result of long-term climate changes unrelated to the impacts. Most recently, Bodiselsch *et al.* (2004) concluded that two iridium peaks associated with very brief (1–5 cm) 0.7–1.0 ‰ negative excursions in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ reflect short-term climate warming due to CO_2 -rich comets, or methane hydrate release after the impacts on continental shelves. Methane release is the favoured interpretation for the latest Paleocene short-term warming and 3–4‰ shift in $\delta^{13}\text{C}$, but this interpretation is not credible for the minor isotope variations of the Late Eocene.

While it is not unreasonable to assume that an impact of the size of Chesapeake or Popigai could cause at least short-term warming, the data from the Massignano section are not convincing, as they reveal significant diagenetic alteration and therefore cannot be used to infer temperature changes. This is evident in the covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values, with a correlation of $R^2 = 0.477$ in the two critical intervals (10.0–10.5 m and 6.0–6.4 m: Figure 5b, c). Since diagenesis can alter $\delta^{18}\text{O}$ values by more than 1.0‰,

Bodiseličš *et al.*'s (2004) conclusion that the short-term 0.7–1.0‰ negative $\delta^{18}\text{O}$ excursions associated with the two iridium peaks reflect impact-induced warming is not justified. Moreover, there is a $\delta^{18}\text{O}$ excursion of similar magnitude preceding the lower iridium peak with no evidence of impact (Figure 5b), which suggests that these excursions may be part of normal environmental and/or diagenetic variations.

The idea of short-term climate warming due to the Late Eocene impacts is intriguing, but requires investigation of diagenetic effects. In addition, small-scale $\delta^{18}\text{O}$ variations must be viewed in the context of well-established long-term climate trends during the Late Eocene. The progressive, or stepwise extinctions observed through the Middle and Late Eocene are closely related to this long-term climate record. Given the well-studied and well-documented Late Eocene impacts and environmental changes, we can conclude that impacts with crater diameters of up to 100 km do not cause mass extinctions or any significant species extinctions, nor do they cause significant short-term or long-term climate and environmental changes.

K/T boundary transition

The internationally accepted criteria for identifying the K/T boundary are illustrated in Figure 6 based on the official K/T Global Stratotype Section and Point accepted by the International Commission on Stratigraphy at El Kef, Tunisia (Cowie *et al.* 1989; Keller *et al.* 1995; Remane *et al.* 1999). The El Kef section was chosen for its continuous deposition and expanded sedimentary record with excellent preservation of calcareous microfossils, geochemical and mineralogical marker horizons, absence of disconformities, hard grounds or any other breaks in sedimentation across the K/T boundary. At El Kef the upper Maastrichtian consists of monotonous grey marls with an abrupt change to a 50 cm-thick black organic-rich clay with a 2–3 mm-thick red layer enriched in iridium, nickel-rich spinels, clay and pyrite spherules at its base (Robin *et al.* 1991). This lithologic shift is characterised by a 2–3‰ negative excursion in $\delta^{13}\text{C}$ (Keller & Lindinger 1989). All tropical and subtropical species disappeared at or below the K/T boundary and the first new Tertiary species evolved immediately after the boundary event. These markers identify the K/T boundary worldwide, although the thickness of the boundary clay is highly variable.

The K/T boundary is thus easily recognised in sediment strata worldwide, including Mexico where the iridium-rich red layer is also present and the boundary clay is only a few centimetres thick (Figure 7). Age control is excellent based on microfossils, particularly planktic foraminifers and calcareous nannofossils and the biotic effects of the mass extinction are well documented in hundreds of sections worldwide, as shown for northeastern Mexico (Figure 8). Planktic foraminifers suffered the most severe mass extinction of all marine micro- and macro-invertebrates, which makes them a particularly good group to test the biotic effects of catastrophes (Keller 1996, 2001).

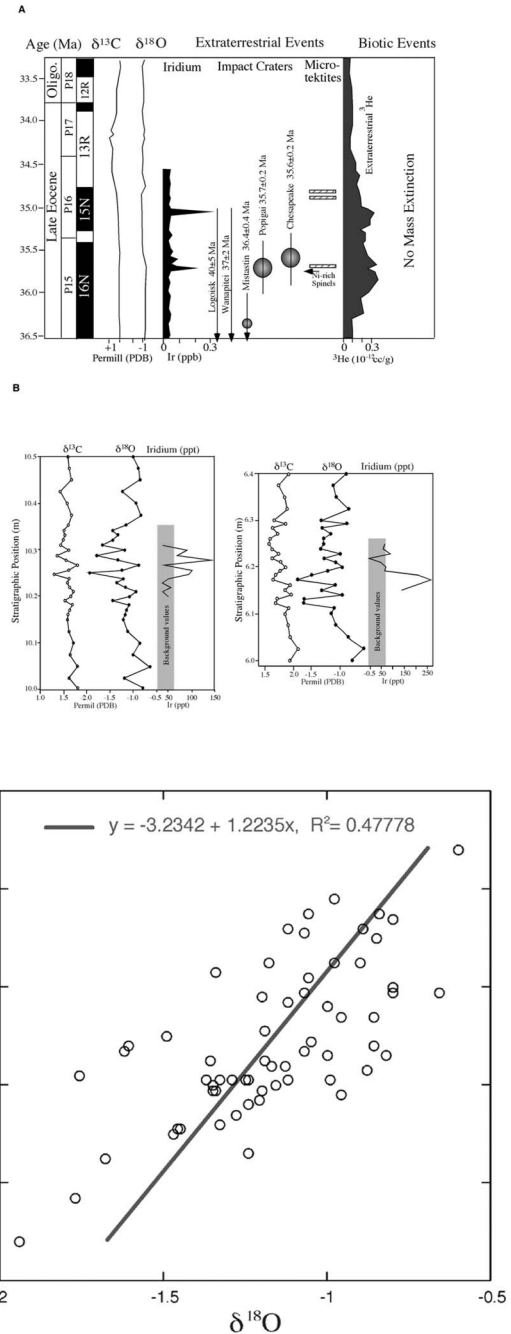


Figure 5 (a) Summary of Late Eocene impact events, including craters, iridium anomalies, ^3He and microtektites (data from Keller *et al.* 1983; Montanari & Koeberl 2000). There are no impact-related species extinctions in any microfossil groups during the Late Eocene. (b) Covariance of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of bulk carbonate across the interval with elevated iridium concentrations and extraterrestrial ^3He at the Massignano section, Italy, reflects diagenetic alteration, rather than impact-caused climate warming of up to 8°C as interpreted by Bodiseličš *et al.* (2004). (c) There is a good correlation ($R^2 = 0.477$) between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of bulk carbonate from the Massignano section, Italy, which reflects a strong diagenetic overprint, rather than a temperature signal (data from Bodiseličš *et al.* 2004).

By K/T boundary time, all tropical–subtropical species went extinct, or two-thirds of the total species diversity (Figure 8). The extinction coincides precisely with the global iridium anomaly, and suggests that a large impact may have been directly responsible for their demise, though Deccan volcanism may have contributed. Most non-tropical species suffered declines in diversity as well as body size and never recovered prior to their extinction in the Early Paleocene. Calcareous nannofossils suffered a similar decline in diversity leading to eventual extinction of many species (Gartner 1996; Henriksson 1997; Gardin & Monechi 1998; Gardin 2002). There are no significant extinctions prior to the K/T boundary, though major biotic stress (e.g. dwarfing, decrease in abundance) is associated with the global greenhouse warming between 200 000 and 400 000 years prior to the K/T boundary (Abramovich & Keller 2002; Keller 2002). Macro-invertebrates (e.g. ammonites, inoceramids, rudistids: Johnson & Kauffman 1996; Zinsmeister & Feldmann 1996; Tsujita 2001; Steuber *et al.* 2002) suffered their declines well before the K/T boundary, primarily during the late Maastrichtian global cooling followed by greenhouse warming. Planktic foraminifers also experienced major stress at this time. The extinction pattern of dinosaurs is more difficult to ascertain because the fossil record is sparse and often lacks the necessary time control, but a general decrease in diversity during the Late Cretaceous is well documented (Archibald 1996).

IMPACT EJECTA IN NORTHEAST MEXICO PRE-DATES K/T

It is generally assumed that the Chicxulub crater represents the K/T impact. The search in support of this hypothesis has led workers in the early 1990s to northeast Mexico where numerous outcrops reveal thick siliciclastic deposits between the K/T iridium anomaly and the Chicxulub impact ejecta (microtektites). To tie the microtektites to the Ir anomaly above the siliciclastic units, Smit *et al.* (1992, 1996) and Smit (1999) hypothesised that the Chicxulub impact generated a giant tsunami that deposited the siliciclastic sediments after the fallout of the microtektites, but prior to the fallout of the iridium. Although this

hypothesis has been generally accepted over the past decade, there is little if any evidence to support it (Keller *et al.* 2003).

The classic El Penon section is a good example to illustrate the problems with this impact–tsunami hypothesis. At this locality, an 8 m-thick siliciclastic sequence consists of thick sandstone underlying several metres of alternating sand, silt and shale (Figure 9). Similar siliciclastic deposits can be observed in over 40 outcrops examined to date in northeastern Mexico: all of them were deposited in submarine channels and are thickest at the centres of the channels and disappear laterally over 100–200 m. Microtektites invariably underlie these siliciclastic deposits. The K/T boundary mass extinction and Ir anomaly are always above it (Figures 7, 8).

The first problems that surfaced with the impact–tsunami interpretation were the presence of burrows in

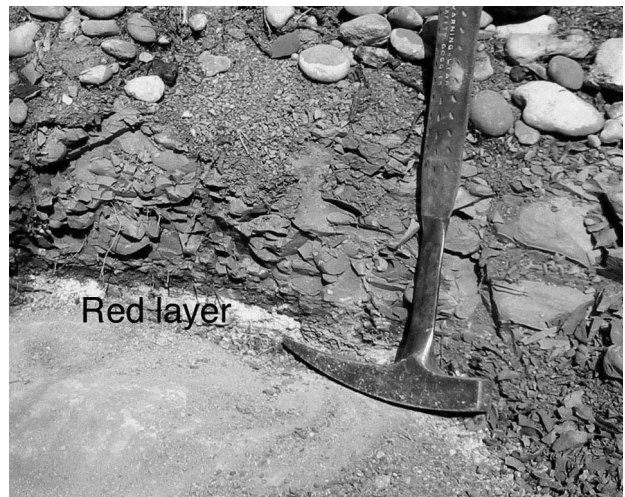
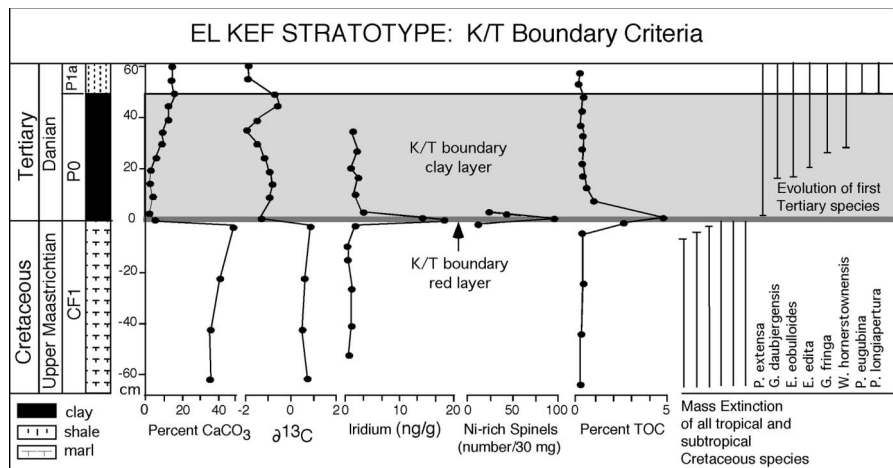


Figure 7 K/T boundary red clay layer and Ir anomaly at the La Sierita section of northeast Mexico. This red clay layer, which marks the K/T mass extinction worldwide, is stratigraphically several metres above the microtektite layer that marks the Chicxulub impact in the late Maastrichtian.

Figure 6 K/T boundary defining criteria based on the El Kef stratotype. The K/T boundary is defined by the extinction of tropical and subtropical species, the first occurrence of Tertiary species, a dark organic-rich boundary clay with thin red layer at its base that is enriched in iridium, nickel-rich spinels, pyrite and rare clay spherules, a negative $\delta^{13}\text{C}$ excursion, high TOC and low CaCO_3 .



