25 March 1983, Volume 219, Number 4591 **SCIENCE**

The Cretaceous-Tertiary Transition

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Summary. The fossil sequences from cores across the Cretaceous-Tertiary boundary show a range of transition times and transition time intervals depending on the fossil indicators and the location of the site. These variations, together with the pattern of iridium distribution with depth at some sites, differences in total amounts of iridium, variations in noble metal abundances normalized to extraterrestrial concentrations, the depositional effects that might be expected in a reducing environment, and the clay mineralogy of the boundary layer clays, put into question the interpretation that an extraterrestrial event was the cause of the faunal changes and the iridium anomaly in the vicinity of the Cretaceous-Tertiary transition. It seems more likely that an explanation for the changes during the transition will come from continued examination of the great variety of terrestrial events that took place at that time, including extensive volcanism, major regression of the sea from the land, geochemical changes, and paleoclimatic and paleoceanographic changes.

A nonstratigrapher may well assume that period or epoch boundaries are well defined isochrons just as a stratigrapher might assume that the geochemists really know the bulk composition of the earth or that the seismologists know the makeup of the core. Stratigraphers, who have to wrestle with the problem, know the pitfalls involved in the establishment of a time stratigraphic boundary or in ensuring that it represents the same time at different places (*1*, *2*).

The magnetic polarity time scale is proving most helpful in determinations of whether rocks or boundaries that are geographically widely separated may be of the same age. One can quickly decide whether two samples can have the same age because the magnetic polarity preserved in each would be the same. Although the time scale of magnetic reversals is no better in absolute time than the faunal, radiometric, or other clocks against which it is calibrated, if a particular reversal event can be identified in two sections, then we know that the ages of the reversal intervals are simultaneous to within approximately 10^4 years.

This assumes that the sediments have not been disturbed after deposition; however, bioturbation is common in marine sediments. Instead of a sharp boundary between two epochs, we find a transition zone in which the fauna change from those of an earlier epoch to those of a later one. One would not expect, however, for bioturbation to result in the separation of initially simultaneous deposits; this was suggested by Ganapathy (*3*) to explain the 30-centimeter difference in a core of the occurrence of a concentration of microtektites and an anomalously large amount of iridium near the end of the Eocene.

The boundary problem has come into sharp focus in recent years, stimulated by the suggestion that the Cretaceous-Tertiary (K-T) boundary is marked by the products of impact of a large extraterrestrial object on the earth. The primary indicator cited is an anomously large amount of iridium in a narrow zone in the sedimentary rocks that cross the boundary (*4*); its presence is attributed to the impacting body. There may well be sources or concentration mechanisms other than a single large impact. Goldschmidt (*5*), for example, noted: "There seemed to be quite a distinct concentration of platinum metals in the red deep-sea clay, but unfortunately volcanic dust from the large explosion of Krakatoa also contained such metals. Therefore at present it cannot be decided whether an increase in platinum metals in the mineral matter of deep-sea sediments can be accounted for by meteoric dust only, or whether world-wide falls of andesite dust can be responsible for the increase. . . . In the choice of sediments one should avoid basins with anaerobic bottom water, to avoid cases of precipitation of solute precious metals from sea water."

It is a pity that the suggested object chose this particular time to collide with the earth because this was, geologically, very noisy. In many areas the K-T boundary is marked by an unconformity related to the last major regression of the sea from the continents (*6*). The Late Cretaceous was a time of enormous volcanic activity all around the Pacific (*7*); most of the Mancos-Pierre shale sequence, some 6 million cubic kilometers in the western United States, is of volcanic origin (*8*). The marine geologists tell us that there were changes in the carbonate compensation depth in the oceans during this time, sometimes up (*9*), sometimes down (*10*), as well as changes in the isotopic composition of significant stable elements (*11*, *12*). There were major extinctions of fauna in the vicinity of this boundary, ranging from corals and planktonic organisms to dinosaurs. In the Tertiary, the mammals became dominant. The Laramide orogeny, which created the spectacular mountains in the western United States, culminated about this time.

