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Sea-Level Changes, Clastic Deposits, and Megatsunamis across the Cretaceous-Tertiary Boundary

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INTRODUCTION

Along basin margins and continental shelves, marine sedimentation is frequently packaged as unconformity-bounded clastic depositional strata. These sequences are generally related to eustatic sea-level changes and in particular to the sea-level fall and rise inflection points when unconformities and condensed sedimentation, respectively, form (Posamentier and others, 1988; Posamentier and Vail, 1988). Sequence stratigraphic models provide a conceptual framework for understanding eustatic control on sedimentation and for applying these models to field observations (Posamentier and others, 1988; Posamentier and Vail, 1988; Donovan and others, 1988; Baum and Vail, 1988; Van Wagoner and others, 1990). On the basis of these models, sequences frequently encountered in continental shelf to basin margin settings of Cretaceous-Tertiary (K-T) transitions or near the K-T boundary in the gulf states (Texas, Alabama, Georgia), Mexico, and the Gulf of Mexico are traditionally interpreted as sea-level lowstand followed by sea-level rise deposits (Donovan and others, 1988; Baum and Vail, 1988; Mancini and others, 1989; Keller, 1989a; Savrda, 1991, 1993; Habib and others, 1992; Moshkovitz and Habib, 1993; Stinnesbeck and others, 1993; Keller and others, 1993a, 1994a).

With the suggestion of the circular gravity anomaly structure at Chicxulub in northern Yucatan as a probable K-T boundary bolide impact crater (Pope and others, 1991; Hildebrand and others, 1991; Swisher and others, 1992; Sharpton and others, 1992, 1993), these same clastic deposits have been reinterpreted by some workers as impact-generated megatsunami deposits (Hildebrand and Boynton, 1990; Hildebrand and others, 1994; Smit and others, 1992, 1994a, 1994b; Alvarez and others, 1992; Bohor and Betterton, 1993). This reinterpretation is based mainly on the theory that the Chicxulub structure is the K-T boundary impact crater, and the extrapolation that such an impact created megatsunami waves which left their marks in the sedimentary strata, as hypothesized earlier by Bourgeois and others (1988) for clastic deposits along the Brazos River, Texas. Proponents of impact-generated tsunami deposition generally agree that it is not known what such deposits look like, that is, whether bedding would be chaotic or graded, locally channelized, or sheet-like over large geographic areas; or whether they would look substantially different from sea-level lowstand deposits. They agree that such tsunami deposits would be emplaced during a single catastrophic event lasting no more than a few days and ideally that it would contain impact products such as iridium, Ni-rich spinels, shocked quartz grains, and glassy microspherules or microtektites.

To date, no K-T clastic deposit has been identified unequivocally as of impact origin, although Hildebrand and Boynton (1990), Alvarez and others (1992), Hildebrand and others (1994), and Smit and others (1992, 1994a, 1994b) have claimed such an origin for virtually all known K-T or near-K-T clastic deposits. For example, coarse-grained sands in sections from Braggs, Mussel Creek, Millers Ferry, and Moscow Landing in Alabama (known as Clayton sands), long interpreted as sea-level lowstand deposits (Donovan and others, 1988; Baum and Vail, 1988; Mancini and others, 1989; Savrda, 1991, 1993; Habib and others, 1992; Moshkovitz and Habib, 1993), are now postulated to be impact-generated tsunami deposits by Smit and others (1994a, 1994b) and Hildebrand and others (1994), largely on the basis of their stratigraphic proximity to the K-T boundary and geographic proximity to Chicxulub. Similarly, coarse-grained deposits of near-K-T boundary age at the Brazos River section, otherwise devoid of impact indicators, were interpreted as tsunami deposits by Bourgeois and others (1988).

In some sections from northeastern and east-central Mexico (Mimbral, Peñon, Mulato, Lajilla, and Tlaxcalantongo), near K-T boundary clastic deposits contain thin layers or lenses of unusual spherule-rich sediments with very rare glass fragments on a basal unconformity. These spherule-rich sediments are overlain by a thick laminated sandstone, which is overlain by alternating sand and siltstone beds (Stinnesbeck and others, 1993, this volume; Keller and others, 1994a). These sediments have also been postulated as impact-generated megatsunami deposits by Smit and others (1992, 1994a, 1994b). This interpretation was challenged by Bohor and Betterton (1993), who prefer turbiditic deposition, and by Stinnesbeck and others (1993, 1994b, this volume) and Keller and others (1994a), who argue for deposition during a sea-level lowstand followed by a transgression preceding the K-T boundary. Deep-water clastic deposits in Deep Sea Drilling Project (DSDP) Site 540 in the Gulf of Mexico, originally interpreted as gravity-flow deposits or submarine slumps (Worzel and others, 1973), were postulated by Alvarez and others (1992) as Chicxulub impact-generated tsunami deposits. However, restudy of these deposits reveals no independently age controlled K-T boundary sediments at Site 540, or elsewhere in the Gulf of Mexico and Caribbean (Keller and others, 1993a).

Are there any K-T boundary impact-generated clastic deposits? Such deposits, of course, are technically possible. We believe, however, that the K-T and near-K-T boundary clastic deposits must be interpreted within the context of their stratigraphic position, geographic location (proximity to sediment source), and paleodepth of emplacement, and within the framework of eustatic sea-level changes and their effects on deposition of sedimentary strata. In this study we address these issues not only for selected K-T boundary sequences that contain siliciclastic strata, but for K-T boundary sequences worldwide (Fig. 1), and emphasize the integration of hitherto neglected global aspects. We begin by reviewing evidence for eustatic sea-level changes across the K-T transition based on K-T sequences worldwide. Clastic deposits in all known marine K-T sequences are then reviewed for biostratigraphy, paleodepth of deposition, and sedimentary characteristics, and correlated to the global eustatic sea-level curve.

BIOSTRATIGRAPHY

Over the past decade the K-T boundary debate has indirectly resulted in the accumulation of the most extensive high-resolution paleontological database of any geologic interval, including more than 45 biostratigraphic zone complete (all biozones present) or nearly complete K-T boundary sections in addition to numerous K-T successions that contain major hiatuses (MacLeod and Keller, 1991a, 1991b; Keller and others, 1993a). Our results are based upon the boundary sec-

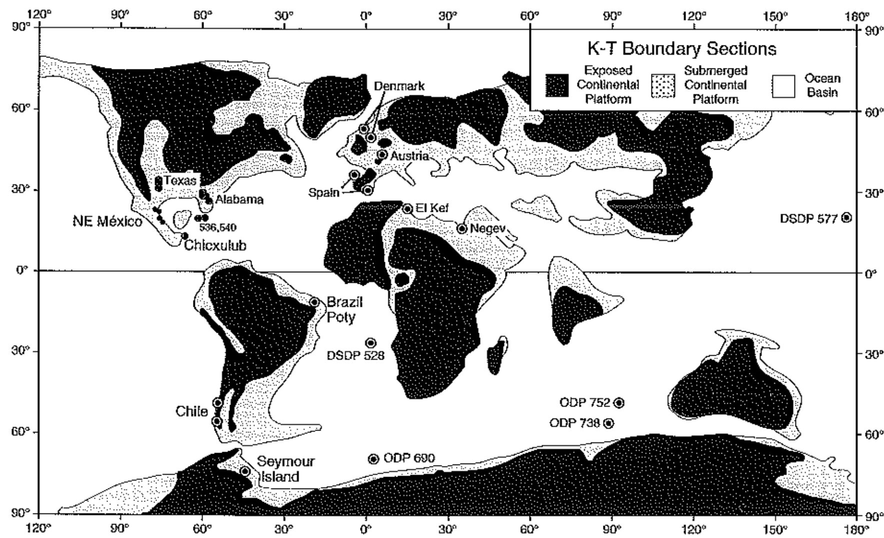

Figure 1. Locations of complete and near-complete K-T boundary sections plotted on a paleogeographic reconstruction of continental positions at the time of the K-T boundary. White = ocean basins; light stippling = continental platforms; black = inferred extend of terrestrial exposure (from MacLeod and Keller, 1991a). DSDP = Deep Sea Drilling Project; ODP = Ocean Drilling Program.

tions shown in Figure 1, where many location points represent multiple sections (e.g., 10 sections for northeastern Mexico, 3 sections each for Brazos and Alabama, 5 sections for the Negev). Each of these sections contains an abundant, well-preserved and well-documented planktonic, foraminiferal, nannoplankton, or dinoflagellate record. In addition to geochemical markers (e.g., Ir anomaly, Ni-rich spinels, $\delta^{13}C$ shift), these biostratigraphies provide the necessary high-resolution time control for accurate placement of the K-T boundary, along with dating of sea-level changes and coarse-grained clastic units.

Many recent high-resolution biostratigraphic studies have lead to significant refinements of the K-T boundary biozonations, especially for the Danian interval. Consequently, the original planktonic foraminiferal zonation of Keller (1988a) was modified and updated in Canudo and others (1991) and Keller (1993) to reflect the improved database and increased confidence in additional and alternative zonal markers. Figure 2 shows the updated planktonic foraminiferal biozonation with first and last appearance datums and comparison with biozonations of Smit (1982) and Berggren and Miller (1988). Based on the Agost and El Kef sections, Pardo and others (this volume) defined the

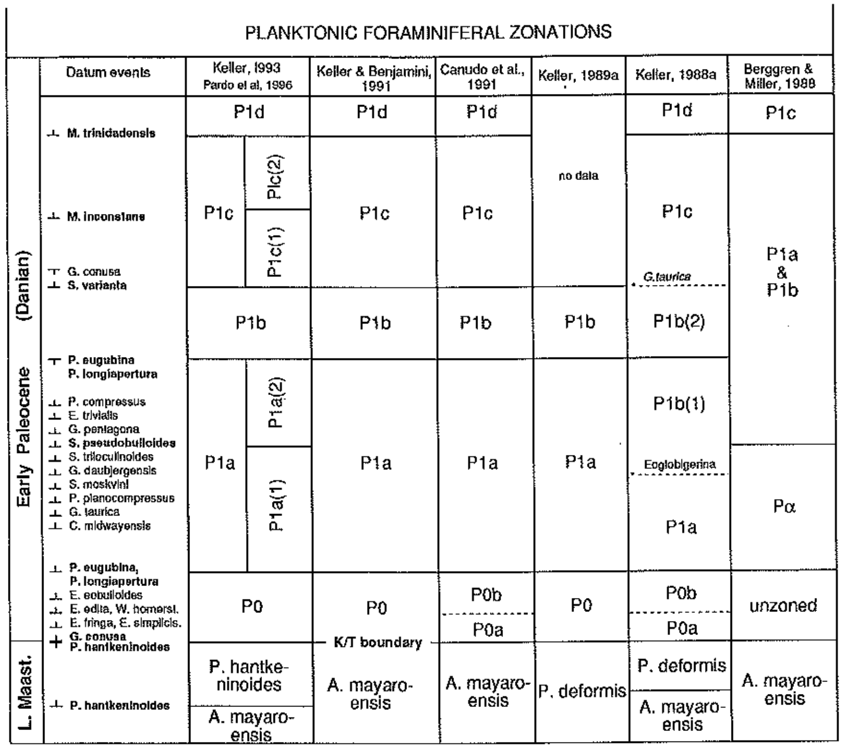

Figure 2. The most commonly used planktonic foraminiferal biozonations for the K-T boundary transition showing successive refinements as a larger global database was developed. Zonation follows Keller (1993); the uppermost Maastrichtian biozone (Plummerita hantkeninoides) *is added to mark the topmost 170–200 k.y. of the Maastrichtian.*

Plummerita hantkeninoides biozone (see also Masters, 1984, 1993) to characterize the topmost (~170–200 k.y.) of the Maastrichtian and to be within chron 29R below the K-T boundary.