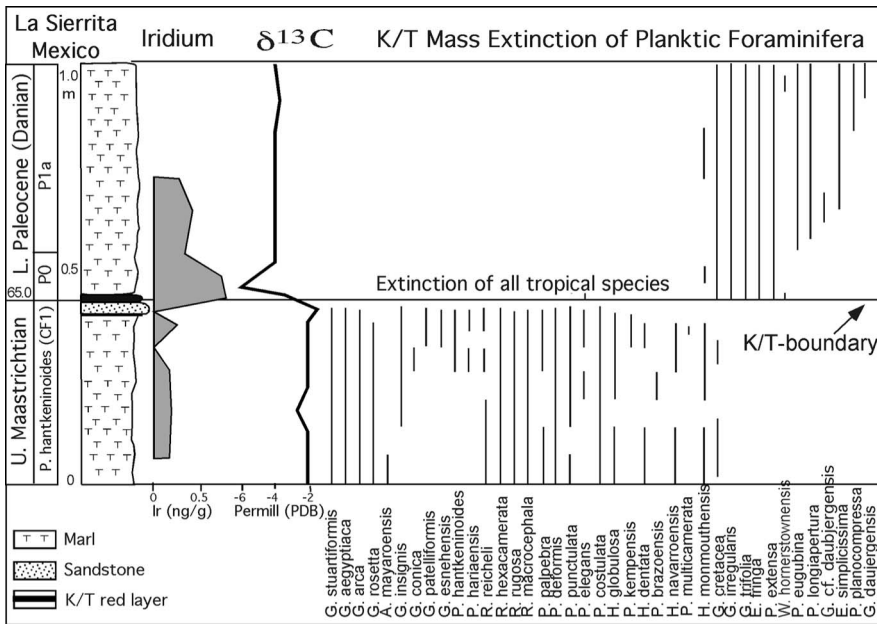


Figure 8 Mass extinction of all tropical Cretaceous planktic foraminifers at or near the K/T boundary, Ir anomaly and $\delta^{13}\text{C}$ shift in the La Sierrita section of northeast Mexico. The microtektite layer that marks the Chicxulub impact is stratigraphically several metres below in the late Maastrichtian.

the sandstone and the alternating sand, silt and shale layers above it (Figure 10a). These burrows of *Chondrites*, *Thalassinoides* and *Ophiomorpha* reveal the repeated colonisation of the ocean floor during deposition of these siliciclastic units (Keller *et al.* 1997; Ekdale & Stinnesbeck 1998). Moreover, the shale interlayers are finely laminated, rich in foraminifers, and burrowed (Figure 10b), which indicates there were intermittent quiet periods of normal marine sedimentation lasting many seasons, alternating with rapid influx of sand. This is inconsistent with either the high-energy deposition of a mega-tsunami, or the short time duration of any tsunami event. As early as the middle 1990s it seemed that rather than a mega-tsunami, these siliciclastic deposits recorded the normal ebbs and flows of channelised transport related to sea-level changes and slope instability.

But more inconsistencies surfaced early on in the microtektite deposits beneath the siliciclastic units. In many localities throughout northeast Mexico there are two layers of microtektites separated by a 20 cm-thick sandy limestone (Figure 11). This limestone contains microtektites at the base and top, and there are microtektite-filled J-shaped burrows (Figure 11b) (Ekdale & Stinnesbeck 1997; Keller *et al.* 1997, 2002, 2003). Deposition of this sandy limestone layer represents a lengthy interruption in microtektite deposition and a later resumption. It means that the two microtektite layers and the limestone in between them could not have been deposited during the few hours between the Chicxulub impact and the arrival of the tsunami waves, as suggested by Smit *et al.* (1996) and Smit (1999). Nevertheless, these authors reconciled this problem with the tsunami hypothesis by postulating the injection of the limestone layer by seismic shaking as a result of the Chicxulub impact. But the problem with this interpretation is that the same limestone layer can be traced between the two spherule layers over 300 km. Seismic injection of a 20 cm-thick limestone layer over

such a distance and within the same spherule layer is highly unlikely if not impossible.

Another problem with interpreting these two microtektite layers as the original Chicxulub fallout is the content itself. In addition to the microtektites, these beds also contain abundant siliciclastic grains (Figures 11a, 12a–c), reworked shallow-water microfossils, wood and plant debris. This indicates that they were originally deposited in a shallow nearshore area and subsequently reworked and transported via submarine channels into deeper waters (the depth of deposition at El Penon is > 500 m). This had to happen in two separate reworking and transportation events separated by the time it took to deposit the sandy limestone layer. To validate the impact–tsunami hypothesis, the following series of events would have had to take place: the microtektite fallout from the Chicxulub event, followed by the first reworking and transport, followed by deposition of the limestone layer, followed by the second reworking and transport. All this would have occurred within the few hours prior to the arrival of the mega-tsunami. This is physically impossible. Based on these observations, Keller *et al.* (1997) concluded that the Chicxulub impact must pre-date the K/T boundary and could not be responsible for the mass extinction or the iridium anomaly.

More evidence to support this conclusion was uncovered by a subsequent study of the late Maastrichtian Mendez Formation underlying the siliciclastic deposits in northeast Mexico (Stinnesbeck *et al.* 2001; Keller *et al.* 2002). These studies uncovered additional microtektite layers interbedded in marls of the planktic foraminiferal zone CF1, which spans the last 300 000 years of the Maastrichtian (Keller 2003). At El Penon the oldest microtektite layer is about 1.8 m-thick and interbedded in marls 5 m below the two reworked microtektite layers and sandy limestone (Figure 13). This appears to be the original Chicxulub ejecta deposit, as indicated by an abundance of



Figure 9 At El Mimbral in northeast Mexico a siliciclastic deposit consisting of massive sandstone and alternating sand, silt and shale layers lies between reworked impact spherules (microtektites) at the base and the K/T boundary at the top. To reconcile the age of the microtektites with the K/T boundary, these sediments have been interpreted as deposited by a mega-tsunami within hours to a few days of the Chicxulub impact.

elongate, compressed and partially fused glass spherules with concave/convex contacts (Figure 12), interstitial calcite cement, only very rare marl clasts, absence of microfossils and any reworked shallow-water material (e.g. sand, wood, plant debris, shallow-water benthic foraminifers). Biostratigraphy indicates that these microtektites are near the base of biozone CF1, and within the lower part of calcareous nannofossil zone *Micula prinsii*, and the lower part of magnetochron 29R. These biostratigraphic and palaeomagnetic data, as well as extrapolation from sediment-accumulation rates, indicate that the Chicxulub impact pre-dates the K/T boundary by about 300 000 years (Keller *et al.* 2003).

In a recent debate on the age of Chicxulub sponsored by the Geological Society of London (<http://www.geol-soc.org.uk/template.cfm?name=NSG1>), Smit claimed that the microtektite layers interbedded in late Maastrichtian marls are the result of slumping induced by the Chicxulub impact. However, in all of northeastern

Mexico there is no significant disturbance of the original sedimentary layers and there is no evidence of slumping. In over 40 outcrops only two minor (2–3 m) folds related to internal deformation could be documented in one area (Mesa Juan) (Soria *et al.* 2001; Schulte *et al.* 2003). But there are many additional reasons why the microtektite layers in the marls of the Mendez Formation are not the cause of tectonic–seismic disturbance including: (i) the marls overlying the microtektite deposit are undisturbed; (ii) two localities 30 km apart show similar successions (Loma Cerca and El Penon: Keller *et al.* 2003); (iii) microfossils show the normal succession of species abundances for the late Maastrichtian, rather than the chaotic distribution that is characteristic in slump deposits (Keller *et al.* 2002); (iv) the microtektite layers are distinctly different from those below the siliciclastic deposit in that they lack any shallow-water components (Figure 12); and (v) the base of the microtektite layer consists of welded vesicular glass and compressed microtektites with concave–convex contacts that indicate deposition occurred while the glass was still hot and malleable (Figure 12).

All of these factors point to rapid deposition of microtektites into deep water immediately after the Chicxulub impact 300 000 years before the K/T boundary mass extinction. After deposition, normal pelagic sedimentation resumed during the late Maastrichtian for at least 200 000 years. The so-called tsunami deposits of Smit *et al.* (1992, 1996) can be explained by normal depositional events related to sea-level and tectonic activity. For example, more than 200 000 years after the Chicxulub impact, the shallow-water microtektites were exposed to erosion and transported into deeper waters during at least two different times (separated by deposition of the sandy limestone), either as a result of lower sea-levels and/or tectonic activity related to the uplift of the Sierra Madre Oriental in northeastern Mexico. Slope instability and continued erosion and transport via submarine channels continued for some time and resulted in deposition of the so-called tsunami deposits of Smit *et al.* (1992, 1996).

CHICXULUB IMPACT BRECCIA PRE-DATES K/T

The new corehole Yaxcopoil 1, drilled in the Chicxulub crater, was supposed to prove that Chicxulub is the K/T boundary impact and responsible for the mass extinction. Instead, it caused renewed controversy over the age of the Chicxulub crater and whether impacts the size of Chicxulub can cause mass extinctions. The new Yaxcopoil 1 core recovered a 100 m-thick impact breccia, which overlies subtidal platform carbonates and is separated from the K/T boundary by a 50 cm-thick micritic limestone (Figure 14). Smit *et al.* (2004) interpreted this 50 cm-thick layer as high-energy backwash and crater infill following the Chicxulub impact and hence contemporary with the K/T boundary mass extinction. Keller *et al.* (2004a, b) interpreted the same interval as evidence of *in situ* normal marine deposition over a period of about 300 000 years following the Chicxulub impact, and hence supporting a pre-K/T age for this impact crater. The arguments for each side are outlined below. The critical evidence is based on the

a



b

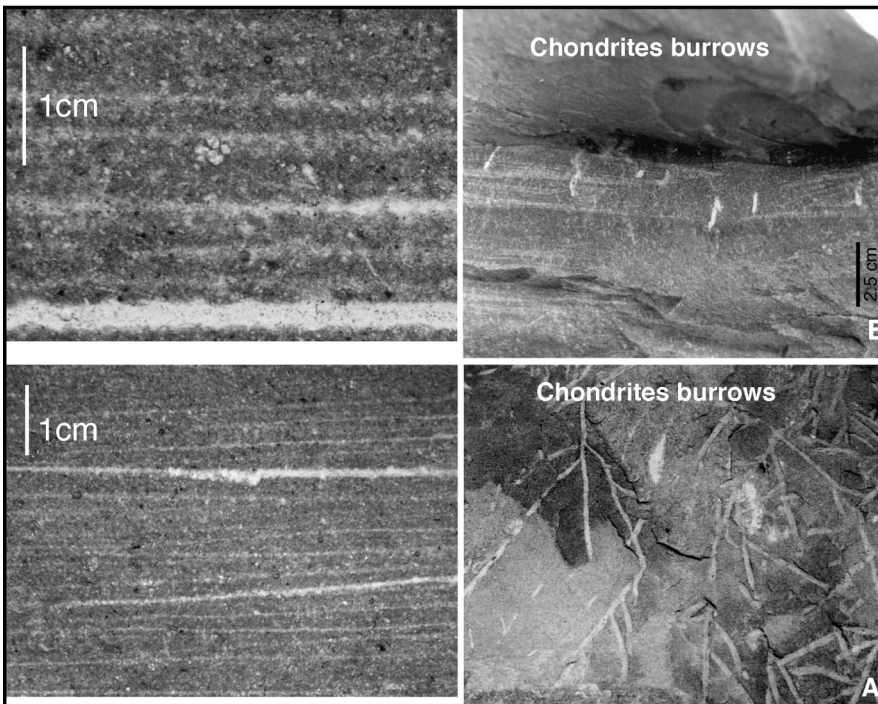


Figure 10 (a) Alternating thin layers of sand, silt and shale are present in the upper part of the siliciclastic deposit beneath the K/T boundary at El Penon (and throughout north-east Mexico). Many of these layers are burrowed by *Chondrites*, *Thalassinoides* and *Ophiomorpha* (see inset). A mega-tsunami interpretation of these deposits cannot be reconciled with the burrows and finely laminated shales, which indicate normal marine sedimentation and repeated colonisation of the ocean floor over an extended period of time. (b) Finely laminated shale layers alternate with thin sand layers in the upper part of the siliciclastic deposit. These laminated shales are rich in foraminifers and bioturbated by *Chondrites* (A, horizontal view, B, vertical view). These laminated shales indicate deposition in a normal low-energy marine environment, with no indication of rapid high-energy deposition as a result of an impact generated megatsunami event.

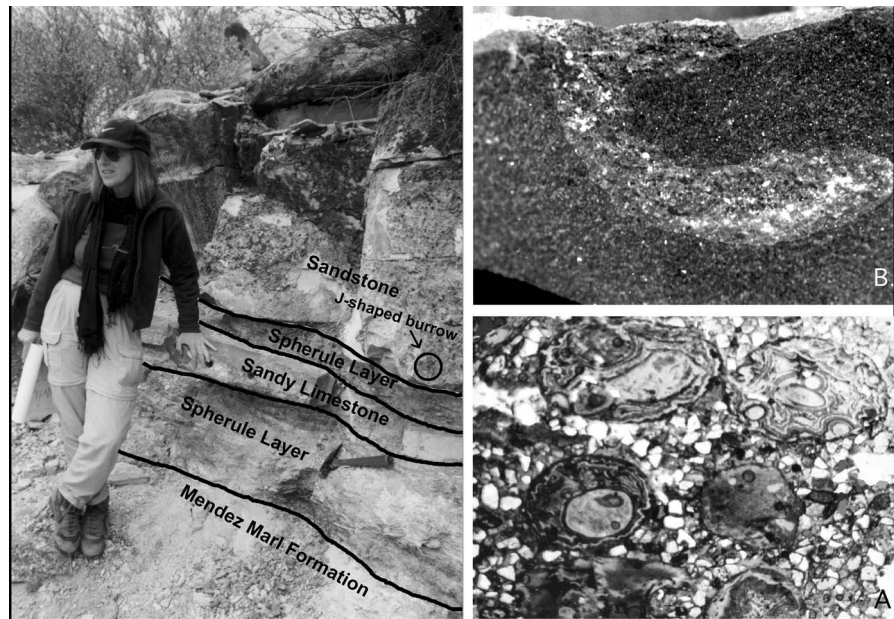
lithology, mineralogy, magnetostratigraphy, stable isotopes, iridium and planktic foraminifers.

Smit *et al.* (2004) argued that the 50 cm interval shows high-energy backwash and crater infill. But the lithology of this interval consists of finely laminated micritic limestone with microlayers or patches of anhedral dolomite crystals. This indicates deposition under low-energy quiet-water conditions where dolomite formed by diagenetic replacement of the precursor limestone. Two 1 cm layers of oblique bedding suggest slightly agitated waters, but insufficient to support the backwash hypothesis. A slight change in the dip angle in the

lower part suggests compaction/settling of the underlying ejecta material that locally changed the seabed slope, rather than backwash. Glass is very rare and no breccia clasts have been observed, though this should be common if the sediments represent backwash and infill from the surrounding crater walls.