The methods of timing these various events are several and the precision of the timing is at best on the order of thousands of years and probably not nearly that good. If indeed the iridium

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anomaly, from whatever cause, marks an isochron of worldwide extent, it would be an exceedingly valuable marker against which we could calibrate other dating methods. Toward this end, it is useful to examine in detail the reported stratigraphic sections that appear to be continuous from Cretaceous to Tertiary, as has been done, for example, in Gubbio, Italy (*13–17*). Can we relate events that happen during the lifetime of organisms, say 10^0 to 10^2 years, to events that are recorded in the transition zone in measurable geologic time, say 10^4 to 10^6 years?

As shown in Fig. 1, our concern is with the time (T_n) of transition of a faunal sequence from one epoch to another and the time interval (δt_n) over which this transition takes place. We would like to know when it occurred and how long it took, as measured against a magnetic reversal sequence.

From three investigations (*18–20*) of the time intervals of normal (N) and reversed (R) polarity sequences in the vicinity of the K-T transition (Table 1), we adopted the scale of LaBrecque *et al.* (*19*). To examine the K-T boundary, we arbitrarily chose a reference time, T_0, at the end of interval 29N. Determinations of transition time T_n will be with reference to this time.

For an instantaneous worldwide extinction event the value of the T_n from various geological sections and determined by various faunal sequences should be the same and δt_n should be zero. If there are measurable time differences in T_n from place to place, the transitions cannot represent synchronous events. If δt_n is finite, the particular faunal assemblage or assemblages by which it is measured have changed gradually in response to changes in the environment.

With the exception of an instantaneous event, the T_n values may not necessarily be the same for given paleontological indices defining the K-T transition nor for the same index from area to area. Even for microfossils, which often show an abrupt transition as measured in geologic time, the T_n values may be different from one ocean environment to another. The δt_n values, which may differ among species and among regions, depend on the magnitude of the environmental change and its effect on a given category of organism. We have determined and examined T_n and δt_n values from available geological sections that appear to be continuous across the K-T transition to see how they conform with various hypotheses that have been called upon to explain the faunal transition.

The magnetic reversal stratigraphy provides a time scale independent of index fossil time comparisons in the sense of providing equivalent time markers. Thus, comparisons of T_n with reversal boundaries provide an indication of synchroneity although they do not give absolute time. To determine δt_n, a reasonable rate of sediment accumulation is the principal requirement, and this can be acquired from magnetic, paleontological, or lithological data.

Deep Sea Drilling Project Sites with K-T Samples

Among sites of the Deep Sea Drilling Project (DSDP) where samples from the K-T transition were taken are six holes of particular interest (A through F, Fig. 2), which are discussed in the following paragraphs.

DSDP site 384. In July 1975, this site was considered "the most complete biostratigraphic record of the Cretaceous-Tertiary transition obtained in a pelagic facies" (*21*). It is located in the western North Atlantic (40°22′N, 51°40′W), and from empirical age-depth curves the depth at this site at the time of the K-T transition is estimated to have been 3 to 4 km (*10*). The calcite compensation depth (CCD) is thought to have descended during the Late Cretaceous and early Tertiary (*10*) because of the continuous sequence of calcareous oozes preserved at the site. From the range of index fossils, the average sediment accumulation rate over the K-T transition was estimated to be 0.6 centimeter per 1000 years (ky) (*21*), a change from 0.3 cm ky^{-1} in the Maestrichtian and 1.0 cm ky^{-1} in the Danian [figure 14 in (*21*)].

Thierstein and Okada (*22*) have described in detail the nannofossil distributions and extents across the K-T transition. The plot of summed relative abundance of Cretaceous and Tertiary taxa to total taxa across the transition (Fig. 3) shows that the change from dominantly Cretaceous to dominantly Tertiary taxa occurs over a depth interval of about 60 cm which, combined with sedimentation rate, yields a δt_n value of 100 ky. This is an integrated δt_n value for the entire assemblage of nannofossils. Transition times for individual species are reported by Thierstein and Okada [table 1 and figure 7 in (*22*)]. Some of the individual species disappear between the sampling interval of 12 cm, $\delta t_n < 20$ ky; others persist for 50 ky.