The age and duration of early Danian biozones can be estimated relatively accurately. The K-T boundary is currently placed within the upper half of paleomagnetic anomaly C29R at 65 Ma based on $^{40}Ar/^{39}Ar$ dating of melt rock from K-T boundary sections at Beloc, Haiti, and Chicxulub (Swisher and others, 1992; Sharpton and others, 1993). The portion of C29R below the K-T boundary is estimated as 350 k.y., and the portion above the boundary is estimated as 230 k.y. (Herbert and D'Hondt, 1990; Berggren and others, 1985). The C29R-C29N boundary corresponds closely to the top of zone P1a (extinction of *P. eugubina*), as observed in sections from Brazos River and DSDP Sites 577 and 528 (Keller, 1989a; Chave, 1984; Bleil, 1985). Biozones P0 and P1a thus span a total of 230 k.y. and of this time interval, zone P0 is estimated to span ~40 to 50 k.y. (MacLeod and Keller, 1991b; D'Hondt and Herbert, 1991).

RELATIVE SEA-LEVEL CHANGES

Relative changes in sea level frequently determine the nature and characteristics of sediment deposition. It is therefore important to determine paleodepths of deposition of each K-T boundary sequence and to trace the relative sea-level changes across the boundary horizon. Figure 3 illustrates the average paleodepth

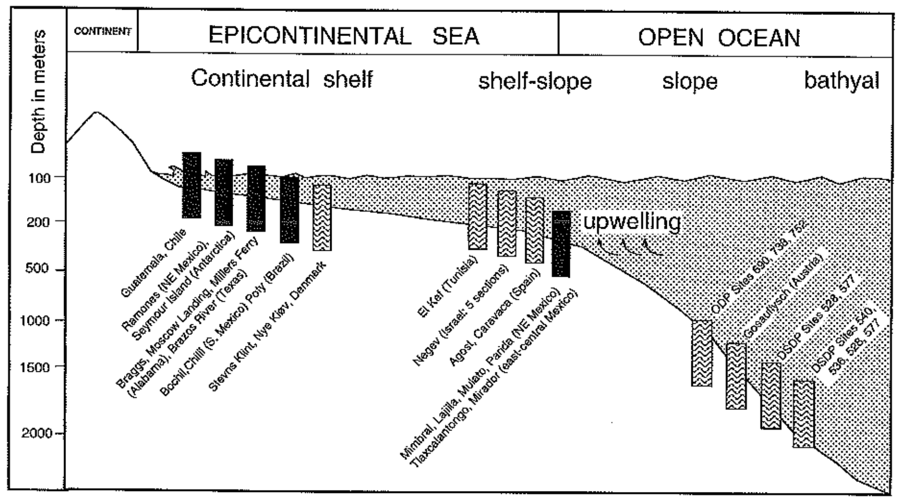

Figure 3. Depositional environments and average paleodepths of primary (complete or near-complete) K-T boundary sections from nearshore (inner neritic), middle shelf, outer shelf to upper slope, and the deep sea. Note that clastic deposits (black) are generally common in nearshore environments and rare in deeper waters unless locations are near continental margins such as the northeastern Mexico sections. Black columns mark sections with near K-T clastic deposits. Patterned columns mark sections without near K-T clastic deposits.

of deposition of all primary (complete or near-complete) K-T boundary sections. These sections span depositional environments from nearshore (inner neritic), middle shelf, outer shelf to upper slope, to the deep sea. Note that clastic sediments and breccias are deposited primarily in inner to middle shelf regions and on continental margins, but are absent in the deep sea. The estimated depth of deposition is based upon benthic foraminifera, ostracodes, dinoflagellates, invertebrates, and paleodepth backtracking of the deep-sea sections.

Determining relative sea-level change across the K-T boundary is more difficult and involves detailed quantitative studies of benthic foraminifera, shallow-water planktonic foraminifera, spores and pollen, dinoflagellates, and macrofossils. Such studies have been done for many K-T boundary sections, and they reveal a consistent pattern of global sea-level changes (Fig. 4) based on sections from northern to southern high latitudes (MacLeod and Keller, 1991a, 1991b; Schmitz and others, 1992; Keller, 1988b, 1992, 1993; Brinkhuis and Zachariasse, 1988; Keller and others, 1993b; Askin, 1992; Askin and Jacobson, this volume; Stinnesbeck and Keller, this volume). This database indicates generally rising seas after a sea-level lowstand, which Baum and others (1982) and Haq and others (1987) estimated occurred 0.5 m.y. below the K-T boundary, and Pardo and others (this volume) estimate at 0.2 to 0.3 m.y. below the K-T boundary based on

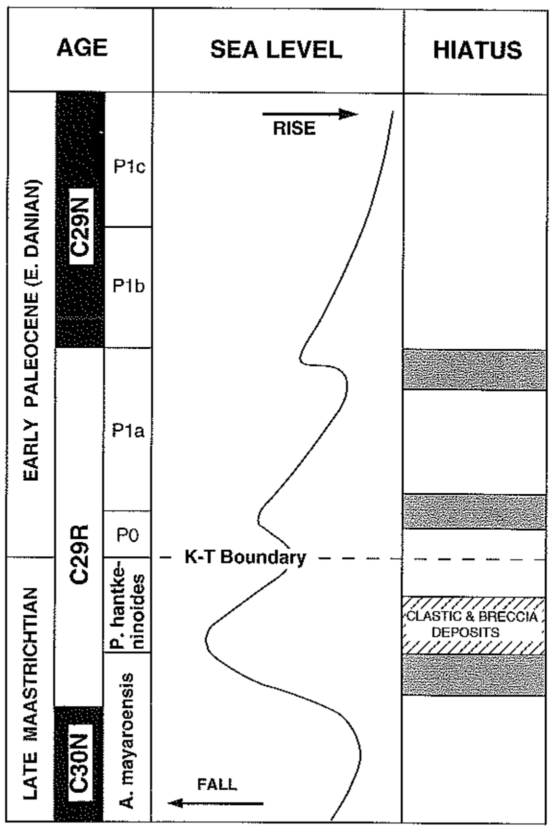

Figure 4. Sea-level changes, hiatuses, and clastic or breccia deposits across the K-T transition based on examination of K-T sections globally. Paleodepth interpretations are based on planktonic and benthic foraminifera, spores and pollen, dinoflagellates, and macrofossils. Biostratigraphy and paleomagnetic stratigraphy are from the Agost section in Spain (Pardo and others, this volume).

paleomagnetic stratigraphy. This rising sea level, beginning during the last 50 to 100 k.y. of the Maastrichtian, was interrupted by at least two short-term sea-level lowstands in the early Danian. In the stratigraphic record, these short-term sea-level lowstands are generally marked by hiatuses that removed part or entire biozones, and/or by condensed sedimentation (as expressed by a type-Z sequence boundary). Thus, they most likely represent sea-level fall and rise inflection points (Posamentier and others, 1988).

We recognize that local tectonics, especially along plate margins and on continental shelves, may amplify, obscure, or obliterate global sea-level signals when subsidence or uplift rates exceed the eustatic (global) rate. To avoid major tectonic overprints in our database, we have searched for consistent sea-level-change signals across latitudes that would indicate global controls. The sea-level changes identified are consistent in both magnitude and timing and thus suggest mainly global controls, any local tectonic overprint being relatively minor.

Most continental shelf and slope sections indicate a major sea-level lowstand, often accompanied by a hiatus, during the latest Maastrichtian preceding the K-T boundary. This sea-level lowstand occurred about 200–300 k.y. below the K-T boundary (Keller, 1989a; Keller and others, 1993b; Pardo and others, this volume), as indicated by its position in the lower part of C29R just below the *P. hantkeninoides* biozone. This latest Maastrichtian sea-level lowstand and hiatus is global in extent. At Stevns Klint and Nye Kløv in Denmark, this hiatus and sea-level lowstand is marked by a disconformity between white chalk beds that contain only rare bryozoans and the overlying undulating gray-white chalks that are rich in bryozoans (Surlyk, 1979; Hultberg and Malmgren, 1986; Schmitz and others, 1992; Keller and others, 1993b). At Brazos River, the sea-level lowstand and unconformity occurs at the base of a clastic deposit and within the portion of C29R below the K-T boundary (Keller, 1989a, 1989b). In a K-T boundary section at the Poty Quarry in Brazil (near Recife), a sea-level lowstand and unconformity occurs 70 cm below the K-T boundary within the *P. hantkeninoides* zone and is marked by a micritic limestone that is overlain by a 10–20-cm-thick limestone breccia (Stinnesbeck, 1989; Stinnesbeck and Keller, this volume). In K-T boundary sections from east-central Mexico (Mirador near La Ceiba) and southern Mexico (Bochil in Chiapas), evidence for a latest Maastrichtian sea-level lowstand is found in deposition of the clastic and breccia deposits 100 cm below the K-T boundary (Macias Pérez, 1988; Montanari and others, 1994; Stinnesbeck and others, 1994a). In most other K-T sections in northeastern Mexico, coarse-grained clastic deposits have been reported at or below the K-T boundary (Smit and others, 1992; Stinnesbeck and others, 1993, this volume; Keller and others, 1994a), as will be discussed below.