Even more problematic for this interpretation is the presence of five thin green clayey microclast layers (including one at the K/T boundary). Smit *et al.* (2004) first interpreted these as smectite derived from altered impact glass. Altered impact glass is indeed present in the breccia, as indicated by the presence of Cheto

Figure 11 Two spherule (microtektite) layers separated by a 20 cm-thick sandy limestone overlie the marls of the late Maastrichtian Mendez Formation and underlie a thick sandstone layer at El Penon. J-shaped burrows infilled with spherules (B) are present in the limestone and near the base of the overlying sandstone. The spherule layers contain abundant vesicular microtektites (A) in a matrix of siliciclastic grains, reworked shallow-water foraminifers, wood and plant debris. The limestone layer and the burrows indicate deposition over an extended period of time, which is incompatible with the tsunami interpretation of Smit *et al.* (1996).



smectite (Figure 15a). But XRD analysis demonstrates that the five green layers are of glauconite origin (Figure 15a–d: Keller *et al.* 2004a, b). Glauconite forms at the sediment–water interface in environments with very slow detritus accumulation (Chamley 1989). These five green layers therefore indicate long pauses (tens of thousands of years for each layer) with reduced sedimentation, winnowing and small-scale transport by minor current activity. Each of the green layers is also bioturbated, which indicates colonisation of the ocean floor by invertebrates (Figure 15). Both glauconite formation and bioturbation are inconsistent with the backwash and crater infill hypothesis.

The depositional age of the interval between the impact breccia and the K/T boundary can be determined based on magnetostratigraphy, $\delta^{13}\text{C}$ isotopes and planktic foraminifers (Figure 14). Magnetostratigraphy indicates deposition occurred within chron 29r below the K/T boundary. Stable isotope values are consistent with deposition during the late Maastrichtian, followed by the negative excursion into the early Danian. Note that a hiatus at the K/T boundary is the likely cause for the absence of a significant iridium anomaly. Planktic foraminifers reveal a late Maastrichtian zone CF1 assemblage, which spans the last 300 000 years before the K/T boundary (Figure 14). Among these three age indicators, the presence of planktic foraminifers has been the most controversial. Smit *et al.* (2004) argued that none are present and that the images presented as evidence by Keller *et al.* (2004a, b) are merely dolomite rhombs. He further argues that a simple test of the geochemical composition of the dolomite rhombs and the images of the foraminifers will prove that they are one and the same.

This test is misguided. Planktic foraminifers in dolomitic limestones are neither preserved nor do they preserve the original foraminiferal test calcite. Dolomite rhombs form by diagenetic replacement of the precursor limestone, as well as the original

foraminiferal shell calcite. Therefore, the ‘fool proof’ test Smit *et al.* (2004) advocated will only confirm that diagenesis has replaced the original shell calcite, but does not invalidate their presence. Foraminiferal species are identified by their size, morphology, characteristic chamber arrangement, coiling, chamber shape and apertures. These characteristics remain intact even after diagenetic replacement of the original shell calcite. Figure 16 illustrates thin-section images of planktic foraminifers from the critical 50 cm interval above the breccia at Yaxcopoil 1 and compares these with pristine 3D images from El Kef, Tunisia. For each species the chamber arrangement, number of chambers, coil, shape and aperture are the same. These images cannot be mistaken for the dolomite rhombs in Figure 17 as claimed by Smit *et al.* (2004). Furthermore, Arz *et al.* (2004), who originally argued that no planktic foraminifers are present, have now confirmed the presence of late Maastrichtian species, though they interpret them as reworked due to backwash. However, that argument is not supported by the lithology and mineralogy of the glauconite layers, the bioturbation, the absence of breccia clasts and presence of an assemblage characteristic of the last 300 000 years of the Maastrichtian.

The evidence for normal marine sedimentation over a period of several hundred thousand years based on multiple proxies (mineralogy, biostratigraphy, magnetostratigraphy, carbon isotopes) is in agreement with the stratigraphic age (65.3 Ma) of the original microtektite ejecta in northeastern Mexico. The 65.3 Ma age is well within the range of published 65.4–65.0 Ma $^{39}\text{Ar}/^{40}\text{Ar}$ ages of the microtektites (Izett *et al.* 1991; Dalrymple *et al.* 1993; Swisher *et al.* 1992), but provides a more precise age due to the high-resolution biostratigraphic control, the high sedimentation rate and the stratigraphic separation of the microtektite layer from the K/T boundary mass extinction and Ir anomaly.

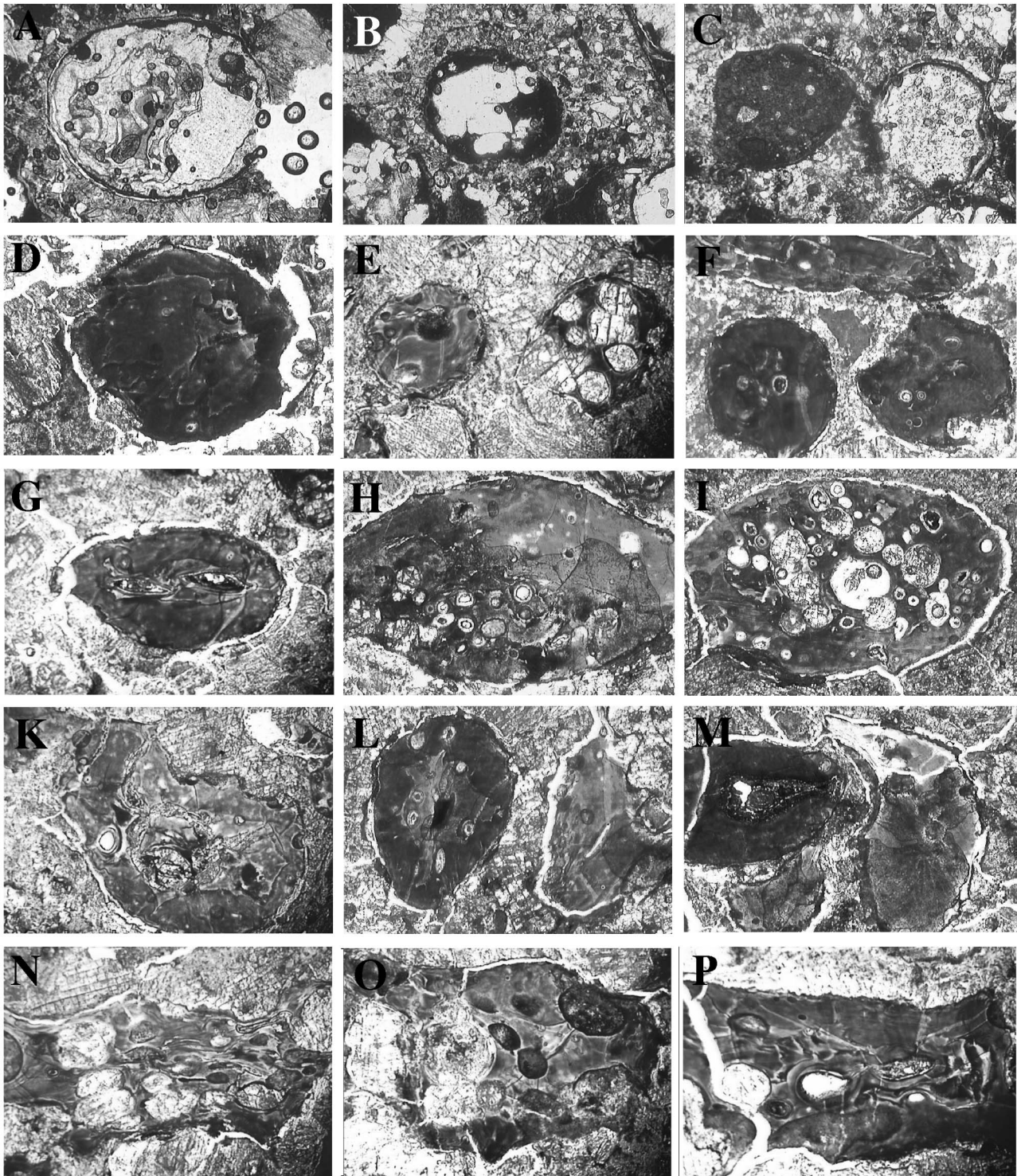


Figure 12 Chicxulub impact spherules (microtektites) from El Penon, NE Mexico. Spherules from the topmost layer (a–c, below the siliciclastic deposit) are reworked from shallow water, as indicated by the matrix of clastic grains with abundant reworked marl clasts (c) containing late Maastrichtian foraminifers. The older spherule layer is interbedded in late Maastrichtian marls ~5 m below and contains abundant rounded (d–f), elongate, and compressed spherules (g–k) with concave–convex contacts (l, m) and vesicular glass (n–p) in a matrix of carbonate cement. The cement matrix and absence of clastic grains indicate that no reworked component is present, and the presence of compressed glass indicates that deposition occurred rapidly while the glass was still hot and immediately after the Chicxulub impact. Spherules range from 2 to 5 mm in size.

Uppermost Maastrichtian at El Penon, NE Mexico

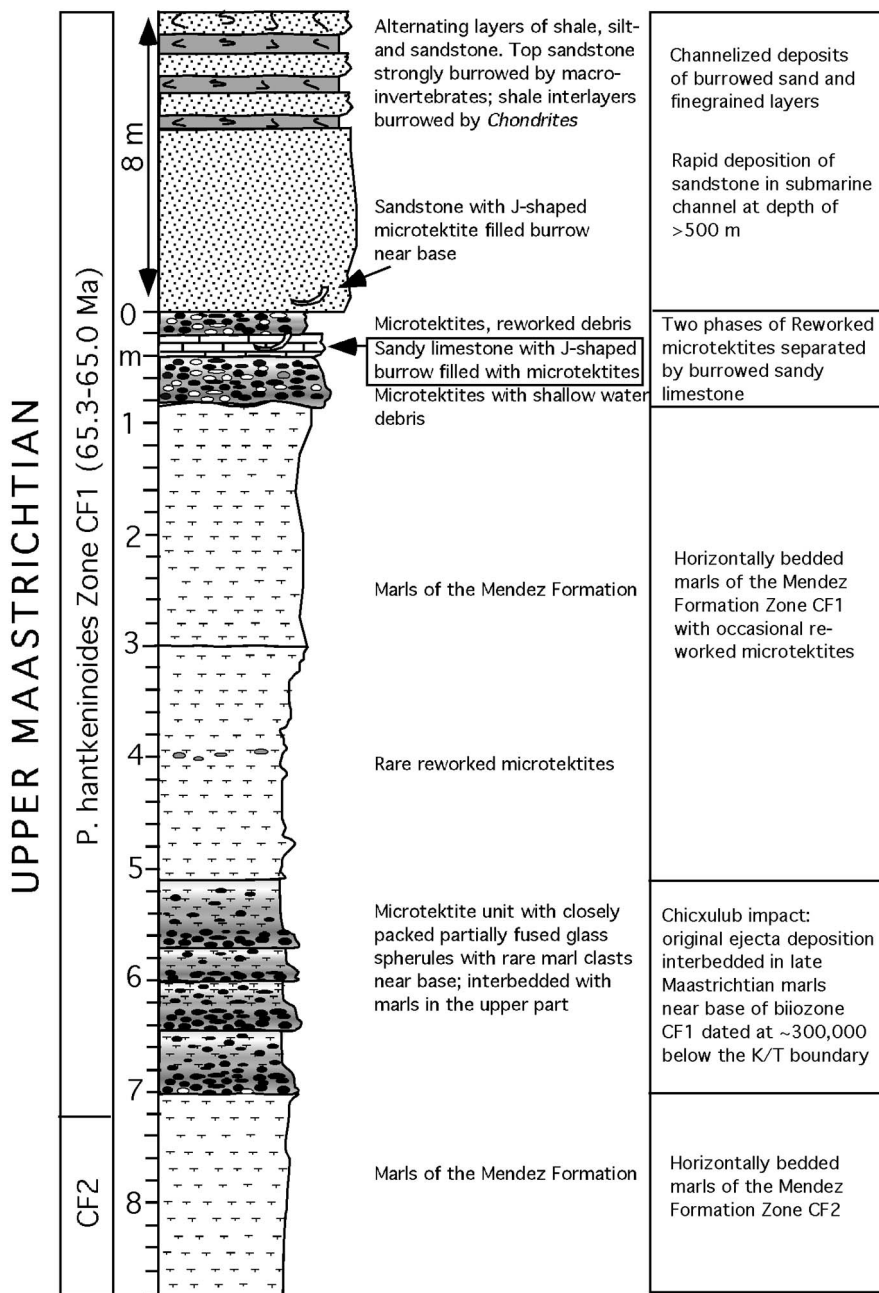


Figure 13 Lithology and stratigraphic position of microtektite ejecta layers from the Chicxulub impact in the El Penon section of northeast Mexico. Note that the original microtektite deposit is between 5 and 7 m below two reworked microtektite layers separated by a burrowed sandy limestone. Above the reworked microtektite layers are about 8 m of burrowed sandstone and alternating sand/silt and shale layers. The overlying K/T boundary and Tertiary sediments are eroded in this sequence.