Thierstein and Okada (*22*) interpret this faunal sequence in terms of a sudden extinction of all Cretaceous taxa and simultaneous appearance of Tertiary taxa with an extended bioturbation, or

Table 1. Magnetic time scale intervals. N, normal; R, reversed; T_0, time at end of normal interval 29N; my, million years.

Source	Time interval (ky)							T_0 (my)
	28N	28R	29N	29R	30N	30R	31N	
Tarling (*18*)	1260	560	1020	830	2030	170	1340	64.03
LaBrecque (*19*)	690	310	560	470	1390	80	730	64.90
Ness (*20*)	960	430	790	610	1820	110	950	66.56

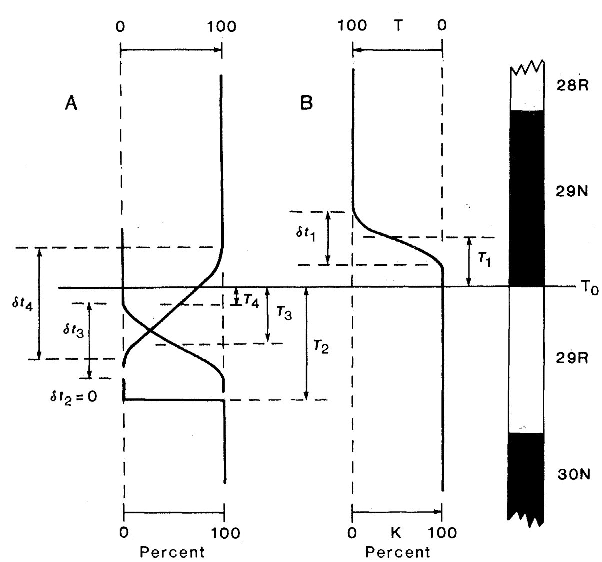


Fig. 1. Definitions of K-T faunal transition times (T_n) and transition time intervals (δt_n) with respect to paleomagnetic time scale for (A) integrated species community and (B) individual species members.

mixing, depth interval of 26 cm through the entire transition sequence. Bioturbation is an important process affecting marine sedimentary sequences and can blur or obscure transition events. Analytic solutions and computer programs have been developed for bioturbation effects for steady state (*23*), instantaneous (*24*), and time dependent (*25*) flux inputs. The solution of Thierstein and Okada, although analytically sound, is not the only one possible. In other words, when the mixing depth is not known, a variety of δt_n values can be ascertained. It is necessary, then, to look elsewhere for determination of mixing depths.

Depths to which bioturbation has been observed in modern deep ocean sediments in the Atlantic, Indian, and Pacific oceans (*26*) are, respectively, 9, 9, and 8 cm. To compare these depths with that from the site 384 core, effects of compaction must be taken into account. The thickness of compacted sediments (h') is given by $h' = (1 - \phi_0)h_0/(1 - \phi')$, where h_0 is the original thickness and ϕ_0 and ϕ' are the original and compacted porosities. For site 384 at the K-T transition, $\phi' = 0.60$ (*27*). Values of ϕ_0 may be in the range of 0.80 to 0.85 so that $h' = (3/8)h_0$ to $(1/2)h_0$. Thus, modern sediments 8 to 9 cm thick would, if equally compacted, end up 3 to 4 cm thick, and the 26-cm interval in the core from site 384 would have an uncompacted thickness of about 61 cm.

In the description of the core from the K-T transition at site 384, an abrupt color change is reported at a depth of 167.93 meters, just at the beginning of the transition (*21*). Further, there is a layer rich in foraminifera within the transition zone at 167.87 m in which the relative abundances are 25 to 49 percent in comparison with 5 to 24 percent at depths of 167.77 and 167.92 m on either side of the peak (*21*). Preservation of the color change and the foraminiferal layer would be unlikely if bioturbation extended over the entire 26-cm interval; it seems more likely that the effects of bioturbation in the compacted sediments have been limited to a few centimeters and that the integrated value for δt_n is around 100 ky.