After the latest Maastrichtian sea-level lowstand, a marine transgression marks the interval up to and across the K-T boundary and through the earliest Paleocene zone P0. At Stevns Klint and Nye Kløv, the pre-K-T boundary transgression is marked by an influx of rugoglobigerinids from lower latitudes, as well as by benthic foraminiferal changes indicating deeper waters and climatic warming (Schmitz and others, 1992; Keller and others, 1993b). Deeper water benthic and palynofloral assemblages are also present in earliest Danian sediments of Spain (Agost and Caravaca), Tunisia (El Kef), Israel (Negev), and Texas (Brazos River)

(Peypoquet and others, 1986; Brinkhuis and Zachariasse, 1988; Keller, 1988b, 1992; Beeson, 1992). Biozone P0, which nearly always consists of a gray to black organic-rich clay with low-oxygen-tolerant faunas, appears to mark a sea-level highstand. Zone P0 is generally absent in settings below 1000 m depth because of nonaccumulation during the transgression (MacLeod and Keller, 1991a, 1991b).

The transition from the black clay of zone P0 to the shales of zone P1a is generally marked by a short hiatus and marks the first short-term Paleocene sea-level lowstand (Fig. 4). Erosion or condensed sedimentation due to decreased planktonic productivity, reduced upwelling, or decreased terrestrial input, frequently results in all or part of zone P0 being missing, as observed in sections from Brazos River (Keller, 1989a), Spain (Canudo and others, 1991), Israel (Keller and Benjamini, 1991), Denmark (Schmitz and others, 1992; Keller and others, 1993b), Mexico (Keller and others, 1994a, 1994b), the Gulf of Mexico (Keller and others, 1993a), and the southern Indian Ocean (Keller, 1993). Transgressing seas marked zone P1a (*P. eugubina*) with the deposition of increasingly shaly sediments that are frequently rich in glauconite (e.g., Brazos, Negev, Spain, Tunisia, Seymour Island).

The second early Paleocene hiatus and sea-level lowstand nearly coincides with the zone P1a-P1b boundary just prior to the extinction of *Parvularugoglobigerina eugubina* (Fig. 4). In many sections, erosion has removed part or all of zone P1a, including sections in Denmark, Spain, Israel, Mexico, Texas, and the southern Indian Ocean (Keller, 1989a, 1989b; MacLeod and Keller, 1991a, 1991b; Canudo and others, 1991; Keller and Benjamini, 1991; Keller and others, 1993b, 1994a, 1994b), and frequently removed sediments into the upper Maastrichtian (Keller and others, 1993a)

Increasingly carbonate rich marls or limestones are deposited in zones P1b and P1c across latitudes, and benthic foraminifera and palynofloras indicate deeper water environments approaching depths similar to the latest Maastrichtian prior to the sea-level fall (Brinkhuis and Zachariasse, 1988; Keller, 1988b, 1992).

Although these hiatuses mark the physical expression of sea-level fall and rise inflection points, the same relative changes in sea level can be observed in sections where no hiatuses are apparent in the biostratigraphic record. For example, the El Kef stratotype represents the most complete K-T boundary sequence known to date with no apparent hiatuses (Donce and others, 1985, 1994; Brinkhuis and Zachariasse, 1988; Keller, 1988a; MacLeod and Keller, 1991a, 1991b; Keller and others, in prep.), although the same relative changes in sea level are indicated in benthic foraminiferal and dinoflagellate assemblages (Brinkhuis and Zachariasse, 1988; Brinkhuis and Leereveld, 1988; Keller, 1988b, 1992). Similar sea-level fluctuations based on dinoflagellate assemblages are also reported from Alabama boundary sections (Habib and others, 1992), Brazos, Texas (Beeson, 1992), and Seymour Island, Antarctica (Askin, 1992; Askin and Jacobson, this volume). Thus, the sea-level lowstands of the latest Maastrichtian and early Danian (Fig. 4) record fluctuations in eustatic sea levels, and their effects on sedimentation must be considered in interpreting K-T clastic deposits.

Relative changes in sea level, timing of sea-level lowstands, and the relative magnitude of sea-level falls across the K-T transition, based on the global database discussed above, are shown in Figure 4. The latest Maastrichtian sea-level

lowstand occurred within the lower part of paleomagnetic chron 29R below the K-T boundary and just below the *P. hantkeninoides* zone, which spans the last 170–200 k.y. of the Maastrichtian (Pardo and others, this volume). Thus, this sea-level drop must have occurred within ~100 k.y. or less, and, on the basis of benthic foraminiferal data, reached a magnitude of ~70 to 100 m (Keller, 1992; Keller and others, 1993b; Schmitz and others, 1992; Stinnesbeck and Keller, this volume). At this magnitude, the sea-level fall would have occurred at a rate of 0.7 to 1.0 m/k.y. (or less than 1 mm/yr).

Generally rising seas characterize the top of the Maastrichtian across the K-T boundary and through the early Danian, reaching the pre-K-T boundary sea-level high by zone P1c (C29N), ~500 k.y. after the K-T boundary. This trend is interrupted by two short-term sea-level lowstands at the zone P0–P1a boundary, ~40 to 50 k.y. above the K-T boundary, and near the top of zone P1a, ~230 k.y. above the K-T boundary. The magnitude of these short-term sea-level drops is significantly less than during the latest Maastrichtian.

What caused the relatively rapid sea-level drop (less than 1 mm/yr) near the end of the Maastrichtian? A sea-level change due to mid-ocean ridge activity, which averages 1 cm/k.y., is too slow to be a causal agent. Changes in continental ice volume must be considered. Sea-level change due to buildup in continental ice volume during the Pleistocene averages 1 cm/yr. In comparison, the late Maastrichtian sea-level change averaged less than 1 mm/yr, or less than one-tenth of Pleistocene rates.

There is evidence for significant continental glaciation during the middle Maastrichtian in Weddell Sea Ocean Drilling Program (ODP) Site 690, in a marked cooling and increase in $\delta^{18}O$ values associated with a major sea-level drop (Barrera and Huber, 1990; Barrera, 1994). The magnitude of the $\delta^{18}O$ increase is similar to that of the middle Eocene cooling and major Antarctic glaciation, which is associated with ice-rafted debris (Barrera and Huber, 1991). This suggests that significant continental glaciation may also have occurred during the middle Maastrichtian (Barrera, 1994). Stable isotope values indicate that relatively cool temperatures prevailed through the late Maastrichtian, suggesting that the sea-level drop just preceding the *P. hantkeninoides* zone is likely the result of increased continental glaciation. At ODP Site 690, stable isotope values record a rapid warming (greenhouse warming?) just preceding the K-T boundary and following the maximum late Maastrichtian cooling (Stott and Kennett, 1990; Barrera, 1994), and at Stevns Klint and Nye Kløv a rising sea level and incursion of tropical species also mark this pre-K-T boundary climatic warming (Schmitz and others, 1992; Keller and others, 1993b).

The cause for this climatic warming is still unclear. It is most likely related to the extensive global volcanism (Deccan Traps, India) and the associated increase in atmospheric carbon dioxide levels at that time, although a similar atmospheric carbon dioxide increase and climatic warming could have been triggered by a pre-K-T boundary bolide impact.

CLASTIC DEPOSITS

Clastic coarse-grained or breccia deposits have been described from many K-T boundary sequences: the best known are from Texas, Alabama, Georgia, Cuba, Haiti, and Mexico (Fig. 5), where they have been variously related to sea-level changes or impact-generated megatsunami deposits. The geographic concentration of such K-T sections is not accidental, but rather the result of many years of intensive search for evidence of a bolide impact in the Caribbean in the form of tsunami deposits. A similarly intensive search elsewhere may yield similar K-T or near K-T clastic deposits in other marginal seas beyond the reach of a proposed Caribbean megatsunami. In this section we review the age, biostratigraphy, lithology, and depositional environment of these clastic deposits to evaluate whether they are coeval, as assumed by proponents of the tsunami model and the late Maastrichtian sea-level lowstand model, or whether they represent different sea-level lowstand ages below, at, or above the K-T boundary.

Alabama: Braggs, Mussell Creek, Moscow Landing

Clastic deposits in Alabama are restricted to the Clayton Sands, which are widespread though intermittent throughout the region. They occur near the K-T boundary and have been interpreted as transgressive infilling of incised valleys cut during a preceding sea-level lowstand (Donovan and others, 1988; Baum and Vail, 1988; Mancini and others, 1989; Savrda, 1991, 1993; Moshkovitz and Habib, 1993; Habib and others, 1992) and as high-energy impact-generated megatsunami deposits (Hildebrand and Boynton, 1990; Olsson and Liu, 1993; Habib, 1994; Smit and others, 1994a, 1994b).

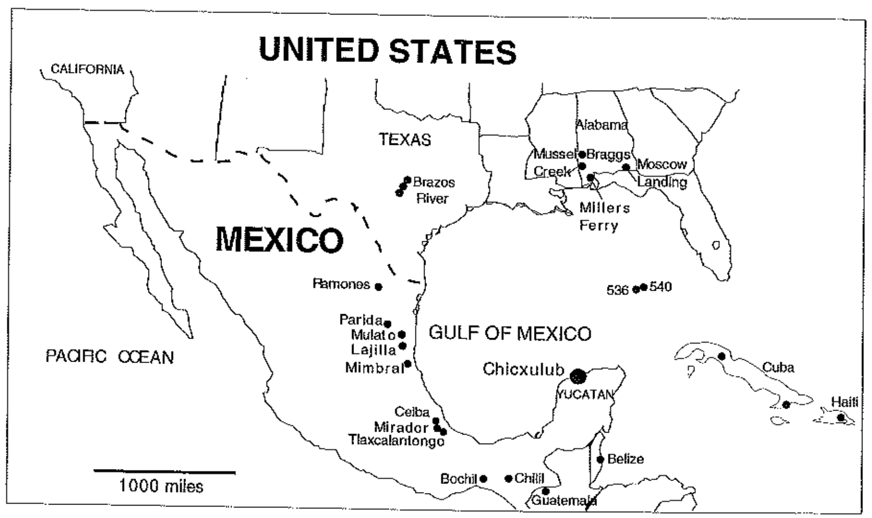

Figure 5. Location of K-T boundary sections in the Caribbean and Central America from which coarse-grained clastic or breccia deposits have been described and interpreted as impact-generated megatsunami deposits.

The lithology and stratigraphic sequences are similar in the Alabama sections, including Braggs, Mussell Creek, Millers Ferry, and Moscow Landing (Mancini and others, 1989; Savrda, 1993; Olsson and Liu, 1993). Here, the gray to olive-gray silty to sandy marls of the late Maastrichtian Prairie Bluff Chalk are intensively burrowed and contain abundant Cretaceous body fossils (principally bivalves). On the basis of the absence of *Abathomphalus mayaroensis*, Mancini and others (1989) concluded that the upper Maastrichtian is missing at the unconformity at the top of the Prairie Bluff Chalk. However, the presence of the latest Maastrichtian *Micula prinsii* nannofossil zone in the uppermost part of the Prairie Bluff Chalk indicates that this "hiatus" is much shorter (Moshkovitz and Habib, 1993; Olsson and Liu, 1993). On the basis of the irregular geometry of consolidated marlstone and truncation of thick-shelled fossils, Savrda (1993) estimated at least 2–3 m of erosion. Furthermore, he estimated an extended period of subaerial exposure on the basis of microkarstification (solution-enhanced erosion).