BIOTIC EFFECTS OF K/T AND CHICXULUB IMPACTS

It seems reasonable to conclude that the K/T impact (but not Chicxulub) directly contributed to or caused the demise of all tropical planktic foraminifers and indirectly the decline and eventual extinction of most other Cretaceous planktic foraminifers and calcareous nannofossils (Figure 18). Species extinctions in most other groups cannot be tied to the K/T impact in a similar temporal cause-effect relationship since they generally occurred earlier and more gradually. Clearly, only those biotic effects at or shortly after the impact may be considered a direct consequence thereof. In addition, environmental stress induced by massive volcanism must be considered in any cause-effect scenario.

Deccan Traps volcanism was particularly intense during the last several hundred thousand years of the Cretaceous and at the K/T boundary (Courtillet & Renne 2003) and may have predisposed the tropical fauna to extinction. This is evident in the fact that all tropical species are rare, and most are only sporadically present during the last 300 000 years of the Maastrichtian in Mexico and worldwide. The K/T impact occurred at a time of high biotic stress with all tropical and subtropical species already on the brink of extinction.

All evidence indicates that the pre-K/T age Chicxulub impact (65.3 Ma) did not cause any species extinctions. Preliminary species population analysis from the intervals directly below and above the

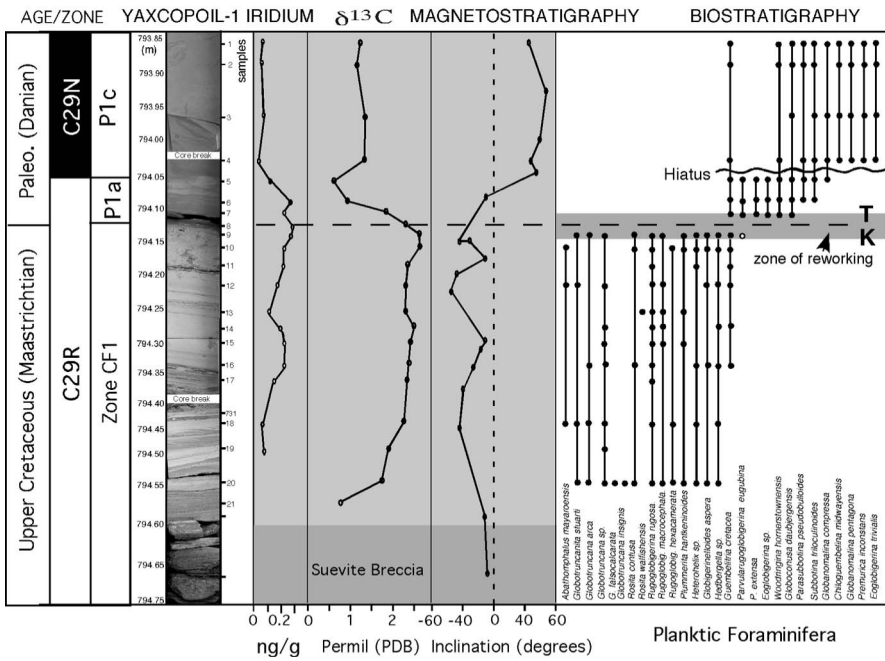


Figure 14 Stratigraphy of the impact breccia and late Maastrichtian to early Danian sediments in the Chicxulub crater corehole Yaxcopoil 1. Five sediment and age-related proxies reveal late Maastrichtian sediments overlying the impact breccia. A hiatus is present at the K/T boundary. The absence of the Ir anomaly is likely due to this hiatus.

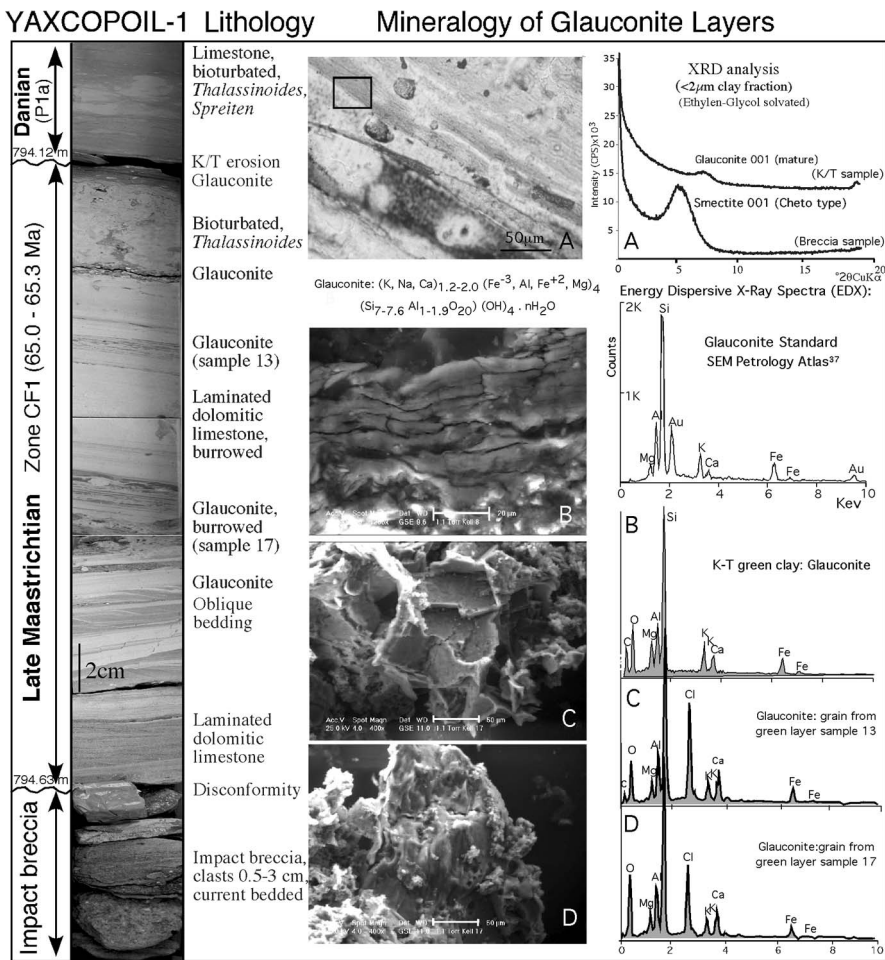


Figure 15 Lithology and glauconite mineralogy of the 50 cm interval between the disconformities at the top of the impact breccia and the K/T boundary in Yaxcopoil 1. (a) The XRD diffractogram of the K/T green clay indicates mature glauconite. In contrast, XRD analysis of breccia samples indicates the presence of well-crystallised Cheto smectite, a typical glass alteration product. (b–d) Environmental scanning electron microscope micrographs of the K/T and lower glauconite layers with electron diffractometer X-ray analyses indicate glauconite composition for all green layers (Note that the CL peak is due to the chlorhydric acid used in preparation of insoluble residues.) The glauconite reference standard from Welton (1984) is shown for comparison.

Chicxulub impact ejecta in Mexican sequences indicates relative abundance variations related to environmental changes (Keller *et al.* 2002), but further analyses are

needed to ascertain the nature of these variations and whether they can be attributed to the Chicxulub impact. This will not be easy since the Chicxulub impact

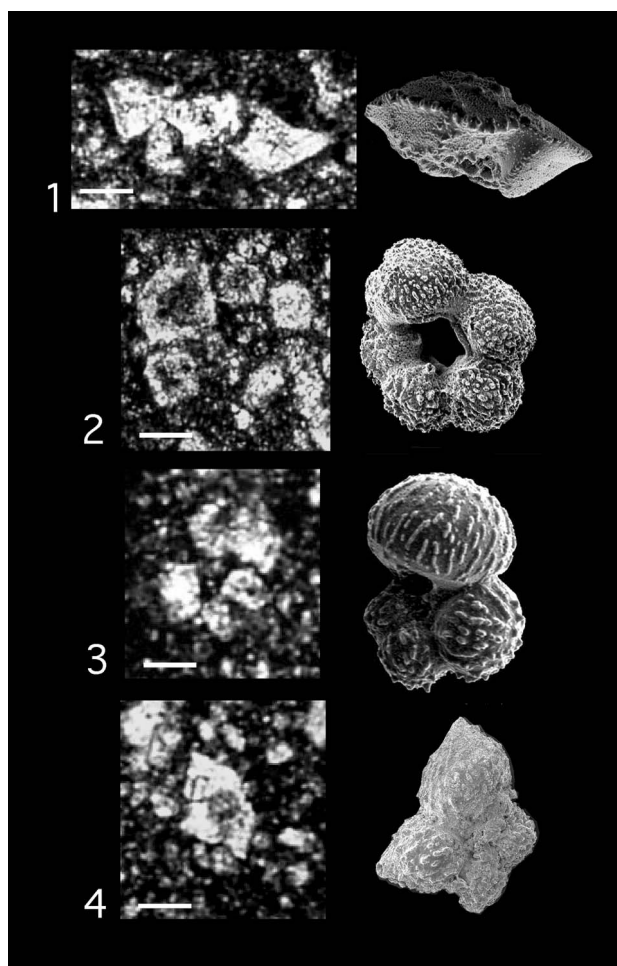


Figure 16 Thin-section photomicrographs of late Maastrichtian zone CF1 planktic foraminifera from Yaxcopoil 1 compared with pristine 3D images of the same species from the El Kef, Tunisia, stratotype section (scale bar = 200 μm for all images). (1) *Globotruncana insignis* (sample 20). (2) *Rugoglobigerina rugosa* (sample 20). (3) *Rugoglobigerina macrocephala* (sample 9). (4) *Plummerita hantkeninoides* (sample 20). Note the overall chamber arrangement, number of chambers, coiling direction and apertures. Sample numbers correlate to litholog in Figure 14.

occurred at a time of high biotic stress primarily induced by Deccan volcanism and associated greenhouse warming (Figure 18).

The Chicxulub data thus indicate that this large impact did not cause any species extinctions. Similarly, the large (90–100 km) Late Eocene Popigai and Chesapeake impacts did not cause any species extinctions. Therefore it can be generalised that impacts with crater sizes of up to 150 km in diameter do not cause mass extinctions (Figure 4), though they may temporarily alter the climate and lead to changes in dominant species populations. It can also be assumed that the impact that caused the K/T boundary mass extinction must have been significantly larger than the Chicxulub impact, unless volcanism contributed to greater biotic stress than currently thought. A reasonable estimate of the K/T impact crater is 250–350 km in diameter (Figure 4).

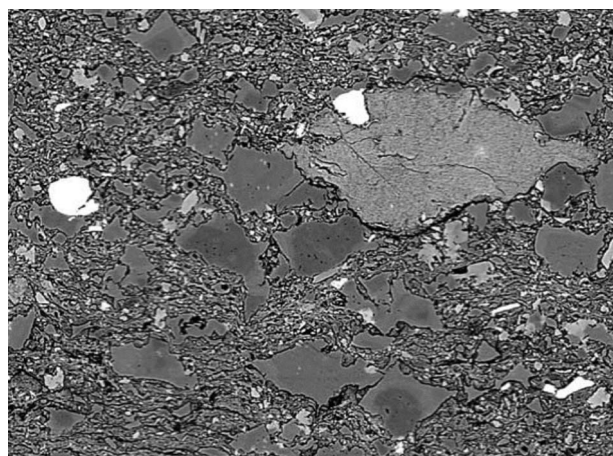


Figure 17 Thin-section photomicrograph of Yaxcopoil 1 dolomite rhombs. Jan Smit claims that the planktic foraminifera shown in Figure 16 are really dolomite rhombs that have been mistaken for foraminifera. Note that there are no similarities between the foraminifera and the dolomite rhombs, which are much smaller in size, lack the overall morphology of foraminifera and do not show the number and shape of chambers, orderly chamber arrangements, coiling direction and apertures.

CALIBRATING BIOTIC EFFECTS OF VOLCANISM

Unlike large impacts, major volcanic eruptions (e.g. large igneous provinces and continental flood-basalt provinces) do not leave short-event marker horizons in the sedimentary records; they typically continue over several hundred thousand years. This makes it much harder to evaluate the biotic effects of major volcanism (Wignall 2001). Nevertheless, it can be demonstrated, based on the late Maastrichtian Ninetyeast Ridge mantle plume eruptions at DSDP Site 216, that the biotic effects of environmental catastrophes, whether impact or volcanism, are the same in planktic foraminifera and microfossils: namely the disappearance of all tropical species, dramatic population decrease in survivors, dwarfing of species sizes, and opportunistic blooms of disaster species (Keller 2003; Tantawy & Keller 2004). All of these biotic effects are apparent during the last 400 000 years of the Maastrichtian in the tropics as well as in middle latitudes (Abramovich *et al.* 2003; Abramovich & Keller 2002, 2003; Keller & Pardo 2004), culminating in the mass extinction at the K/T boundary. Based on the end-Permian mass extinction, Twitchett (2001) hypothesised that survival during biotic crises is optimal for species that retain large populations but reduce body size (e.g. dwarfing). This seems abundantly supported by the late Maastrichtian and K/T transition.