Larson and Opdyke (*28*) determined the magnetic stratigraphy in the site 384 core for the depth interval from 156 to 176 m. Although the quality of the data is not the best, they clearly indicate that the nannofossil K-T transition lies in a normal polarity sequence which the authors identify as 29N. This sequence occurs over a 150-cm interval in the core and is defined by 13 normal polarity measurements. If T_n is taken as the midpoint in the nannofossil transition, this corresponds to a depth in the core of 167.60 m, a paleomagnetic age of 29.9N, and a T_n of -60 ky. Sedimentation rates calculated from the magnetic stratigraphic correlations of Larson and Opdyke are 0.7, 0.4, and 1.0 cm ky^{-1} for the Tertiary intervals 28N, 28R, and 29N and 0.2 and 0.2 cm ky^{-1} for the Cretaceous intervals 29R and 30N with an average rate for the entire sequence of 0.4 cm ky^{-1}, rates comparable to those given above on the faunal ranges.

The occurrence of the nannofossil K-T transition in a normally magnetized sequence is at odds with the well-studied section at Gubbio, Italy, where the microfossil transition occurs in an earlier sequence of reversed polarity identified as 29R (*16*). Larson and Opdyke (*28*) suggest that the position near a field reversal, together with the changes in lithology and paleontology noted by sci-