At Mussell Creek and Moscow Landing, and many other localities in central and western Alabama, the Prairie Bluff is unconformably overlain by the Clayton Sands, which consist of irregular bodies of uncemented yellowish-brown silty sands. At Moscow Landing, this member is characterized by inclined parallel and ripple cross-laminations, rip-up clasts of the Prairie Bluff Chalk, and reworked Cretaceous fossils. The resident ichnofauna includes rare *Ophiomorpha* burrows in the lower beds and *Thalassinoides* and *Planolites* burrows in the middle and upper beds of the Clayton Sands (Savrda, 1993). Sediments and fossils both indicate that deposition occurred in a marginal marine tidally influenced environment (Savrda, 1993; Moshkovitz and Habib, 1993).

The Clayton Sands are now generally considered to be of early Danian age. The Clayton Sands exposed at Mussell Creek and Moscow Landing contain planktonic foraminifera indicative of zones P1b and P1c (*Subbotina pseudobulloides*, *S. triloculinoides*, *S. trivialis*, and *Globoconusa daubjergensis*; Mancini and others, 1989), whereas at Millers Ferry the earliest Danian zones P0 and P1a (*P. eugubina*) span the entire Clayton Sand, as reported by Olsson and Liu (1993) and Liu and Olsson (1992). This suggests that the older (zones P0 and P1a) Clayton Sand is present at Millers Ferry, but absent at Mussell Creek, Braggs, and Moscow Landing, whereas the younger (zone P1b or P1c) Clayton Sand occurs in the latter outcrops but not at Millers Ferry. This, in turn, implies that the Clayton Sands are not coeval in all sections and may represent a time-transgressive facies spanning the interval from zone P0 to at least zone P1b in the early Danian.

At all four Alabama sections (Mussell Creek, Braggs, Moscow Landing, Millers Ferry), the Clayton Sands and/or Prairie Bluff Chalk are overlain by the basal limestone bed of the Clayton Formation, which truncates the Clayton Sands and Prairie Bluff Chalk. This contact is sharp and clearly erosional at Mussell Creek, Braggs, and Moscow Landing (Mancini and others, 1989; Savrda, 1993), and the abrupt lithological change at Millers Ferry may also indicate discontinuous sedimentation.

Is deposition of the Clayton Sands the result of a K-T boundary impact-generated tsunami? The presence of Danian microfossils in the Clayton Sands alone argues against deposition caused by a K-T boundary impact. Moreover, no shocked quartz, tektite glass, or other impact indicators have been found. The presence of microkarstification and a firmground at the Prairie Bluff–Clayton Sands contact argues for an extended period of subaerial exposure rather than a catastrophic event (Savrda, 1993). As indicated by zones P0–P1a and P1b–P1c planktonic foraminifera in the Clayton Sand, deposition occurred over an extended time period spanning at least 200–300 k.y. An extended period of deposition is also indicated by the resident ichnofauna, characterized by the presence of multiple phases of habitation and permanent domiciles (Savrda, 1993).

Sedimentological and biological indicators strongly indicate that deposition of the Clayton Sands occurred over an extended time period bounded by disconformities formed during two sea-level lowstands. The erosional surface at the top of the Prairie Bluff most likely represents the latest Maastrichtian sea-level lowstand (Baum and Vail, 1988; Donovan and others, 1988) characterized by microkarstification and prolonged subaerial exposure. By early Danian time sediment deposition resumed with the Clayton Sands; deposition occurred during the sea-level transgression and was probably intermittent. The disconformity between the top of the Clayton Sands and the basal limestone of the Clayton Formation marks a sea-level lowstand at or near the top of zone P1a as indicated by the presence of *P. eugubina* at Millers Ferry (Olsson and Liu, 1993). At Mussell Creek and Moscow Landing, deposition of Clayton Sands may have continued into zone P1b or P1c as suggested by the absence of *P. eugubina* (Mancini and others, 1989).

Donovan and others (1988) and Baum and Vail (1988) interpreted the upper Prairie Bluff Chalk as a prograding highstand systems tract (HST) that was followed by a major sea-level fall that truncated the sequence, forming an irregular locally incised topography that was temporarily exposed (type 1 sequence boundary). During the subsequent transgression, these incised valleys were filled with marginal marine sands and gravels (lowstand systems tract, LST). Further rise of sea level formed the transgressive erosional surfaces and firmgrounds at the base of the Clayton limestone (transgressive systems tract, TST) and was followed by the return to normal marine conditions. This interpretation is consistent with the biostratigraphic and paleontological evidence.

Texas: Brazos River

Outcrops along the Brazos River in central-east Texas contain a nearly continuous record across the K-T boundary (Jiang and Gartner, 1986; Keller, 1989a, 1989b; Hansen and Upshaw, 1990; MacLeod and Keller, 1991a, 1991b; Hansen and others, 1993; Beeson, 1992). The K-T transition consists of calcareous claystones of the Kinkaid Formation, with a thin (<20 cm) clastic unit at its base. The clastic unit rests unconformably upon a scoured surface of the Maastrichtian Corsicana Formation and consists of coarse sand containing shell hash and shale intraclasts capped by a thin limestone layer. This clastic event deposit has been variously interpreted as an impact-generated tsunami deposit (Bourgeois and others, 1988; Hildebrand and Boynton, 1990) linked to the proposed Chicxulub impact structure (Smit and others, 1992, 1994a, 1994b), or to noncatastrophic sedimentation related to the latest Maastrichtian sea-level lowstand and subsequent transgression (Keller, 1989a, 1992). No shocked quartz, Ni-rich spinels, or

tektite glass have been found within the clastic unit. A recent study indicates an Ir anomaly 15–17 cm above the limestone layer that marks the top of the clastic unit (Rocchia, 1994, written commun.).

There has been disagreement in the past as to whether the K-T boundary should be placed at the unconformity at the base of the clastic deposit, at the top of the clastic deposit, or between 15 and 22 cm above the deposit. The general consensus seems to be to place the K-T boundary 17 to 20 cm above the top of the clastic deposit at the first appearance of Tertiary foraminifera (Keller, 1989a, 1989b), nannofossils (Jiang and Gartner, 1986), and palynomorphs (Beeson, 1992; Figure 6), coincident with the Ir anomaly as observed by Rocchia (1994, written communication).

All planktonic foraminifera observed by Keller (1989a) in the clastic deposits of three Brazos River sections are Late Cretaceous forms. Also present is the latest Maastrichtian index species *Plummerita hantkeninoides*, which indicates that deposition occurred during the last 170–200 k.y. of the Maastrichtian. Montgomery and others (1992), however, reported late Danian planktonic foraminifera, suggesting that deposition occurred during the late Danian. Examination of Montgomery and others' (1992) data indicates, however, that the species illustrated as Danian

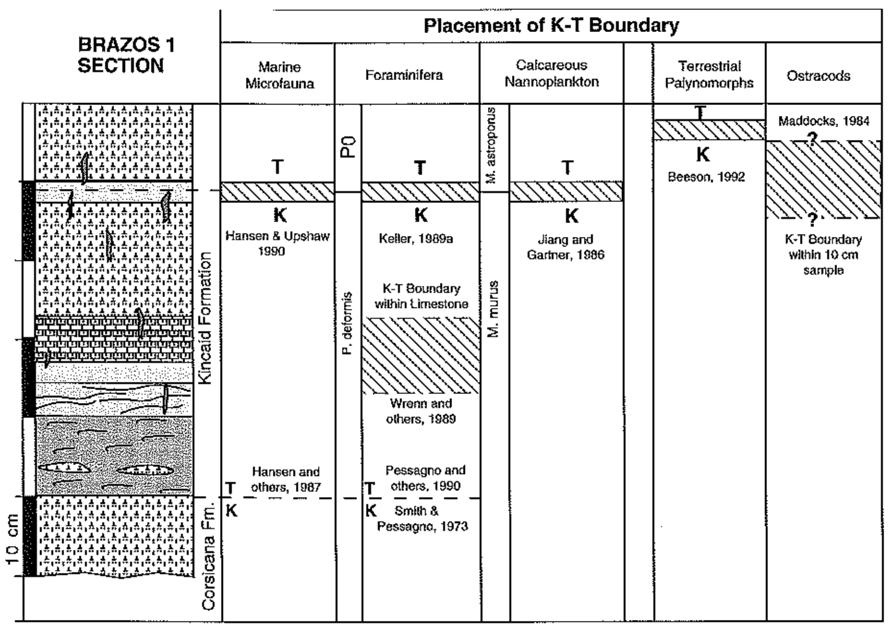

Figure 6. Placement of the K-T boundary in the Brazos-1 section, Falls County, Texas. Note the general biostratigraphic consensus between macrofossils, planktonic foraminifera, nannofossils, palynomorphs, and ostracodes in placing the K-T boundary between 15–20 cm above the clastic deposit. The paleontological boundary placement coincides with the iridium anomaly 15–17 cm above the clastic deposit as reported by Rocchia (1994, written commun.).

taxa are either too poorly preserved for positive species identification or lack precise location data relative to the clastic deposit, and in some cases are from Danian strata above the clastic deposit (MacLeod and Keller, 1994). Thus, no confirmed Danian microfossils are known in this clastic deposit.

Between the clastic deposit and the K-T boundary, the marly sediments contain well-preserved late Maastrichtian shallow-water benthic foraminifera and surface-dwelling planktonic foraminifera (heterohelicids, guembelitrids, rugoglobigerinids, hedbergellids, globigerinellids, and *P. hantkeninoides*). There is no evidence of size sorting by currents or upward fining of sediments. Above the K-T boundary, the same well-preserved late Maastrichtian assemblages continue (except *P. hantkeninoides*) with the addition of the evolving new Tertiary species. All early Danian zones are present, including zones P0, P1a, P1b, and P1c. A short hiatus is present at the zone P1a–P1b boundary, indicated by a lithological change, abundant glauconite, and the sudden truncation of species ranges (Keller, 1989a). A short hiatus or condensed sedimentation is also present at the P0–P1a boundary (Fig. 7; MacLeod and Keller, 1991a, 1991b).