The onset of the late Maastrichtian biotic stress coincides with rapid global warming of 4°C in deep waters between 65.35 and 65.1 Ma (Li & Keller 1998; Nordt *et al.* 2003; Wilf *et al.* 2003), which appears to be related to major Deccan Traps volcanism (Figure 10). The Chicxulub impact at 65.3 Ma may have contributed to this greenhouse warming and exacerbated the already high stress conditions. The gradual decrease in macro-invertebrates, rudistid extinction, and onset of

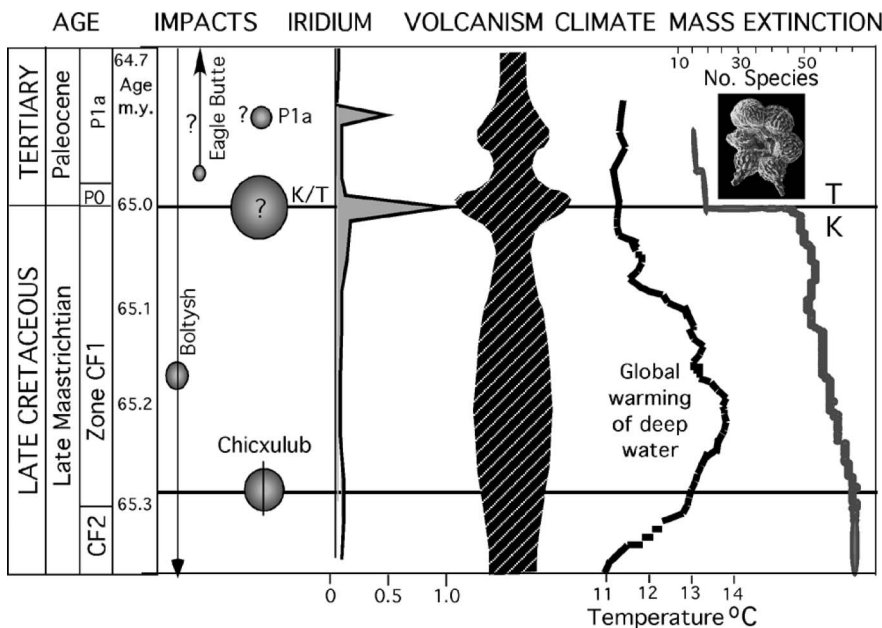


Figure 18 Summary of impacts, volcanism, climate and diversity decline in planktic foraminifers across the K/T transition. The K/T impact is inferred based on the global iridium anomaly; an early Danian impact is inferred based on the Ir anomaly (Keller *et al.* 2003). Volcanism data from Courtillot (1999); climate data from Li & Keller (1998); diversity data from Abramovich & Keller (2002).

gradual diversity decrease in planktic foraminifers all occurred during this global warming (Johnson & Kauffman 1996; Zinsmeister & Feldmann 1996; Keller 2001; Abramovich & Keller 2003). Therefore, it can be said unequivocally that late Maastrichtian volcanism induced high biotic stress that predisposed faunas to eventual extinction. The high stress conditions appear to be largely the result of greenhouse warming and eutrophication of the water column as a result of hydrothermal and magmatic eruptions. Without these existing biotic stress conditions during the late Maastrichtian, the K/T boundary impact may not have had the same catastrophic effect.

PHANEROZOIC MASS EXTINCTIONS: CASE-BY-CASE STUDIES

Triassic – Jurassic

MASS EXTINCTION AND IMPACTS

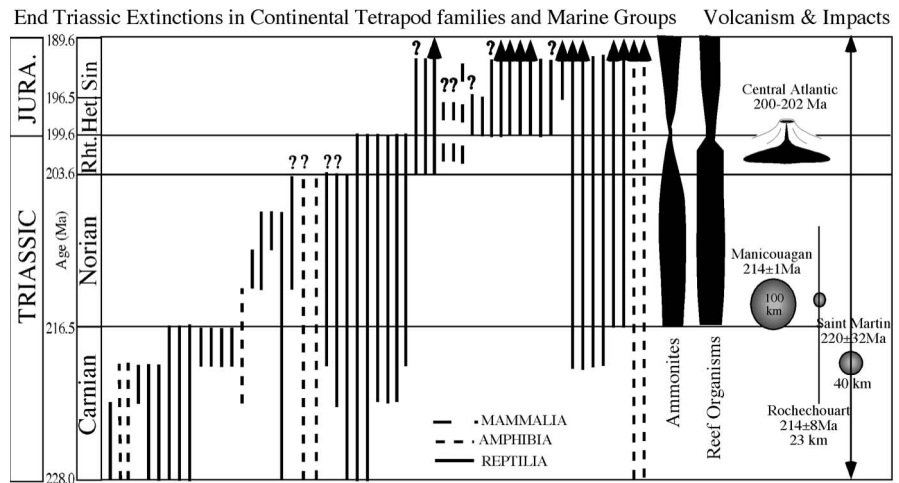
The extinctions in the Late Triassic occur in two stages, in the late Carnian and late Norian to Rhaetian. Marine extinctions in the Carnian are not very well known, though fairly large faunal turnovers mark the ammonites and reef builders (Hallam & Wignall 1997) (Figure 19). The global extent and magnitude of these extinctions in marine and terrestrial realms have yet to be demonstrated (Weems 1992; Lucas 1994; Hallam & Wignall 1997).

The early Norian was largely a time of recovery in the marine or terrestrial realms at the time of the Manicouagan impact at 214 ± 1 Ma (Figure 19). Olsen *et al.* (1987, 1990, 2002; Olsen 1999) argued that this impact may have caused the end-Triassic terrestrial extinction. But the Manicouagan impact is 14 million years older, as well as too small (100 km diameter) to cause major extinctions. The two Late Eocene impacts of the same

size, demonstrate that neither species extinctions nor other long-term environmental consequences should be expected from impact craters of 100 km diameter. Two other smaller impact craters are also known from around this interval (Walkden *et al.* 2002), though the dating is imprecise and no biotic consequences would be expected. Hallam and Wignall (1997) Ward *et al.* (2001) and Tanner *et al.* (2004) discussed the evidence, or lack thereof, for impact related mass extinctions at the Triassic–Jurassic boundary.

The end-Triassic (Rhaetian) mass extinction is estimated to have been greater than that at the K/T boundary based on the percent genera extinct per million years (48% for invertebrates: Sepkoski 1986), though this seems to be an artifact of the short Rhaetian stage (Benton 1995). In the marine realm, ammonites suffered near-total extinction at the end of the Triassic followed by radiation in the Early Jurassic. Reef organisms also suffered mass extinctions at this time along with conodonts and bivalves (Hallam & Wignall 1997). The terrestrial crisis appears somewhat later and possibly associated with flood-basalt volcanism. However, it is unclear how rapid these extinctions occurred, or whether the timing is the same among the various groups. This is largely due to lack of comprehensive data, which at this time exist primarily from the Kendelbach section of Germany, the Austrian Alps, and the Newark Supergroup in the northeastern United States. These localities suggest a sudden crisis in terrestrial plants (Fowell & Olsen 1993). Based on cyclic sedimentation (assuming Milankovitch cycles and a continuous sedimentation record) Olsen *et al.* (1990) suggested the terrestrial extinction may have occurred over less than 40 000 years, with a more conservative estimate of 500 000 years (Olsen & Sues 1986). But marine extinctions are not well dated and records of ammonoids and bivalves suggest a gradual decline in the Rhaetian Stage. Age correlations are hampered by the lack of geochemical data, as no

Figure 19 Faunal turnover, impacts and volcanism across the Triassic–Jurassic transition. Faunal turnover modified after Hallam & Wignall (1997); volcanism after Courtillot & Renne (2003).



stable isotope records exist to date across this mass extinction interval apart from the Kendelbach section, which reveals a diagenetic signal. No iridium anomaly has been confirmed.

CENTRAL ATLANTIC MAGMATIC PROVINCE

The Central Atlantic magmatic province with U–Pb ages of 202–200 Ma (Courtillot & Renne 2003) from the Newark Basin suggests that eruptions occurred in the Rhaetian Stage (end-Triassic) based on the new time scale of Gradstein and Ogg (2004). This implies that the marine and tetrapod extinctions began prior to volcanism. However, a U–Pb date of 199.6 ± 0.3 Ma from a marine ash layer 5 m below the top Triassic radiolarian zone in the Queen Charlotte Islands of Canada led Palfy *et al.* (2000) to suggest that the terrestrial crisis may have pre-dated marine extinctions. Both these conclusions are questioned on the basis of dating, as well as the poor correlation between marine and terrestrial sequences. The best dates (U–Pb ages 201 ± 1 Ma, $^{40}\text{Ar}/^{39}\text{Ar}$ age 199.5 ± 2 Ma; Hames *et al.* 2000; Weems & Olsen 1997) are from the Newark Basin, where the base of the lava flows is 30 m above the palynofloral extinctions (Olsen 1999), though volcanism may have started earlier in other regions. Hesselbo *et al.* (2002) argued that light organic carbon excursions in Greenland and the UK, are coeval with marine and terrestrial extinctions and may have been the result of a major carbon cycle perturbation due to massive volcanism. Similarly, Palfy (2003) argued that Central Atlantic magmatic province volcanism may have been the driving force in the end-Triassic mass extinction. Given the current data it is reasonable to assume that such volcanism and the accompanying environmental consequences (e.g. climate and sea-level changes) significantly contributed to the faunal and floral crises in the middle and late Rhaetian.

End-Permian

MASS EXTINCTION

The mass extinction at the end of the Permian is recognised as the largest in Earth's history with

estimates ranging from a low of 75% to a high of 96% species level extinctions (Raup 1979; Hoffman 1986). Two extinction events separated by about 16 million years are recognised (Jin *et al.* 1994a, b; Stanley & Yang 1994), with the best records from the intensively studied Chinese sections. However, any compilation of extinction data across the Permian–Triassic (P/Tr) transition is hampered by problems of biostratigraphic correlation, at the stage and biozone levels as a result of endemism (e.g. ammonoids) and differing zonal schemes. The recent revision of the geological time-scale (GST2004: Gradstein & Ogg 2004) and consequent changes in stage levels has introduced further problems in correlating published data based on regional subdivisions. For these reasons, extinction trends can be shown at best as summary for the number of families extinct for each stage (Figure 20).

An extensive review of this mass extinction and potential causes is given by Hallam and Wignall (1997) and only a brief summary is given here of the critical players in the mass extinction. The Late Permian (Maokouan) mass extinction (now placed in the Capitanian Stage) appears to have been largely a crisis in the tropics, where it decimated echinoderms, corals, brachiopods, sponges, fusulinid foraminifers and ammonoids (Ross & Ross 1995; Stanley & Yang 1994). Dating for this mass extinction is still poor.

The catastrophic end-Permian (Changhsingian) mass extinction in the marine realm appears to have come rapidly, as suggested by sudden termination of species radiations at the time the disaster hit (Hallam & Wignall 1997; Racki 2003). Twitchett *et al.* (2001) concluded that the terrestrial and marine ecosystems collapse began at the same time, preceded the negative excursion in $\delta^{13}\text{C}$, and occurred over 10–60 thousand years. Brachiopods, bivalves, complex tropical foraminifers, bryozoans, gastropods and ammonoids were all nearly wiped out (Figure 20) (Hallam & Wignall 1997). Nektonic forms survived relatively unscathed (e.g. fish, conodonts, nautiloids), possibly because of their mobility and living in the upper water column, which allowed them to escape the deadly bottom water dysoxia. Dysaerobic benthic groups also did well, including palaeo-psychrospheric ostracods, nodosarioids, agglutinated foraminifers and various 'paper pecten' bivalve

lineages, with the latter dominating the subsequent Mesozoic black-shale assemblages (Hallam & Wignall 1997). In the post-extinction phase impoverished but cosmopolitan groups survived (four bivalve genera, some microgastropods and echinoderms), marine nekton and terrestrial biota recovered rapidly, but reefs and most benthic groups remained absent.

Although the mass extinction occurred rapidly, it was not geologically instantaneous, as indicated by extinctions spread over several decimetres to a few metres of sediment. In southern China, the most intensively studied section at Meishan shows a diverse late Changhsingian fauna within 20 cm of the P/Tr boundary. Most of this fauna abruptly disappears at the base of a 5 cm-thick white ash layer enriched with pyrite (Yang & Li 1992; Yin *et al.* 1992; Hallam & Wignall 1997). The white ash layer and overlying black clay reflect anoxic conditions. The first primitive species (ammonoids, brachiopods, echinoids) appear in the overlying wackestone, which is followed by alternating centimetre-thick beds of micrites, marls and dark shales containing some Permian survivors and the first primitive species.

GEOCHEMICAL SIGNALS

The P/Tr and K/T mass extinctions and lithologic transitions show superficial similarities. For example, both boundary intervals show lithologic transitions marked by boundary clays, though the white/black double clay layer that marks the P/Tr boundary in South China is not global, whereas the K/T black clay and basal red layer are observed globally. Both boundary intervals show a succession from predominantly carbonate to marl or shale separated by a low oxygen clay layer. In both boundary transitions the majority of the taxa disappeared at or near the base of the ash (P/Tr) or clay layer (K/T). Both boundary events are also

characterised by major negative $\delta^{13}\text{C}$ excursions, although the excursion at the P/Tr boundary (Holser *et al.* 1989, 1991) (Figure 20) is twice the magnitude of that at the K/T boundary (Figure 14). In the Late Permian a smaller and more gradual $\delta^{13}\text{C}$ excursion also occurs in the late Wuchiapingian.

The gradual decrease in $\delta^{13}\text{C}$ in the upper Changhsingian is interpreted as increased supply of ^{12}C -enriched CO_2 due to oxidation of ^{12}C -rich coals exposed during a major orogenic phase in southern Gondwana that exposed glossopterid coal deposits (Faure *et al.* 1995; Berner 2002). The $\delta^{13}\text{C}$ shift at the P/Tr boundary is interpreted as a combination of productivity collapse and enhanced organic matter preservation due to anoxic deposition (Wang *et al.* 1994; Hallam & Wignall 1997; Wignall & Twitchett 1996, 2002; Kidder & Worsley 2004). Sulfur and strontium isotopes both show prolonged declines during the Late Permian followed by a rapid increase in the late Changhsingian and into the Early Triassic, which has been interpreted as increased precipitation and burial of ^{32}S -rich pyrite and increased rate of continental weathering, respectively (Korte *et al.* 2003).

Kidder and Worsley (2004) suggested that the reason the P/Tr mass extinction was so much more severe than the K/T mass extinction was due to a hot climate, salinity fluctuations, low atmospheric and oceanic oxygen levels, high sulfide levels and limited nutrient availability. They acknowledged that all of these factors are also characteristic of the K/T mass extinction, but suggest that the difference lies in the less severe environmental conditions.

IMPACT SIGNALS

A diligent search for an extraterrestrial event at the P/Tr catastrophe has been unsuccessful to date. Although Ir anomalies of 8 ppb and 2 ppb have been reported from the volcanic ash of two Chinese sections (Sun *et al.* 1984),

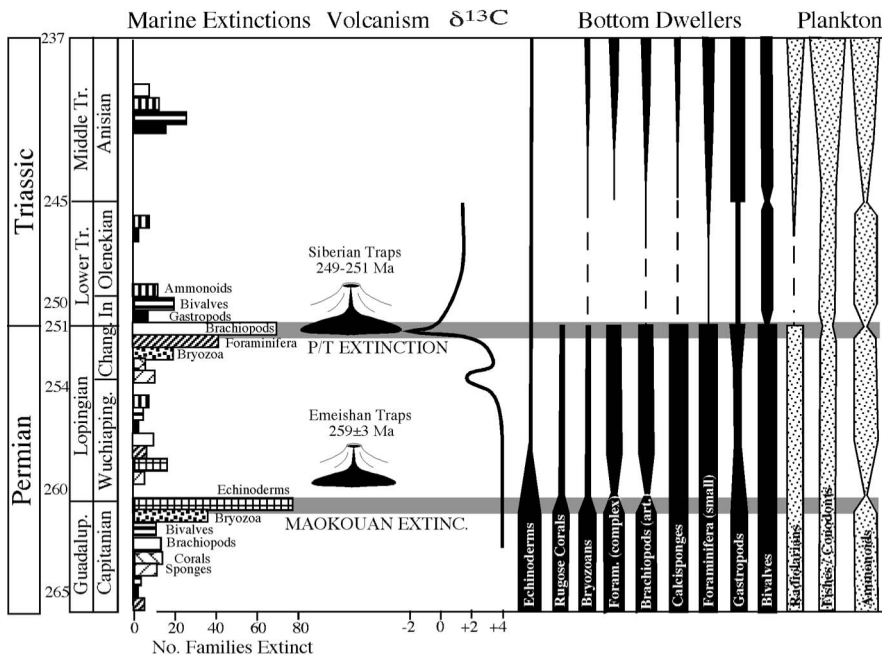


Figure 20 Faunal turnover, impacts and volcanism across the Permian–Triassic transition. Faunal data modified after Hallam & Wignall (1997); volcanism after Courtillot & Renne (2003).

these could not be confirmed by subsequent analyses (Clark *et al.* 1986; Zhou & Kyte 1988). There have been further unsubstantiated claims of an extraterrestrial impact as potential cause of the mass extinction, as well as trigger of the Siberian flood-basalt province. For example, Chai *et al.* (1992) suggested that an Ir enrichment of 0.123 ppb in the Meishan ash and Ir/Au ratio characteristic of mixed crustal and cosmic sources indicate that volcanism was triggered by an impact (Hallam & Wignall 1997). Most recently, Becker *et al.* (2004) suggested that the Bedout structure of Western Australia is a large impact crater and likely responsible for the end-Permian mass extinction. This claim has been very controversial and unsubstantiated. A petrological study of the Bedout 1 breccia shows that none of the acknowledged impact criteria such as shock metamorphic signatures are present (Glikson 2004b), i.e. no planar deformation features (PDF), diaplectic glass (maskelynite) or impact melt components. Renne *et al.* (2004) also came to this conclusion and suggested that the Ar–Ar plagioclase age of the Bedout breccia is unsubstantiated. Further, Thomas *et al.* (2004) and Wignall *et al.* (2004) reported that faunal studies of the Hovea core, located about 1500 km south of the Bedout structure in Western Australia, reveal an excellent record of the P/Tr extinction (e.g., bioturbated mudstones with abundant fauna, replaced by black shale with impoverished fauna, just below the P/Tr boundary), but no evidence for an impact (e.g., no Ir/shocked minerals).

SIBERIAN TRAP VOLCANISM

The Siberian traps represent one of the largest basalt provinces, as detailed in numerous studies and summarised in Wignall (2001) and Courtillot and Renne (2003). Renne *et al.* (1995) determined that the climax of the Siberian trap volcanism is synchronous with the P/Tr boundary within 0 ± 0.04 million years and that the main extrusion occurred over about one million years across the P/Tr boundary dated at 251 Ma, though eruptions likely occurred between 251 and 249 Ma (Courtillot & Renne 2003; Kamo *et al.* 2003). These authors warn that comparing the timing of various volcanic events based on the numerous published dates is misleading due to variable uncertainties.

The major environmental effect of the Siberian traps is probably the explosive eruption of S- and CO₂-rich basalts leading to global climate changes (Yin *et al.* 1992; Wignall 2001; Courtillot & Renne 2003; Maruoka *et al.* 2003; Racki 2003). Although volcanogenic CO₂ is a likely increased source, the amount is insufficient to cause the observed 4–5‰ negative δ¹³C spike. Other suggested sources include isotopically light carbon from melting of shallow mantle rocks (Anderson 1994), the sudden influx of C due to collapse of bioproductivity, and methane gas eruptions (Erwin 1993; Bowring *et al.* 1998). Global cooling can be ruled out as a cause for the P/Tr mass extinction because of the lack of evidence for prolonged cooling and associated sea-level regression (Wignall 2001). Global warming is more likely as suggested by increased atmospheric CO₂, increased strontium ratios, marine anoxia, oxygen isotope signals suggesting higher tropical tem-

peratures (Holser *et al.* 1989; Kidder & Worsley 2004), and marine productivity collapse (Wignall & Twitchett 1996, 2002; Sheldon & Retallack 2002). Floral evidence includes the migration of calcareous algae to boreal latitudes, the loss of the high-latitude glossopteris fauna and development of palaeosols more typical of lower latitudes (Retallack 1995; Retallack *et al.* 2003).

Current evidence suggests that the P/Tr mass extinction is likely due to the interaction and culmination of multiple environmental events, with the principle culprit global warming caused by increased CO₂ due to Siberian traps volcanism (Gruszczynski *et al.* 2003), oxidation of coal due to Gondwana tectonic activity, and possibly methane outgassing (clathrates) triggered by climate warming (Kidder & Worsley 2004).

Late Devonian

The Late Devonian Frasnian–Famennian biotic crisis marks one of the most intriguing and largest mass extinctions in Earth history and has all the aspects of a major thriller, complete with multiple impact events, major magmatic eruptions, climate and sea-level changes, ocean anoxia and multiple extinction events. On the surface, this biotic crisis is a major testing ground for biotic consequences of impacts and volcanism, as earlier explored by Racki (1999a). But on closer examination, there is still insufficient data to test for a cause-and-effect scenario.

MASS EXTINCTIONS

Four relatively sudden extinction events are recognised in the Devonian: the mid-Givetian Taghanic event, the two Frasnian upper and lower Kellwasser events [in the late Frasnian *Palmatolepis gigas* (or upper *P. rheana*) and *P. linguiformis* zones], and the late Famennian Hangenberg event (in the *Siphonodella praesculcata* zone). Hallam and Wignall (1997) reviewed these extinction events (Figure 21) and concluded that the upper and lower Kellwasser events represent the most serious biotic crises, resulting in decimation of all pelagic and most benthic groups, including the sudden crash of the stromatoporoid coral reefs. Racki (1999a) also identified the Kellwasser biotic crisis as the most severe extinction event (Figure 22). A more recent review by House (2002) concluded that on the basis of families extinct per million years the Givetian and Famennian extinctions exceed the Frasnian, though he cautioned that this assessment is based on poorly known ages of biozones and stages. However useful these summaries, they lack the precise chronostratigraphic control necessary to determine the timing and extent of the mass extinctions, although a stepwise decimation of pelagic and benthic faunas through the late Frasnian is well documented (McGhee 1996; Hallam & Wignall 1997; Racki 1999a).

Cause-and-effect scenarios can be most reliably demonstrated based on faunal, sedimentary and geochemical proxies within the same stratigraphic horizons, which can be correlated regionally and/or globally. The Late Devonian biotic crisis has been intensively studied and a wealth of data supports selective extinctions and multiple Earth-derived crises

among pelagic and benthic faunas and reef ecosystems as a result of variations in climate, sea-level and oxygenation (Buggisch 1991; Joachimski & Buggisch 1993, 1996; Copper 1986, 1998, 2002; Walliser 1996; Racki 1999a; Racki *et al.* 2002; Tribovillard *et al.* 2004), eutrophication (Murphy *et al.* 2000), nutrient-driven bioerosion masking the post-extinction record (Peterhansel & Pratt 2001), and probably linked with magmatic extrusions due to tectonic rifting (Wilson & Lyashkevich 1996; Racki 1999a, b; Ma & Bai 2002; Sandberg *et al.* 2002; House *et al.* 2000; Racki *et al.* 2002).

IMPACT SIGNALS

A multitude of hypotheses ranging from purely cosmic to exclusively Earth-derived perturbations and a combination of both, have attempted to explain the ultimate cause of the end-Devonian mass extinctions. In theory it should be easy to determine the cause-and-effect nature of terrestrial and extraterrestrial events in these biotic crises, but in practice this has proved surprisingly difficult, as discussed in a number of reviews (McGhee 1996, 2001; Hallam & Wignall 1997; Walliser 1996; Racki 1999a; Racki *et al.* 2002). The major problems in this endeavor are: (i) the absence of sedimentary sequences containing both the biotic crisis and impact signals; sequences with reported impact signals (iridium and microtektites) in South China are not associated with significant faunal changes (Ma & Bai 2002), whereas the extinction horizons reveal no impact signals; (ii) the Devonian time-scale is still conjectural; at issue are the various commonly used geochronological schemes, which differ with respect to the timing and duration of the Frasnian and Famennian Stages and result in a dating error of about 10 million years (Racki in press); and the revision by Gradstein and Ogg (2004) has not alleviated this problem because there is no easy conversion of published records to their newly proposed stage boundaries, which also require further testing; and (iii) radiometric ages of stage boundaries and dating of impact craters introduce further errors in timing and

correlation; in addition, some impact craters are poorly dated with errors of millions of years, excluding them from any cause-and-effect correlation. Racki (in press) concluded that with the current conjectural Devonian time-scale an impact-extinction event correlation is practically impossible. Here we will attempt the impossible, while keeping in mind the limitations of the current datasets and time-scales.

The Late Devonian to Early Carboniferous was clearly a time of multiple impacts with at least two of them in the range of 100–120 km -diameter (Alamo and Woodleigh: Glikson *et al.* 2005), two craters around ~50 km (Siljan and Charlevoix) and two smaller craters (~20 km: Ilynets, Kaluga) (Figure 23). In addition, several microtektite layers have been reported from sections in South China and Belgium (Claeys *et al.* 1992), though the latter were found to be man-made glass spherules from road reflectors, and some of the South China spherule layers are of unknown origin (Wang 1992). Ma and Bai (2002) identified microtektite-like microspherules of probable impact origin in several layers across the Frasnian–Famennian (F/F) boundary in two South China sections with peak abundance distributions below the end-Frasnian (*P. linguiformis* zone) mass extinction. They concluded that the microtektites are not directly associated with the mass extinction.

Similarly, the impact craters could not be linked to the mass extinctions. The Alamo breccia is biostratigraphically dated as occurring in the lower Frasnian *P. punctata* conodont zone (Sandberg *et al.* 1997; Warme & Kuehner 1998), and hence pre-dates the end-Frasnian mass extinctions by at least 5 million years. The *P. punctata* conodont zone spans about 0.5 million years based on the old time-scale (McGhee 1996 pp. 14, 15) and is currently dated between 385.3 and 382.8 Ma based on the new time-scale of Gradstein and Ogg (2004). The Woodleigh impact, radiometrically dated at 359 ± 4 Ma, and the smaller Siljan impact dated 361 ± 1.1 Ma, span the interval of the latest Famennian Hangenberg extinction, based on the new time-scale. But to link