At the Brazos River section, catastrophic tsunami deposition is difficult, if not impossible, to reconcile with the biostratigraphy and the absence of tektite glass and shocked quartz. Proponents of this theory explain the sediments between the top of the clastic bed and the K-T boundary as the result of settling through the water column after the tsunami (Bourgeois and others, 1988; see also Keller, 1991; Smit and others, 1994b), and a similar explanation is given for the 5–100-cm-thick Maastrichtian sediments above the clastic deposits in Mexico (Smit and others, 1992, 1994a, 1994b; Montanari and others, 1994). Because there is no evidence of impact-derived material or size sorting, and the dwarfing of some

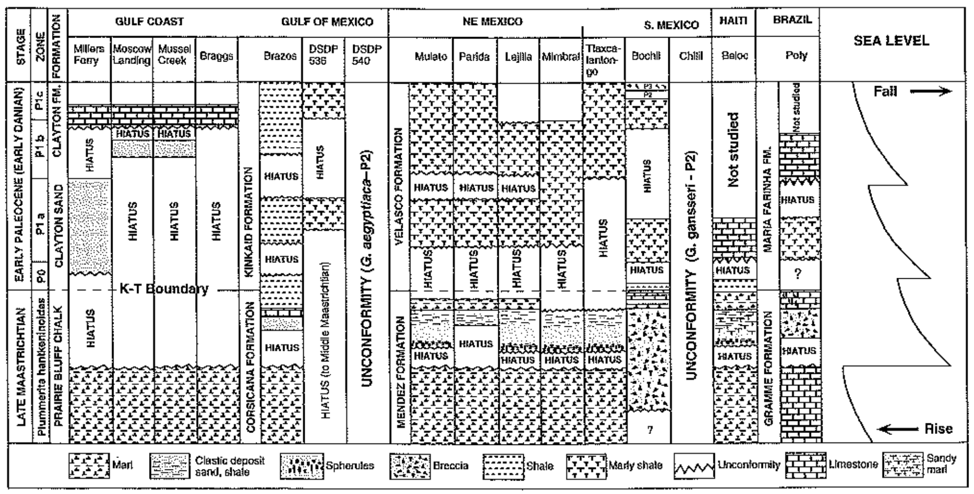

Figure 7. Biostratigraphy, lithostratigraphy, and hiatuses (white intervals) of K-T sections with clastic or breccia deposits correlated with sea-level fluctuations. Note that the clastic or breccia deposits, which have been reinterpreted by some workers as impact-generated tsunami deposits, occur stratigraphically below the K-T boundary and correlate with the latest Maastrichtian sea-level lowstand.

species (e.g., heterohelicids) begins below the clastic deposit at the Brazos sections, catastrophic impact-generated tsunami deposition is unlikely. Prevailing evidence suggests that the unconformity and clastic deposit are related to the latest Maastrichtian sea-level regression and subsequent transgression across the K-T boundary. On the basis of magnetostratigraphy at the Brazos core section, the unconformity formed in C29R just below the K-T boundary, and an estimated 300 k.y. of the latest Maastrichtian may be missing (Keller, 1989a). Sediment deposition at Brazos thus differs from Alabama sections where Montgomery and others (1992) suggested that a major Upper Cretaceous to Danian unconformity is present based on the absence of *A. mayaroensis*. Stable isotopic data indicate, however, that this species is a deeper water dweller probably living at depths of 200 m or deeper (Boersma and Shackleton, 1981; Barrera and Huber, 1990) and that its absence in the shallow waters of the Gulf of Mexico coastal plain is likely due to ecologic exclusion, rather than a hiatus. This is also indicated by the presence of the latest Maastrichtian nannofossil index species *Micula prinsii* (Savrda, 1993; Olsson and Liu, 1993) and planktonic foraminifer *P. hantkeninoides*.

A latest Maastrichtian sea-level fall at Brazos is indicated by middle to inner neritic benthic foraminifera (Keller, 1992), a decline in invertebrate diversity, and an increase in oysters (Hansen and others, 1993). These shallowing waters may have been accompanied by reduced salinity as suggested by the decrease in stenohaline carnivorous snails (Hansen and others, 1993). Paleobathymetric trends among benthic foraminifera and palynomorph assemblages indicate that the sea-level lowstand was followed by rapidly rising seas across the K-T boundary (Keller, 1992). Short-term sea-level lowstands appear to have occurred at the P0–P1a and P1a–P1b boundary (Fig. 7), as indicated by abrupt changes in planktonic foraminifera at both horizons, and a lithological shift to sand and glauconite at the P1a–P1b boundary (Keller, 1989a; MacLeod and Keller, 1991a, 1991b).

Cuba

A big boulder bed present in the uppermost Cretaceous sediments of Cuba has been interpreted as a proximal ejecta blanket of a nearby K-T boundary bolide impact (Bohor and Seitz, 1990). This boulder bed has been shown, however, to be the result of exfoliation weathering in the coarse calcarenaceous part of a megaturbidite (Brönnimann and Rigassi, 1963; Dietz and McHone, 1990; Iturralde-Vincent, 1992). This megaturbidite forms a sheet-like unit that is widely known in Cuba as the Cacarajicara, Amaro, and Peñaver formations (Pszolkowski, 1986; Iturralde-Vincent, 1992).

Whatever the cause of the sudden collapse of sediment prisms on the platform and slope of Cuba, whether earthquake, bolide impact, or sea-level lowstand, it did not coincide with the K-T boundary. Planktonic foraminifera from the top of the Cacarajicara Formation indicate an early to middle Maastrichtian age (Pszolkowski, 1986), whereas the top of the Peñaver and Amaro formations, which contain the megaboulder bed, are of late Maastrichtian *A. mayaroensis* zone age, and contain the index taxon (Brönnimann and Rigassi, 1963; Iturralde-Vincent, 1992). Moreover, the Amaro Formation and megaboulder bed is overlain by a 150–250-m-thick limestone unit that also contains late Maastrichtian planktonic foraminifera (Iturralde-Vincent, 1992). Thus, the megaboulder bed of

Cuba is of late Maastrichtian age and unrelated to the K-T boundary event. Continuous K-T boundary sequences are present in marly sediments in eastern and southern Cuba that contain no evidence of the megaboulder bed, or any other clastic deposit (Iturralde-Vincent, 1992; Fernandez and others, 1991).

Haiti: Beloc

The K-T boundary sections of Beloc in Haiti contain a 40–60-cm-thick clastic sandstone deposit rich in calcareous and glassy spherules that have been variously reported as of tektite origin (Izett, 1990; Sigurdsson and others, 1991; Blum and Chamberlain, 1992; Koeberl and Sigurdsson, 1992) or of volcanic origin (Jéhanno and others, 1992; Lyons and Officer, 1992; Koeberl, 1994; and Robin and others, 1994). The clastic bed rests upon marls containing latest Cretaceous planktonic foraminiferal (*A. mayaroensis* zone) and nannofossil (*Micula murus* zone) assemblages (Maurasse and Sen, 1991; Sigurdsson and others, 1991). The spherule-rich sandstone is graded in size and abundance and has cross-laminations toward the top of the unit.

In most Beloc sections, the spherule-rich sandstone and overlying sediments are incomplete due to redeposition. However, two sections have been reported with a less disturbed and more expanded sedimentary record (Jéhanno and others, 1992). In these two sections, a 25–30-cm-thick bed of Maastrichtian sandy marl conformably overlies the clastic deposit, similar to sections in northeastern Mexico. This sandy marl layer is succeeded by a thin clay layer that contains the cosmic markers Ni-rich spinels and the Ir anomaly that characterize the K-T boundary elsewhere. Thus, the sandy marls below the K-T boundary indicate that a period of normal hemipelagic sedimentation followed deposition of the spherule-rich clastic unit which must predate the K-T boundary event by at least several thousand years. Moreover, faunal assemblages indicate that water depth decreased in this sandy marl as compared to the water depth below the spherule-rich clastic bed (Jéhanno and others, 1992). On the basis of these data, we suggest that the Haiti spherule-rich clastic sediments were deposited during the latest Maastrichtian sea-level lowstand (Fig. 7) that may have triggered gravity flow or turbidite deposition containing spherules from an earlier or coeval widespread Caribbean volcanic event or ejecta from a pre-K-T boundary impact.

Northeastern and East-Central Mexico

More than a dozen K-T boundary sections containing siliciclastic deposits at or near the boundary are known from northeastern Mexico and two are known from east-central Mexico (Mirador and Tlaxcalantongo near La Ceiba, Fig. 5). Most of these sections show comparable depositional sequences, although clastic units are variable in lithology and thickness, ranging from only 2 cm to a maximum of 11 m thick. Additional sections have no clastic deposit. The geographic distribution of these sections spans over 500 km in a north-northwest–south-southeast–trending area that is parallel to and 40–80 km east of the front range of today's Sierra Madre Oriental (Keller and others, 1994a).

The regionally developed clastic member has been described by a number of

Longoria and Grajales Nishimura, 1993; Keller and others, 1994a, 1994b; Adatte and others, 1994) and has been interpreted as a K-T boundary deposit produced by an impact-generated megatsunami (Smit and others, 1992, 1994a), a series of normal high-energy turbidity currents triggered by an impact (Bohor and Betterton, 1993), and a series of gravity flows or turbidity currents related to the latest Maastrichtian sea-level lowstand and tectonic activity (uplift of Sierra Madre) (Stinnesbeck and others, 1993, 1994a, this volume; Keller and others, 1994a; Adatte and others, 1994).

In sections where no clastic deposit is present at the K-T boundary (e.g., Parida where clastic deposit disappears), the K-T boundary occurs between the fine-grained marls of the Maastrichtian Mendez Formation and the overlying marly shales of the Tertiary Velasco Formation. Although the contact between these formations is transitional and no erosional surface is recognizable, the earliest Danian zone P0 is generally not present and the lower part of zone P1a also appears to be missing, either due to a hiatus or condensed sedimentation (Fig. 7, type Z sequence boundary, Lopez-Oliva and Keller, 1994). The clastic deposit is stratigraphically near the K-T boundary. In the most complete sections, a 5–10-cm-thick layer of Maastrichtian marls overlies the clastic deposit (e.g., Lajilla, Mulato, Parida, in northeastern Mexico), and a 1-m-thick Maastrichtian age sequence of sand, shale and clay overlies a breccia deposit at Bochil in Chiapas, Mexico. In some sections (e.g., Mimbral), Tertiary shales of the Velasco Formation directly overlie the clastic deposit and in still other sections, Tertiary sediments are missing (e.g., Peñon, Ramones, Sierrita; Keller and others, 1994a). The latest Maastrichtian index species *P. hantkeninoides* has been oberved in all sections (except Mulato) within, below, or above the clastic member, but always below the K-T boundary (Lopez-Oliva and Keller, 1994). The presence of this species indicates that deposition of the clastic member occurred within the last 170–200 k.y. of the Maastrichtian.