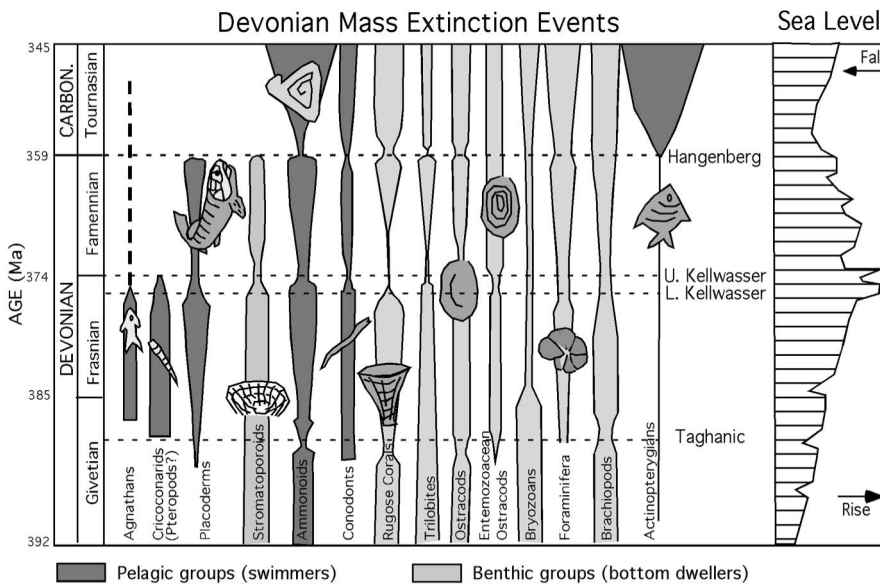
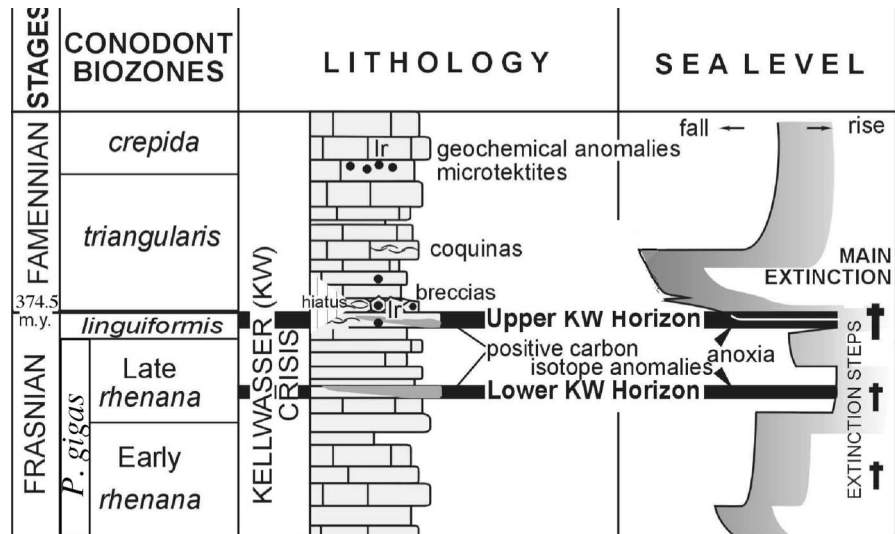


Figure 21 Biotic effects of Devonian mass extinction events (after Hallam & Wignall 1997). The most severe biotic crises are the lower and upper Kellwasser events, which occurred at the end of the Frasnian and decimated the reef system, shallow benthos and some pelagic swimmers. These crises were associated with major sea-level rises, warm climate and widespread ocean anoxia. The Hangenberg extinction affected mainly the pelagic groups (ammonoids and fishes), whereas the Middle Devonian Taghanic extinction was part of a long-term diversity decline.

Figure 22 Composite diagram of the Frasnian–Famennian transition with the two Kellwasser extinctions, sea-level changes and geochemical anomalies (modified after Racki 1999a).



these two impacts to the mass extinction requires more precise dating, as well as impact signals within the same stratigraphic sequences that document the extinction horizon. Attempts to identify impact signals based on geochemical analyses across this and other extinction events have so far failed.

LAG-TIME MULTIPLE IMPACT HYPOTHESIS?

McGhee (2001) and Warme and Kuehner (1998), strong advocates for impact-caused mass extinctions during the Late Devonian, conceded that the prolonged biotic crisis with its extinction pulses fails the test of impact-caused mass extinctions. Instead, they argued that the lag-time multiple impact hypothesis of Poag (1997) and Poag *et al.* (2002), developed for the Late Eocene multiple impacts and for which there is no supporting evidence (see above), is applicable to the Late Devonian mass extinctions. McGhee (2001) argued that the Alamo and Siljan impacts caused long-term greenhouse warming resulting in the mass extinction at the end of the Frasnian, about 3 million years later based on the old time scale, and 6 million years later based on the new time-scale (Gradstein & Ogg 2004). A test of this model by Racki (in press; pers. comm 2004) based on the palaeotemperature curve of Joachimski *et al.* (2004) failed to show evidence of climate warming until the lower *P. rhenana* zone, several million years after the impacts of the *P. punctata* zone. Thus, there can be no correlation between these early impacts and the late Frasnian extinctions. Further problems in McGhee's correlation of impacts and extinctions are introduced by the new chronostratigraphic time-scale, which places the Siljan and Woodleigh impacts in the Famennian with error bars spanning the Hangenberg extinction event (Figure 23) (Gradstein & Ogg 2004).

There seems to be no support for long-term climate warming induced by multiple impacts in either the Late Eocene or Late Devonian in the current database. Although climatic effects over decades would be expected from large impacts, it is uncertain whether such short-term changes can be detected in the geological

record, and if so whether they can be distinguished from normal climatic variations, diagenetic overprints, or the effects of pulsed, long-term volcanism. There seems to be general agreement, however, that the multiple impacts of the Late Devonian did not directly cause the mass extinctions. This is consistent with the Late Eocene record and confirms the observation that impacts with crater sizes around 100 km in diameter do not cause species extinctions.

VOLCANISM: THE VILUY TRAPS

No major flood-basalt provinces, similar to the Siberian or Deccan traps, are known from the Late Devonian. Nevertheless, widespread volcanism occurred around the time of the mass extinctions, as recognised by basement uplift, magmatism and rifting spanning about 2000 km across the East European Platform (Wilson & Lyashkevich 1996; Racki 1998; Courtillot & Renne 2003). Two major volcanic phases are recognised at the time of the mass extinctions at the end of the Frasnian and Famennian (Racki 1998). The volcanic province is estimated to have been relatively small (< 10 000 km³), although this estimate is derived largely from erosional remnants and the actual volcanic province may have been much larger even within the East European Platform (Wilson & Lyashkevich 1996) and may be a minor part of plume clusters in eastern Laurussia and Siberia (Ernst & Buchan 2001; Abbott & Isley 2002). Courtillot and Renne (2003) suggested that the kimberlite pipes from the eastern Siberian platform (including the Viluy and Markha basins) studied by Kravchinsky *et al.* (2002) could be the remnants of a major continental flood-basalt province that caused the mass extinctions in the Late Devonian. The age of this volcanic province, termed the Viluy Traps, is currently estimated between 377 and 350 Ma (Kravchinsky *et al.* 2002).

Although the current age data of the Viluy Traps spans the time of the F/F mass extinctions, dating is too imprecise to allow for an estimate of the environmental and biotic consequences of this volcanism, whether it

AGE (MA) IMPACTS & EJECTA VOLCANISM EXTINCTIONS

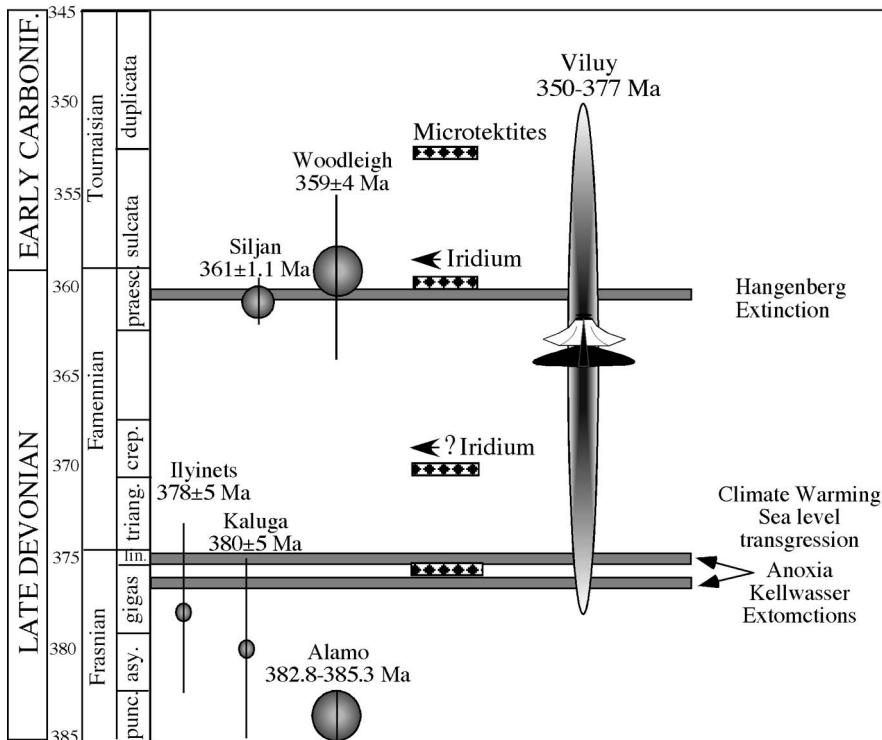


Figure 23 Late Devonian mass extinctions, impacts and volcanism reveal a time of major environmental upheaval, though no cause-effect scenario is supported between either impacts or volcanism and mass extinctions. Viluy magmatic province after Courtilot & Renne (2003); impact craters after Earth Impact Database 2005 and Haines (2005).

caused global warming or cooling, whether it resulted in hydrothermal activity leading to eutrophication and anoxia or dysoxia, or indeed whether it caused the biotic stress that lead to the mass extinctions. However, such a multicausal scenario appears likely given the current evidence of biotic stress from the fossil record, including high silica productivity and/or eutrophication phases (Racki 1999a, b; Peterhansel & Pratt 2001), and environmental stress from geochemical indicators (Racki *et al.* 2002; Joachimski *et al.* 2001; Girard & Lécuyer 2002). Although Late Devonian mass extinctions and environmental stress conditions have yet to be directly linked to Late Devonian volcanism in the same stratigraphic sequences, this has been done for the late Maastrichtian mantle plume volcanism on Ninetyeast Ridge (Keller 2003). At this locality, the onset of magmatic eruptions led to immediate catastrophic biotic effects in planktic foraminifers causing the disappearance of all but disaster opportunist species. The direct cause appears to have been hydrothermal activity-related eutrophication of the water column which eliminated all but one disaster opportunist surface dweller, reduced low-oxygen-tolerant species, and led to dwarfing of all survivors. Biotic recovery began as soon as volcanic activity ended.

The more severe and prolonged volcanic activity of the Late Devonian and its environmental effects are likely causes for the Kellwasser (F/F) biotic crises. The devastating Kellwasser mass extinctions in shelf habitats may primarily have been caused by a lethal combination of tectonically induced hydrothermal and volcanic processes, with the resultant eutrophication, climate warming, sea-level transgression, sluggish circulation and anoxia (Hallam & Wignall 1997; Racki

1999a, b; Joachimski *et al.* 2001, 2002; Peterhansel & Pratt 2001; Racki *et al.* 2002). However, oxygen isotope records from conodont apatite (Joachimski & Buggisch 2002) also suggest repeated cooling episodes of up to 5–7°C during the late Frasnian creating a ‘destabilized greenhouse warming’ (Racki 1998) with high stress conditions for tropical shallow-water faunas. In contrast, the smaller Hangenberg extinction at the end of the Devonian appears to have been related to biotic stress induced by climate cooling, a sea-level regression and volcanism.

End-Ordovician

The end-Ordovician biotic crisis is the first of the five major Phanerozoic mass extinctions with an estimated 57% genera extinct (Hallam & Wignall 1997). The pelagic graptolites seem to have been wiped out in a single extinction phase, whereas most benthic groups suffered two extinction phases with the first phase hitting low-latitude taxa (trilobites, brachiopods, bryozoans, echinoderms). The second phase occurred about half a million years later and wiped out the low-diversity cool-water fauna of middle and deep shelf environments. No significant magmatic events or impact signals have been identified for this interval. Barnes *et al.* (1996) suggested a possible middle Ordovician ‘superplume activity’, though this has yet to be confirmed (Abbot & Isley 2002). Minor concentrations of iridium and trace elements detected in the Yangtze Basin are interpreted as a combination of sediment starvation and anoxia during the sea-level transgression (Wang *et al.* 1993, 1994). The mass extinction appears to have been due to several factors, including cooling, glaciation, sea-level

regression and major changes in oceanic circulation leading to the first extinction phase in pelagic groups (e.g. graptolites, conodonts). The second phase appears to have been related to climate warming and ocean bottom water anoxia eliminating shelf habitats (Hallam & Wignall 1997). There seems to be no evidence that volcanism or impacts played significant roles in this mass extinction event.

NEOPROTEROZOIC RADIATION: SNOWBALL EARTH

The Cambrian explosion of life is commonly attributed to the aftermath of a global glaciation period (Snowball Earth hypothesis: Hoffman *et al.* 1998), though some have suggested it was likely due to the Acraman impact (Grey *et al.* 2003). Others have questioned the reality of this 'explosion' and suggested that it may be an artifact of fossil preservation and incomplete sedimentary records (Shields 1999). Precision in correlation of geological and faunal events decreases with increasing age and the dating of Neoproterozoic rocks is no exception. The problems are confounded by absent to rare fossils, poor preservation, and incomplete stratigraphic records. Nevertheless, the generalised outline of this early life history is apparent.

The radiation of life largely began at the Marinoan glaciation (620–600 Ma and rapidly diversified after the Gaskiers glaciation (580 Ma: Evans 2000; Lund *et al.* 2003; Kendall *et al.* 2004). These glaciation episodes are accompanied by major negative $\delta^{13}\text{C}$ excursions (Shields 1999). Three impacts have been identified over this interval. The 90 km-wide Acraman impact crater is the largest and dated at 580 Ma (Grey *et al.* 2003; Glikson 2004a; Williams & Gostin 2005), and hence close to the Gaskiers glaciation. The 60 km-wide Beaverhead structure is dated at *ca* 600 Ma (no error margins given: Earth Impact Database 2005), and the 25 km-wide Strangways structure is dated 646 ± 42 Ma. Neither of these relatively small impacts can be assumed to have had long-term global effects on either climate or evolution, considering that not even the largest known impact (Chicxulub) can be demonstrated to have had long-term effects or caused species extinctions. The smaller Acraman impact is the same size as the two Late Eocene impact craters (Popigai and Chesapeake), which caused no species extinctions or evolution and no demonstrated long-term environmental effects (see above).

DISCUSSION

Mass extinctions

Mass extinctions are generally defined as geologically brief intervals, usually less than 1 million years, where rates of extinctions far exceed normal background turnover, where diverse taxa from a wide range of habitats are affected, and where the disappearance of taxa is global and permanent (Hallam & Wignall 1997). In practice, a 1 million years interval of elevated extinction rates is far too long when testing for

impact-caused mass extinctions. The proximate biotic effects of impacts (e.g. acid rain, global cooling, wildfires, ozone depletion—though all are speculative and controversial) are not expected to range beyond 2–24 months (Toon *et al.* 1997; Pollack *et al.* 1983) and would be observed as a mere geological instant in the sedimentary record. Such short-term cooling (nuclear winter) was recently claimed to have been identified after the K/T impact based primarily on a change in the coiling ratio of the benthic foraminifer *Cibicides pseudoacutus* and incursion of cooler water species into lower latitudes (Coccioni & Galeotti 1998; Galeotti & Coccioni 2002). Stable isotope analysis of left and right coiling specimens of *C. pseudoacutus*, however, reveal no change in either $\delta^{13}\text{C}$ or $\delta^{18}\text{O}$ signals (Keller *et al.* 2002) and hence fail to support the post-K/T impact nuclear winter. It remains doubtful that such short-term climate signals can be obtained from the sedimentary record and differentiated from seasonal variations or diagenetic effects. Longer term environmental effects, such as greenhouse warming due to increased CO_2 as a result of volcanism during the latest Maastrichtian (65.4–65.2 Ma), may last more than 200 thousand years and significantly increase biotic stress, but has not been shown to cause mass extinctions (Abramovich & Keller 2003). Volcanism-related biotic stress, in principle, should be much the same as that due to large impacts.

In general, rapid extinctions—those occurring over a short stratigraphic interval—have been interpreted as evidence favoring the impact model (Smit 1999; Pospichal 1994; Marshall & Ward 1996; Vermeji 2004). But caution is in order. Dramatic environmental changes, such as at the K/T boundary, are generally associated with hiatuses as a result of erosion or non-deposition (MacLeod & Keller 1991; Keller 2002; Keller *et al.* 2003). Even the most well-known K/T sections are incomplete, including Gubbio in Italy, Zumaya, Agost and Caravaca in Spain, Stevns Klint, and Nye Klov in Denmark. Hiatuses often span one or more biozones, or condensed sedimentation results in very reduced biozones. The fact that the iridium anomaly is present in these sections is often cited as evidence for their completeness. But iridium concentration itself is usually the result of condensed sedimentation (Tredoux *et al.* 1988; Crockett *et al.* 1988; Sawlowicz 1993). Extinction horizons that coincide with hiatuses or condensed intervals prevent determination of the rate and tempo of the biotic change. Although such sections still allow us to determine that a mass extinction occurred, it cannot be determined whether species began to disappear prior to the impact, how many succumbed immediately, and what proportion survived and for how long. In fact, hiatus and erosion horizons can transform a prolonged, gradual extinction record into an apparent 'catastrophic collapse' of the ecosystem. Failure to recognise hiatuses frequently results in erroneous interpretations of sudden catastrophes.

High-resolution stratigraphic studies, comparable to the Eocene–Oligocene and K/T boundaries, are largely lacking for the older mass extinction horizons. The invertebrate extinction pattern, upon which the genus-level mass-extinction record of Sepkoski (1994) is based, is generally known to the stage level, rather than

biozones (Benton 1994; Sepkoski 1996; Hallam & Wignall 1997). With stages varying from 1 to 12 million years in duration and averaging 6 million years, this low-resolution record prevents evaluation of true correspondence between impacts, volcanism and mass extinctions. But the geological record clearly demonstrates that impacts, massive volcanism and mass extinctions frequently cluster, indicating heightened environmental crisis, even if no synchronous relationship can be shown (MacLeod 1998, 2003; Tsujita 2001). High-resolution studies currently being undertaken for all major mass extinctions are likely to resolve the questions of cause-and-effect between impacts, volcanism and mass extinctions.

Coincidences between impacts and volcanism

Impacts and volcanism are the two catastrophes frequently invoked as primary causes for the major mass extinctions and it is often suggested that large impacts may have triggered massive flood-basalt volcanism. White and Saunders (2004) statistically evaluated the probability of random coincidences between impacts and 2 million years periods of flood-basalt volcanism (large igneous provinces and continental flood-basalt provinces) over a period of 300 million years and determined the probability at 57% for at least one K/T-sized impact, 19% for two impacts, and only 4% for three impacts. But for smaller craters (≤ 100 km), the probability of three coincidences increases to 46%, and for still smaller craters (≤ 60 km) the probability is 97%. They concluded that one should expect random coincidences between large igneous provinces and continental flood-basalt provinces and no causal relationship between impact and volcanism needs to be invoked. They further concluded that the coincidence of impacts and volcanism combined may be required to cause the largest mass extinctions due to the ecosystem's inability to cope with the effects of a large impact in a system already under stress from prolonged volcanism. This conclusion is in agreement with palaeontological observations of gradually decreasing diversity prior to mass extinctions, or long periods of biotic stress associated with prolonged volcanic eruptions.

Biotic effects of continental flood-basalt provinces and large igneous provinces

The biotic effects of massive volcanism, either of continental flood-basalt or large igneous provinces are much more difficult to assess than for impacts because no single sharp sedimentary horizon can be identified. Volcanic intensity may be variable over thousands, tens of thousands or even hundred thousands of years. Studies generally compare the overall correspondence of mass-extinction episodes with large igneous provinces to infer a cause-effect relationship (McLean 1985; Courtillot *et al.* 1986; Kerr 1998; Palfy 2003) (Figure 3a). Wignall (2001) reviewed this record and confirmed a close correspondence with five mass extinctions: the end of the middle Permian extinction (end-Guadalupian/Emeishan flood basalt), the end-Permian mass extinc-

tion (Changhsingian/Siberian Traps), the end-Triassic extinction (central Atlantic volcanism), and the end of the Early Jurassic extinction (Toarcian/Karoo Traps), but found that the onset of eruptions slightly post-dates the main phase of extinctions. For the K/T mass extinction, the main phase of Deccan Traps eruption pre-dates the main phase of the mass extinction, which coincides with an impact. Wignall concluded that the link between mass extinctions and large igneous provinces remained enigmatic with no clear relationship with size and volume of extrusives, or violence of eruptions, although he found a positive relationship between six out of 11 large igneous provinces with climate warming and ocean anoxia.

Does this mean that massive volcanism has no direct effect on biota? Not necessarily. The lack of faunal studies of sedimentary sequences interbedded with the volcanic eruptions may well have biased Wignall's (2001) survey. Moreover, ocean anoxia is invariably associated with high-intensity benthic extinctions. The biotic effects of volcanism were recently studied based on DSDP Site 216 on Ninetyeast Ridge, based on planktic and benthic foraminifers and calcareous nannofossils in the volcanic sediments (Keller 2003, 2004; Tantawy & Keller 2004). Volcanic activity began at about 66.8 Ma and continued with variable intensity until about 400 000 years before the K/T boundary. The onset of volcanism eliminated all but the disaster opportunistic calcareous nannofossils and planktic foraminiferal species (e.g. *Guembelitra cretacea*). Successively lower intensity volcanism led to the return of low-oxygen-tolerant species, followed by small generalist surface dwellers. Once volcanism ceased, more species migrated into the area, though they remained dwarfed indicating that stress conditions continued. The K/T mass extinction occurred during this recovery phase. Similar stressed assemblages were observed in Madagascar, Israel and Egypt (Keller 2003) indicating the widespread biotic effects of volcanism. Similar studies are needed to assess the biotic effects of large igneous provinces at other mass extinction episodes to determine the true biotic effects of volcanism.

Calibrating the impact-kill effect

Impacts occurred throughout the Phanerozoic, as shown by the Earth Crater Impact Database as well as Ir anomalies, shocked quartz and tektites (Figure 1). The actual impact occurrence is likely to be much higher, as many craters and impact ejecta horizons have yet to be discovered, many have been eroded over time and those which hit the oceans have been lost. MacLeod (1998, 2003) showed that the overall cratering record for the Phanerozoic has a poor correlation (Spearman correlation coefficient -0.219) with the invertebrate extinction record based on stage-level resolution. (Note that the larger time intervals of stage-level resolution favours higher correlation.) Nor did he observe a clear correspondence between frequent small impacts (10–50 km-diameter) and higher extinction intensity, or larger impacts (50–100 km-diameter) and higher extinction intensity, indicating

that impacts of less than 100 km-diameter do not cause significant extinctions. This observation is corroborated by the fact that the two Late Eocene impacts (Popigai, 100 km and Chesapeake, 90 km) are not associated with elevated species extinctions (Keller *et al.* 1983; Keller 1986; Prothero 1994; Montanari & Koeberl 2000).

Only the K/T mass extinction appears to correspond directly with a large impact event based on the global iridium anomaly, as also concluded by Hallam and Wignall (1997), Courtillot (1999), Keller *et al.* (2003) and Alvarez (2003). This impact is widely believed to be the Chicxulub crater on Yucatan (Hildebrand *et al.* 1991; Sharpton *et al.* 1992; Alvarez 2003). Biostratigraphic evidence from the new crater corehole Yaxcopoil 1 and from northeastern Mexico sections indicate that this impact event pre-dates the K/T by about 300 000 years and caused no significant species extinction (Keller *et al.* 2004a, b). This indicates that impacts of crater size up to 170 km-diameter do not cause mass extinctions (Figure 4). No iridium anomaly has been demonstrated to be associated with the Chicxulub impact (Keller *et al.* 2003). The well-documented global Ir anomaly at the K/T boundary represents another, and probably larger (~250–350 km) impact for which no crater has been found to date.

Raup (1992) proposed a 'kill curve' based on the assumptions that the Chicxulub crater is of K/T age, about 180–300 km wide, and caused the extinction of about 80% marine species. These assumptions have proved to be overestimates. The impact crater size is now estimated at 170 km (Bell *et al.* 2004), the mass extinction at about 45%, and most importantly, the Chicxulub impact pre-dates the mass extinction and caused no species extinctions (Keller *et al.* 2003, 2004a, b). In a test of Raup's impact kill-curve, Hallam and Wignall (1997) showed that none of the well-known impact craters, including Ries (30 km), Montagnais (45 km), Popigai (100 km), and Manicouagan (100 km) for which the projected kill-curve predicts up to 50% of the species extinct, caused extinctions. The Chesapeake (90 km) and Chicxulub (170 km) craters can now be added to this list of impacts with no recognised species extinctions. I concur with Hallam and Wignall's (1997 p. 245) conclusion that bolide impacts cannot be plausibly invoked as a general cause of mass extinctions. Based on this study it can be concluded that impacts with craters up to 170 km-wide cause no significant species extinctions. The K/T mass extinction coincides with a global iridium analysis that suggests a significantly larger impact than Chicxulub.

CONCLUSIONS

Large impacts are commonly credited with the most devastating mass extinctions in Earth's history and the K/T boundary impact and mass extinction is the strongest and sole direct support for this view. Even Alvarez (2003) conceded that there is no evidence that links any other major mass extinction to impacts. Of course, one might argue that the real killer impacts

have yet to be found, but this begs the question of how large an impact, or how many impacts, are necessary to cause a mass extinction and over what time interval? But, do large impacts really cause mass extinctions by themselves? Or must it be the unlucky coincidence of one or more large impacts upon an environment already stressed due to volcanism and associated climate and sea level changes? A survey of the major Phanerozoic mass extinctions, the impact record and major volcanism (continental flood-basalt provinces and large igneous provinces) leads to the following conclusions.

(1) The Chicxulub crater is not the K/T impact event, but pre-dates the K/T boundary and mass extinction by 300 000 years as demonstrated by the microtektite ejecta layer interbedded in late Maastrichtian sediments near the base of zone CF1 in the Chicxulub crater core Yaxcopoil 1 and throughout northeastern Mexico. No Ir anomaly has been detected to date for this impact.

(2) The K/T mass extinction (of largely tropical microplankton in the marine realm) coincides with a global Ir anomaly, which indicates a second, significantly larger impact event. The size of this still unknown impact crater is estimated to be > 250 km in diameter.

(3) The sudden catastrophic mass extinction at the K/T boundary appears to be the result of the unfortunate coincidence of a large impact upon a severely stressed environment due to massive volcanism (Deccan Traps) and resultant climate warming with tropical and subtropical species already on the brink of extinction.

(4) Known impact craters of the past 600 million years range from < 10 km to about 170 km in diameter, with Chicxulub at 170 km-wide the largest impact crater. None of them are contemporaneous with mass extinctions.

(5) Excellent age control and correlative faunal changes for the two large (100 km-diameter) Late Eocene impacts (Popigai and Chesapeake) and the pre-K/T Chicxulub impact, demonstrate that they caused no species extinctions or environmental changes. This means that impacts with craters up to 170 km-diameter do not cause mass extinctions or long-term environmental changes.

(6) Multiple impacts over a period of several 100 000 years (e.g. comet shower) are known from the Late Eocene (> 4 impacts), K/T transition (~4 impacts), Late Devonian and Jurassic/Triassic. This indicates that multiple impacts, ranging from small to large, are the norm rather than the exception, but caused no significant species extinctions.

(7) The consistent association of large magmatic provinces with all but one (end-Ordovician) of the five major Phanerozoic mass extinctions suggests that volcanism played a major role. The biotic stress appears to be due to a lethal combination of tectonically induced hydrothermal and volcanic processes, leading to increased CO₂, climate warming and possibly methane outgassing (clathrates), eutrophication, ocean anoxia and the collapse of marine productivity. It must be concluded that major magmatic events and their environmental consequences can be invoked as long-term general cause of mass extinctions.

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