Although the thickness and character of the clastic member vary between individual outcrops, three distinct lithologic units can be recognized in the thickest deposits (e.g., Mimbral, Peñon, Mulato, Sierrita; Stinnesbeck and others, 1993, this volume; Keller and others, 1994a). The basal unit 1, which consists of a carbonate-spherule-rich weathered sediment, is the key evidence for linking the clastic deposit to a bolide impact. This layer is of variable thickness, but generally no more than 10–30 cm thick, and contains abundant spherules. The spherules are 1–5 mm in diameter, commonly infilled with blocky calcite, and surrounded or partly filled by mixed layers of illite/smectite, chlorite, palagonite, or mica (Stinnesbeck and others, 1993; Keller and others, 1994a), or altered to kaolinite, replaced by smectite, pyrite, and glauconite. Many large spherules contain several smaller spherules. Some spherules contain apatite concretions, rutile crystals, clasts of limestone, or foraminiferal tests filled with glauconite. Many spherules have a tan organic coating, indicating that they are calcite infilled algal resting cysts (Stinnesbeck and others, 1993, 1994b; Keller and others, 1994a). Thus, the spherules of unit 1 have multiple origins, including organic, authigenic, volcanic, and accretionary oolites that formed at neritic depths and were probably subse-

Rare glass particles in the spherule-rich layer are similar in chemical composition to glass spherules from Beloc, Haiti (Smit and others, 1992; Bohor and Betterton, 1993; Stinnesbeck and others, 1993). Many workers consider these glass particles and spherules as indicative of impact origin (Izett, 1990; Blum and Chamberlain, 1992; Sigurdsson and others, 1991; Koeberl and Sigurdsson, 1992), whereas others consider them to be indicative of volcanic origin (Jéhanno and others, 1992; Lyons and Officer, 1992; Leroux and others, 1995). For a discussion of this controversy see Robin and others (1994) and Keoberl (1994). If proven to originate from impact, these rare glass shards represent the only potential evidence to date of an impact in the clastic deposits of northeastern and east-central Mexico. A small iridium anomaly (0.8 ppb; Smit and others, 1992; Keller and others, 1994a; Stinnesbeck and others, 1993) is present above the clastic deposit in the basal shales of the Tertiary Velasco Formation.

A 10–20-cm-thick sandy limestone layer containing microlayering of fine and coarse-grained sediments and few spherules (W. Ward, 1994, personal commun.) is present within this spherule-rich layer at the Mimbral, Peñon, and Lajilla sections. We suggest that this sandy limestone layer represents a distinct and separate event during deposition of the spherule-rich sediments of unit 1. It seems unlikely that these distinct depositional events are the products of air-borne fallout, settling through the water column and reworking by wave action, all prior to the arrival of the first tsunami wave within two hours of the bolide impact event, as suggested by Smit and others (1994a). Deposition of these sediments probably occurred over a longer time period and by multiple events, as also suggested by the presence of abundant material from shallow-water areas that was transported and redeposited in the deeper water areas of Mimbral, Peñon, Lajilla, Mulato, Sierrita, and Tlaxcalantogo.

In northeastern Mexico sections, an erosional surface marks the contact between unit 1 and the overlying unit 2, which consists of a weakly laminated sandstone devoid of spherules, glass, or iridium. Plant debris is present in distinct layers at the base of unit 2 at the Mimbral outcrop and mud clasts are frequently present at the unconformity in all sections. Moreover, burrows, some infilled with the underlying spherule-rich sediment, are present and indicate habitation by invertebrates during deposition of unit 2 (Ekdale and Stinnesbeck, 1994). An erosional unconformity also marks the contact between unit 2 and the overlying unit 3, which consists of alternating sand, silt, and shale layers topped by a rippled sandy limestone. Ripple marks, flaser bedding, convolute bedding, climbing ripples, and small-scale cross-bedding are commonly found. Unit 3 is lithologically, mineralogically, and petrographically most variable, and contains thin layers of hemipelagic sediments with rich late Maastrichtian foraminiferal assemblages and two distinct layers rich in zeolites that are correlatable over the region and represent an influx of volcanogenic sediments (Adatte and others, 1994, 1995). The rippled sandy limestone layer that caps unit 3 is bioturbated in all outcrops, and burrows of *Chondrites*, *Zoophycos*, *Thalassinoides*, and *Ophiomorpha* commonly found in two to three discrete layers indicate different levels and different times of habitation. There is no downward mixing of Tertiary sediments (Stinnesbeck and others, 1993; Keller and others, 1994a, 1994b). Benthic

lower depth than either the underlying Maastrichtian Mendez marls or the overlying Tertiary Velasco shales (Keller and others, 1994a, 1994b).

Our studies of more than a dozen K-T boundary outcrops in northeastern Mexico thus indicate that deposition of the clastic member occurred during the latest Maastrichtian sea-level lowstand within the last 170–200 k.y. of the Maastrichtian, but preceding the K-T boundary by at least several thousand years (Stinnesbeck and others, 1993, 1994b; Keller and others, 1994a, 1994b; Lopez-Oliva and Keller, 1994). Moreover, deposition occurred over an extended time period that allowed repeated recolonization of the substrate after erosion by burrowing organisms. Sedimentary processes also indicate multiple event deposition by debris flows, gravity flows, and periods of normal hemipelagic sedimentation. If the rare glass shards in unit 1 prove to be of impact origin, this impact would have preceded the K-T boundary. However, there is no evidence of a major pre-K-T boundary bolide impact in the biotic, stratigraphic, or geochemical records.

Southern Mexico: Chiapas

To date, there are no detailed studies of K-T boundary sequences of southern Mexico, despite visits by numerous geologists (including us) in search of outcrops with megatsunami deposits or continuous sedimentation. The problem is in the complex regional tectonics, which have resulted in repeated uplift and erosion and, hence, a major hiatus, or repeated flysch deposition (Quezada Muñeton, 1990; Michaud and Fourcade, 1989; Stinnesbeck and others, 1994a). We report here on the Chilil and Bochil sections of Chiapas (Fig. 5), which are representative of these two conditions.

The Chilil section, near Cristobal de las Casas, shows the transition between the Late Cretaceous Angostura Formation and the Paleocene Soyalo Formation. Our investigations indicate that the Angostura Formation consists of bioclastic shallow-water limestones with corals, rudists, echinoderms, and large benthic foraminifera. This unit is capped by a thin layer of bioturbated marly limestone with abundant early-late Maastrichtian (*G. gansseri* zone) planktonic foraminifera, indicating drowning of the carbonate platform. The top of this marly limestone contains crusts of large iron- and manganese-oxide nodules that formed during a long period of nondeposition. This surface is overlain by rhythmically bedded pelagic marls and shales of the Paleocene Soyalo Formation, which contains abundant planktonic foraminifera of the late Paleocene zone P2. The Chilil section thus contains a hiatus from the lower-upper Maastrichtian *G. gansseri* zone to the upper Paleocene zone P2 (Fig. 7). We found no evidence of the continuous K-T boundary deposition reported by Longoria and Gamper (1992).

The Bochil section north of Tuxtla Guiterrez contains a more complete K-T boundary record (Montanari and others, 1994; Stinnesbeck and others, 1994a). However, as at Chilil this K-T transition must be viewed within the tectonic and depositional history of the region. Flysch deposition began in the late Campanian and continued into the lower Tertiary, depositing many hundreds of meters of rhythmically bedded shales, silts, and sandstones (Quezada Muñeton, 1990). This flysch sequence is interrupted by numerous debris flows of well-rounded conglomerates in the lower part, and thick breccia beds of platform carbonates and deposits mark the gradual uplift, tilting, and final collapse of the nearby platform carbonates (Michaud and Fourcade, 1989) and sea-level changes. Within this sequence of repeated breccia deposits, the thick clastic breccia containing angular platform limestone clasts (Montanari and others, 1994) near the K-T boundary does not mark an unusual sedimentological or lithological change, but rather a recurring phenomena, and hence provides no evidence for a single marine disturbance triggered by the proposed Chicxulub impact. These near K-T breccia deposits are likely the results of local tectonic activity during the latest Maastrichtian sea-level lowstand. This is supported by the fact that the breccia is overlain by a succession consisting of a 1-m-thick sandstone, silt, reddish clay, silty chalk, 5–10-cm-thick micritic limestone, thin clayey arenite, and white laminated chalk, the latter containing abundant early Danian zone P1a (*P. eugubina*) species (Montanari and others, 1994). Enriched iridium values possibly marking the K-T boundary were reported by Montanari and others (1994) from a reddish clay layer 6–10 cm below the micritic limestone of early Danian zone P1a.

These data indicate that K-T boundary age sediments are present at Bochil, although it is unclear whether sedimentation was continuous. At present, there is no evidence (no shocked quartz or glass) to link the breccia deposit 1 m below the K-T boundary to a K-T impact event (Fig. 7). Because the Bochil section contains many similar breccia deposits throughout the Campanian, Maastrichtian, and early Paleocene, the near K-T breccia layer, similar to the earlier and later deposits, most likely reflects sea-level changes and regional tectonic activity.

Guatemala and Belize

Cretaceous-Tertiary boundary deposits of Guatemala and Belize are still poorly understood, and no detailed studies exist. Preliminary investigations suggest that the depositional setting of the Petén region of Guatemala is similar to that of Chilil in Chiapas, Mexico. Michaud and Fourcade (1989) described Maastrichtian platform limestones and platform margin breccias overlain by rhythmically bedded Tertiary marls and shales in both regions. Neither they nor Hildebrand and others (1994) provided detailed biostratigraphic descriptions of the Petén sections. Hildebrand and others (1994, p. 49) stated, however, that "a dramatic deepening" occurred at K-T boundary time, giving evidence "of the dramatic erosional effects near the point of impact" of the Yucatan bolide. No explanation is provided how a bolide impact on Yucatan would cause dramatic deepening in Guatemala sections, nor did Hildebrand and others (1994) provide biostratigraphic evidence that the deepening is of K-T boundary age.

Our preliminary investigation of Guatemalan K-T boundary sections shows an 8–50-m-thick breccia deposit that is disconformably overlain by lower Tertiary sediments. The breccia consists of thick beds of large rounded clasts, angular clasts and interlayers of smaller-sized clasts, and pebbles in a muddy matrix. The top 2 m of the breccia unit consists of beds with smaller-sized breccia clasts and pebbles. Deposition of limestone clasts within the breccia occurred in a shallow-marine environment at depth of less than 20 m, as indicated by the predominance of miliolid benthic foraminifera and absence of planktonic foraminifera. Subsequent breccia deposition occurred at neritic depth via transport. The age of the breccia is unknown. Limestone or marl layers disconformably overlie the breccia unit, the contact being marked by an undulating erosional surface and breccia clasts in the overlying sediments. These sediments are of early Danian zone P1a age and benthic foraminiferal fauna indicates deposition in an outer neritic to uppper bathyal environment. Further studies are necessary before the age and nature of breccia deposition in Guatemala can be determined.

A K-T boundary section was reported by Ocampo and Pope (1994) from Albion Island in northern Belize. These authors described a K-T boundary marked by a major erosional unconformity separating crystalline Cretaceous dolomites and a poorly sorted dolomitic and carbonaceous breccia which they interpreted as the "product of ballistic sedimentation from the Chicxulub impact" (Ocampo and Pope, 1994, p. 86). It is interesting that, by the authors' own admission, their K-T boundary age interpretation is not based on any biostratigraphic evidence, either supporting a Maastrichtian age of the dolomite or the breccia deposits. Moreover, they did not observe Tertiary strata overlying the breccia deposit, nor did they observe them in other parts of the island. Ocampo and Pope's (1994, p. 86) interpretation of a breccia deposit as the "product of ballistic sedimentation from the Chicxulub impact" is a speculation based solely on a breccia of unknown age and the Chicxulub impact scenario.

Northeastern Brazil: Poty Quarry

The Poty Quarry section north of Recife in northeastern Brazil is outside the region of the disputed megatsunami deposits, but contains a limestone breccia bed 70 cm below the K-T boundary, coincident with the latest Maastrichtian sea-level lowstand. This section contains the most complete marine K-T boundary section known to date in South America (Stinnesbeck, 1989; Stinnesbeck and Keller, this volume). The upper Maastrichtian sediments (Gramame Formation) consist of micritic marly limestones (wackestones) that unconformably underlie a 20-cm-thick limestone breccia bed 70 cm below the K-T boundary (Fig. 7). The breccia bed contains reworked parts of the underlying marly limestone, calcispheres, planktonic foraminifera, bone fragments, phosphatic particles, glauconite, pyrite concretions, and abundant serpulids. In the 50 cm above the breccia bed, the grain size of limestone clasts diminishes from rudite to arenite size fraction, and in the top 10 cm changes into marly limestone similar to the marly limestones below the breccia bed and unconformity. The K-T boundary is placed at a thin clay layer, which contains high concentrations of iridium, between the top of the marly limestone and the overlying 5-cm-thick marly layer. This marl contains abundant planktonic foraminifera indicative of the earliest Tertiary zone P1a (*Parvularugoglobigerina eugubina*). A latest Maastrichtian age of the breccia as well as limestones below and above is indicated by the presence of *P. hantkeninoides*, which marks the last 170–200 k.y. of the Maastrichtian.

The presence of ammonites and neritic benthic foraminifera, the abundance of shallow surface-dwelling planktonic foraminifera (rugoglobigerinids, hedbergellids, guembelitrids, heterohelicids), and near absence of the deeper dwelling globotruncanids in the marly limestones indicate shallow-marine shelf deposition (<150 m depth) during the last 170–200 k.y. of the Maastrichtian. Different paleoenvironmental data appear 70 cm below the K-T boundary coincident with deposition of the limestone breccia which marks a sea-level lowstand followed by transgression. Benthic foraminifera suggest shallowing from middle neritic to inner neritic depths. At the same time, palynomorph taxa indicate a change from tropical to temperate climatic conditions (Ashraf and Stinnesbeck, 1988; Stinnesbeck and Keller, this volume). The sea-level transgression, which continued across the K-T boundary, reached middle neritic depths by K-T boundary time. The earliest Paleocene zone P0 has not been detected and may be missing at the Poty section, although an iridium anomaly is present in the thin clay layer (Albertao and others, 1994). The last 200 k.y. in the Poty Quarry section thus record a major global sea-level lowstand during the latest Maastrichtian, which was accompanied by deposition of a breccia layer, followed by a sea-level transgression and deposition of marly limestone and clay by K-T boundary time.

Albertao and others (1994) interpreted the Poty Quarry K-T boundary sequence as indicating two closely spaced bolide impacts: one at the K-T boundary which they placed at the base of the breccia, and the second in the Danian marked by the iridium anomaly. They based this interpretation on the putative identification of Danian species and microspherules, which they considered to be microtektites in the breccia and overlying marls and limestones. Examination of these sediments does not confirm their observations. Several of the species they considered to be of Danian age actually range into the Maastrichtian and others seem misidentified (for a discussion see Stinnesbeck and Keller, this volume). Moreover, the microtektite-like spherules that are abundantly present throughout the section and appear to be amber colored and glassy, are phosphatic and dissolve in strong acid. They are likely infillings of algal resting cysts which are commonly present in K-T transitions (Stinnesbeck and Keller, this volume).

Chile

Along the Pacific coast of central Chile, late Maastrichtian marine detrital sediments of the Quiriquina Formation are preserved in isolated basins. Late Maastrichtian fossils including ammonites [*Eubaculites carinatus*, *Hoploscaphites quiriquiniensis*, *D. (Diplomoceras) cylindraceum*, and *P. (Neophylloceras) surya*] are abundant (Stinnesbeck, 1986; this volume) and disappear between 5 to 10 m below the K-T boundary.

In the type area near Concepción, glauconitic sand and siltstones with calcareous sandstone concretions characterize the upper part of the late Maastrichtian sediment sequence and indicate deposition in a quiet offshore shelf setting. Bioturbation is intensive, and *Teichichnus* and *Zoophycos* are the prevailing trace fossils. Upsection, the abundance of detritus feeding bivalves suggests stillwater conditions and sediments rich in organic matter for the uppermost 4 to 5 m of the Quiriquina Formation. This suggests the latest Maastrichtian sea-level transgression and maximum flooding surface. Above this interval, the marine sequence is disconformably truncated by (prograding?) gravelly yellow sandstones of the Curanilahue Formation, reflecting brackish to fluviatile conditions.

The K-T boundary is placed at the unconformity between the bioturbated marine siltstones of the Quiriquina Formation and the fluviatile conglomeratic sandstones of the Curanilahue Formation The last unequivocally Maastrichtian index taxon is present 5 m below this lithological contact. Above this contact, only long-rangi-

the base of the Curanilahue Formation. The first palynoflora of unequivocal early Tertiary age (probably Paleocene) was found 18 m above the lithological contact that marks the K-T boundary (Frutos, 1984, personal commun.).

These lithological and faunal data indicate that, during the latest Maastrichtian sea-level regression, marine conditions gradually changed to shallower neritic and eventually brackish to fluviatile conditions in the earliest Tertiary. This is evident in the deposition of glauconite-rich sands near the top of the late Maastrichtian that are overlain by the fluviatile conglomeratic bed near the K-T boundary. There is no evidence of catastrophic deposition at or near the K-T boundary; only normal sedimentary processes associated with a sea-level regression that changes a marine environment to a nearshore depositional environment.

Antarctica: Seymour Island

The K-T boundary on Seymour Island is located in an expanded, shallow-marine clastic sequence consisting of loosely consolidated glauconitic silty sandstones of the Lopez de Bertodano Formation. Except for palynomorphs, the glauconitic interval is sparsely fossiliferous: planktonic foraminifera and calcareous nannoplankton are absent. Invertebrate faunas indicate a gradual faunal turnover throughout this interval, the stratigraphically highest ammonite occurrence (*Maorites densicostatus*) being near the base of the glauconite-rich interval (Zinsmeister and others, 1989; Zinsmeister and Feldmann, this volume). Based on a marked decrease of most invertebrate groups, Macellari (1986, 1988) originally placed the K-T boundary at the base of the glauconite-rich interval.

Dinoflagellate cysts provide the most precise biostratigraphic control for placing the K-T boundary in a 20–30 cm transitional interval just above a sharp iridium peak and ~3.5 m above the last ammonite occurrence (Askin and Jacobson, this volume; Elliot and others, 1994).

Both macrofossil and microfossil assemblages indicate that biotic and environmental changes on the Seymour shelf began well before the iridium layer was deposited at the end of the Cretaceous. Invertebrate species show a rapid turnover and declining ammonite diversity in the upper 40 m of Maastrichtian strata (*Pachydiscus ultima* and *Zelandites varuna* ammonite zones, Macellari, 1986, 1988; Zinsmeister and others, 1989; Zinsmeister and Feldmann, this volume). Dinoflagellates also show an accelerated turnover shortly before the end of the Maastrichtian, marked by an influx of new forms (Askin, 1988; Askin and Jacobson, this volume).

These biotic and lithological changes can be explained, at least in part, by the latest Maastrichtian sea-level regression, followed by transgression across the K-T boundary. In the Seymour Island sections, the glauconite-rich interval marks the sea-level regression during the latest Maastrichtian that temporarily reduced water depths from an offshore marine shelf to a marginal marine environment. At or before K-T boundary time, sea level began to rise, reaching a maximum flooding surface shortly after (30 cm above) the K-T boundary and iridium peak (Askin and Jacobson, this volume; Elliot and others, 1994). Thus, Seymour Island sections provide no evidence for unusual chaotic sediment deposition across the K-T boundary event, but rather show normal marine sedimentary

DISCUSSION AND CONCLUSIONS

We have examined K-T boundary sections from low to high latitudes, spanning depositional environments from continental shelf to shelf-slope and the deep sea (Figs. 1 and 3). Across these latitudes the overwhelming majority of deep-sea sections are marked by condensed sedimentation, nondeposition, or erosion across the K-T boundary transition (MacLeod and Keller, 1991a, 1991b; Keller, 1993; Keller and others, 1993a; Peryt and others, 1993) and, except for flysch deposition near continental margins, no clastic deposits are present. Outer shelf to upper slope environments, currently known from the Tethys region (Tunisia, Israel, Spain, and Mexico, Fig. 3), are characterized by generally higher rates of sedimentation, more continuous deposition and very short intervals of nondeposition or hiatuses (MacLeod and Keller, 1991a, 1991b; Keller and others, 1994a, 1994b). Clastic deposits are present only in northeastern and east-central Mexico sections and the southern United States, where they vary in nature and thickness, ranging from several meters to 10 cm and locally disappear (Keller and others, 1994a, 1994b). Middle shelf to inner neritic environments are known from many sections spanning low to high latitudes (Fig. 3). Sediment accumulation rates are generally high and sea-level fluctuations are marked by erosion, deposition of clastic deposits, and glauconite formation. Of the 12 shallow-water sections examined, only the Danish sections (Stevns Klint and Nye Kløv) lack clastic deposits, although they contain hiatuses (Schmitz and others, 1992; Keller and others, 1993b). Clastic deposits, including coarse sands, glauconite, and breccias, are thus common occurrences in continental shelf or platform settings worldwide at times of sea-level regression to transgression inflection points.

Are near K-T boundary clastic deposits the result of an impact-generated megatsunami wave? This sensational interpretation has received such popular support in some circles that a K-T boundary impact origin has been invoked for breccia deposits of unknown age in Belize (Ocampo and Pope, 1994), Guatemala (Hildebrand and others, 1994), and Chiapas (Montanari and others, 1994), and even for exfoliated sandstone boulders in Cuba (Bohor and Seitz, 1990), without any supporting evidence of impact origin or evidence of K-T boundary age. Clastic coarse sandstone deposits in Texas, Alabama, and Georgia have been reinterpreted as K-T impact-generated tsunami deposits regardless of their stratigraphic positions below or even above the K-T boundary, microkarstification or bioturbation indicating deposition over a longer time interval than a few days (Savrda, 1993), and the absence of any impact material (Bourgeois and others, 1988; Hildebrand and Boynton, 1990; Habib, 1994; Smit and others, 1994a, 1994b). Moreover, breccia, sandstone, and channel-fill deposits of near K-T boundary age in Mexico have been reinterpreted as Chicxulub impact-generated megatsunami deposits (Hildebrand and others, 1994; Smit and others, 1992, 1994a, 1994b; Montanari and others, 1994), although typical late Maastrichtian hemipelagic sediments top the clastic units and discrete layers of bioturbation are present within (e.g., Lajilla, Mulato, Parida and Bochil in Mexico, Poty Quarry in Brazil), indicating pre-K-T boundary deposition. Independent impact evidence is generally lacking (a small iridium anomaly found at Mimbral occurs within the

Parida, see Stinnesbeck and others, this volume) lack any clastic deposit showing normal sedimentation across the K-T boundary.

Of all the sections for which K-T impact-generated tsunami deposits have been claimed, sections in northeast and east-central Mexico and Haiti deserve special attention because of their unique spherule-rich deposits. In the Beloc section of Haiti, this spherule-rich deposit contains many glassy spherules that have been variously interpreted as of impact origin (Izett, 1990; Sigurdsson and others, 1991; Blum and Chamberlain, 1992; Koeberl and Sigurdsson, 1992; Keoberl, 1994) or of volcanic origin (Jéhanno and others, 1992; Lyons and Officer, 1992; Robin and others, 1994). The spherule-rich layer at the base of the clastic deposits at Mimbral and Lajilla contains very rare glass fragments similar in chemical composition to those from Haiti, suggesting a common origin. If these glass spherules and fragments are of impact origin, then impact-triggered earthquake or tsunami deposition of the overlying clastic sediments in the northeastern and east-central Mexico sections becomes probable. However, this bolide impact would not have been of K-T boundary age, but preceded the boundary event by some thousands of years. This is suggested by the 25–30 cm of late Maastrichtian sandy marls followed by the K-T boundary iridium anomaly, overlying the clastic deposit in the undisturbed Beloc sections (Jéhanno and others, 1992), and similarly by the 5–10-cm-thick late Maastrichtian marls above the clastic deposits in sections at Lajilla, Mulato, and Parida, and 100 cm at Bochil (Keller and others, 1994a; Montanari and others, 1994; Lopez-Oliva and Keller, 1995; Macias Pérez, 1988; Stinnesbeck and others, this volume). In these sections as well as others, the Ir anomaly, Ni-rich spinels, first appearance of Tertiary planktonic foraminifera, and the $\delta^{18}O$ shift that are used to mark the K-T boundary worldwide occur well above the clastic deposits.

We do not believe sufficient evidence is present to identify any of the known K-T or near K-T clastic deposits as of a single impact origin because of the following. (1) They are of variable ages and frequently predate or postdate the K-T boundary. (2) They contain no unequivocal impact ejecta (with the possible exception of glass spherules from Haiti and rare glass shards from some northeastern Mexico sections, the origin of which, whether volcanic or impact, is still in dispute; see Keoberl, 1994, and Robin and others, 1994). (3) They do not represent a single event deposit over a few days, but multiple events over a longer time period as indicated by disconformities, different mineralogical contents between strata, layers of hemipelagic sedimentation, microkarstification, and microlayering of fine- and coarse-grained sediments. (4) They are frequently burrowed and contain multiple horizons of resident ichnofauna truncated by erosion. This repeated recolonization indicates deposition over an extended time period, rather than rapid catastrophic accumulation. (5) A tsunami interpretation fails to account for the effects of sea-level changes upon sediment deposition across the K-T transition. (6) A tsunami interpretation ignores the fact that microfossils and/or macrofossils generally indicate a sea-level lowstand at the time of breccia or clastic deposition, which was followed by rising seas. (7) Normal hemipelagic sedimentation of Maastrichtian age above the clastic deposits invalidates a K-T boundary age for these deposits. Ignoring these factors and prematurely placing an impact-generatbasis of their geographic proximity to the proposed Chicxulub impact crater inevitably leads to spurious scenario-driven interpretations.

Our biostratigraphic, lithological, and depositional analyses of K-T boundary transitions for which impact-generated clastic deposits have been claimed are illustrated in Figure 7 along with the sea-level curve derived from high-resolution studies of numerous K-T boundary transitions worldwide (see earlier discussion). In all sections where near K-T boundary clastic deposits were identified, the underlying unconformity is determined to be of latest Maastrichtian age (*P. hantkeninoides* zone, chron 29R) and coeval with the sea-level regression, as indicated by benthic foraminifera, ostracodes, invertebrates, and dinoflagellates (Keller, 1988b, 1992; Moshkovitz and Habib, 1993; Olsson and Liu, 1993; Savrda, 1993; Stinnesbeck and Keller, this volume; Stinnesbeck and others, 1993; Keller and others, 1993b; Schmitz and others, 1992).

This sea-level regression resulted in erosion and nondeposition that varied depending on the depositional environment (increased erosion landward). In the shallowest Gulf Coast sections (Braggs, Mussell Creek, Moscow Landing), sediment deposition (Clayton Sand) did not resume until the later Danian sea-level transgression (Mancini and others, 1989; Savrda, 1993; Moshkovitz and Habib, 1993), whereas in the Millers Ferry section it resumed earlier (Liu and Olsson, 1992), possibly because of its somewhat deeper water environment. Brazos River sections differ in that the clastic deposit is of Maastrichtian age (*P. hantkeninoides* zone), followed by shale deposition, interrupted only by a short interval of nondeposition at the P0–P1a boundary and by a short hiatus and glauconite deposition at the P1a–P1b boundary (Keller, 1989a, 1989b) coincident with short-term sea-level lowstands. This suggests that Brazos sections may have been located in somewhat deeper waters than the Alabama sections.

In the deeper water (400 m) northeastern and east-central Mexico sections, sediment deposition is similar to that of the Brazos section. The basal spherule-rich unit of the clastic deposits rests unconformably upon marls of the latest Maastrichtian Mendez Formation (*P. hantkeninoides* zone), coincident with the sea-level lowstand (Fig. 7). The top of the clastic deposit is also marked by a disconformity followed by a thin layer of Maastrichtian marls ranging from 5–10 cm thick in northeastern Mexico sections (Lajilla, Mulato, Parida) to 100 cm thick in southern Mexico (Bochil, Chiapas) (Keller and others, 1994a, 1994b; Montanari and others, 1994; Macias Peréz, 1988). In most sections, the earliest Danian zone P0 and the lower part of zone P1a are missing, and a short hiatus is present at the P1a–P1b boundary, as also observed at the Brazos section. On the basis of available stratigraphic information, sediment deposition in the Haiti sections appears to be similar to that in northeastern Mexico (Fig. 7).

Stratigraphically, the platform breccia deposits of the Bochil section in Chiapas and the Poty section in Brazil are of late Maastrichtian (*P. hantkeninoides* zone) age. In these locations, the breccia beds are overlain by latest Maastrichtian and early Paleocene sediments. The Chilil section, which has also been claimed to have impact-related clastic deposits, in fact, contains no sediments of K-T boundary age because an erosion surface cuts from the upper Paleocene (zone P2) to the lower upper Maastrichtian *G. gansseri* zone.

impact-generated megatsunami deposits (Alvarez and others, 1992). However, DSDP Site 540 has no sediment deposition between the middle Maastrichtian (*G. aegyptiaca* zone) and the early Paleocene zone P2. Likewise, DSDP Site 540 has a hiatus that spans from the middle or lower upper Maastrichtian to the early Danian zone P1a, followed by a second short hiatus at the zone P1a–P1b interval (Keller and others, 1993a).

Thus, Figure 7 shows that, based on current biostratigraphic information, not all near K-T clastic deposits are coeval. Their stratigraphic positions vary systematically with paleodepth and proximity to shorelines. In upper slope to outer shelf environments, clastic deposition began with the onset of the latest Maastrichtian sea-level regression following the global sea-level lowstand and unconformity. In the stratigraphically more complete sections, normal hemipelagic sedimentation resumed prior to the K-T boundary, such as at Lajilla, Mulato, and Parida in northeastern Mexico, the Mirador section near La Ceiba in east-central Mexico, the Bochil section in Chiapas, southern Mexico, the Beloc section in Haiti, the Poty section in Brazil, and the Brazos section in Texas. In these localities, sediment deposition during the latest Maastrichtian (after deposition of the clastic members) and early Danian occurs in generally deepening environments. Deposition is generally interrupted by short hiatuses or condensed intervals in the early Danian, indicating short-term sea-level lowstands at the P0–P1a and P1a–P1b boundaries. Deposition of clastic deposits in shallow neritic environments such as Braggs, Mussel Creek, Moscow Landing, and Millers Ferry in Alabama seems directly related to water depth. In these shallow-water sections, a major erosional unconformity is associated with the latest Maastrichtian sea-level lowstand, and sediment deposition (in this case of clastic deposits) resumes only with the early Danian sea-level rise (Fig. 7).

Our investigation thus concludes, in agreement with many previous studies, that variations in clastic deposition of near K-T boundary age are associated directly with a sea-level lowstand (Donovan and others, 1988; Baum and Vail, 1988; Mancini and others, 1989; Keller, 1989; Savrda, 1991, 1993; Habib and others, 1992; Moshkovitz and Habib, 1993). At this time, we see no convincing evidence that any of these clastic deposits are generated by a megatsunami wave as a result of a K-T boundary bolide impact on Yucatan. We remain open to this interpretation for at least some deposits (e.g., glass at Mimbral and Haiti). An impact origin for this glass, however, would indicate a pre-K-T boundary bolide impact for which there is no biotic, stratigraphic, or geochemical evidence at this time. Likewise, a volcanic origin for this glass suggests that major volcanic activity preceded the K-T boundary. In either case, these catastrophic events would have occurred during the last 200 k.y. of the Maastrichtian, and possibly coincided with the latest Maastrichtian sea-level lowstand, which influenced the nature of sediment deposition.

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