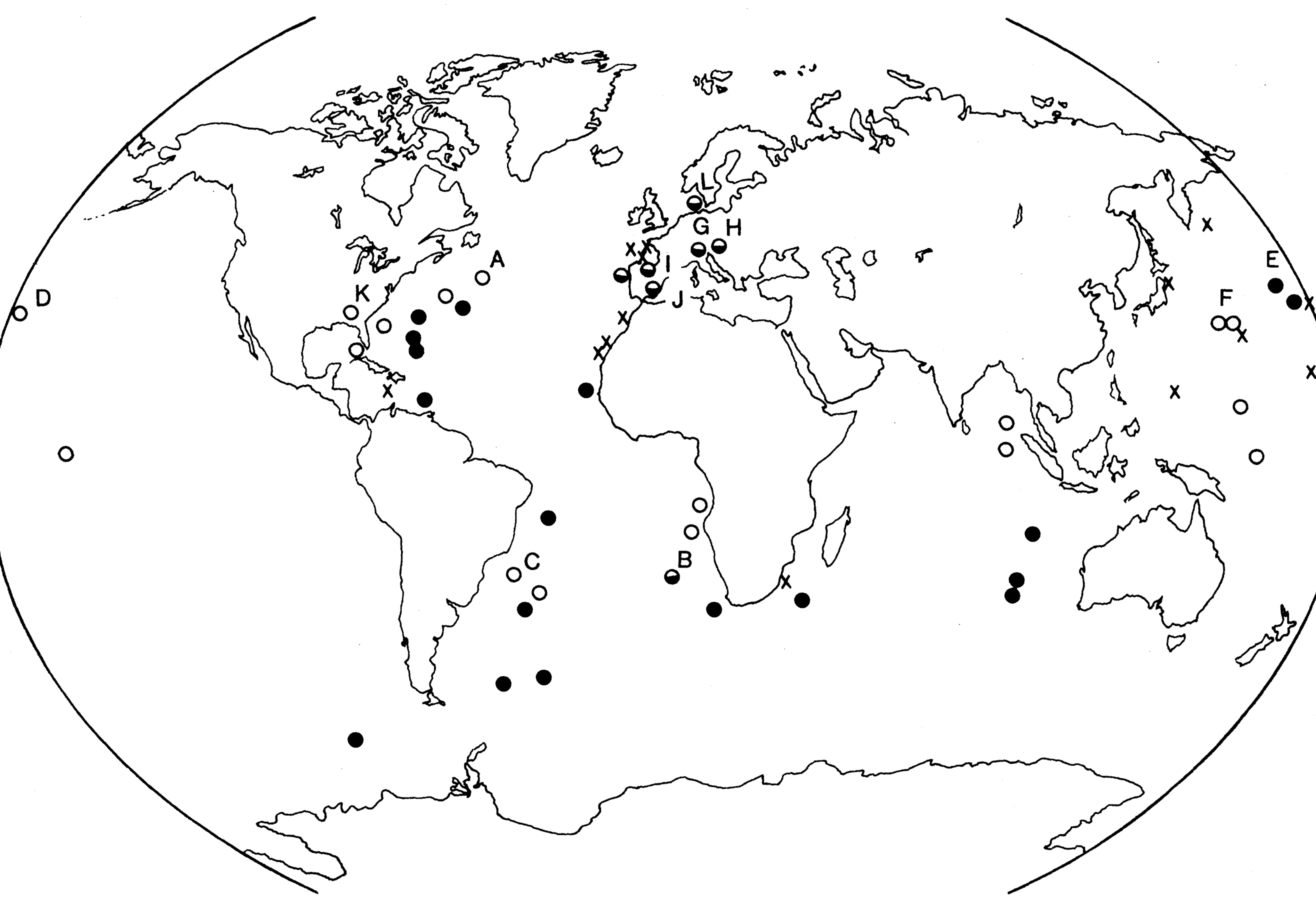


Fig. 2. Lithology of marine sequences near the K-T transition: continuous calcareous sequence (○); continuous pelagic clay sequence (●); calcareous, boundary layer clay, calcareous sequence ⊖; and extended hiatus (X). Locations discussed in text are indicated by *A*, DSDP site 384; *B*, DSDP site 524; *C*, DSDP site 356; *D*, DSDP site 465; *E*, DSDP site 464; *F*, DSDP site 47; *G*, Gubbio, Italy; *H*, Belluno, Italy; *I*, Zumaya, Spain; *J*, Caravaca, Spain; *K*, Braggs, Alabama; and *L*, Stevns Klint, Denmark.

entists on the DSDP ship, might indicate a hiatus at the biostratigraphic boundary. Thierstein and Okada (*22*) have a different interpretation; they do not see a hiatus, but discount the significance of the normal polarity event, placing the K-T transition in sequence 29R. The apparent continuous depositional character across the boundary shown by the core, the well-defined normal polarity event, and the reasonable sedimentation rates determined from the magnetic stratigraphy favor the original interpretation by Larson and Opdyke—that the K-T transition at site 384 occurs in a normal polarity event.

DSDP site 524. At site 524, which is located in the eastern South Atlantic at 29°29′S, 3°31′E (*11*), sedimentation rates at the time of the transition were high, around 2.8 cm ky^{-1}, reflecting inputs of volcanic origin from the nearby Walvis Ridge. The sediments are described as pelagic with turbidites and contain a boundary layer clay about 3 cm thick. The authors (*11*) postulate from the observed $CaCO_3$ distribution that the calcite compensation depth rapidly shallowed during the K-T transition; they do not state how shallow it became nor do they give a paleodepth for the site. Because of continuous calcareous K-T sequences at other DSDP sites, Thierstein (*29*) considers this interpretation as a worldwide event to be untenable.

The transition from dominantly Cretaceous to dominantly Tertiary nannofossils occurs over an interval of 230 cm, a longer transition zone than that at site 384 but commensurate with the higher sedimentation rate. The $\delta^{13}C$ values (the per mil deviations from the PDB isotopic standard) for the Cretaceous and Tertiary forms in the transition zone are about 1 unit less than those for the underlying Cretaceous bulk samples indicating that mixing was not significant and that the Cretaceous and Tertiary taxa lived together in the same $\delta^{13}C$ environment. The calculated δt_n value is 80 ky. From the midpoint of the K-T transition, T_n occurs at a magnetic stratigraphic age of 29.2R, or $T_n = +90$ ky, in close agreement with the results from Gubbio.

An iridium anomaly was reported with a peak amplitude of 3 parts per billion in the boundary clay, considerably smaller than that observed at other locations. It is spread over a depth interval of 20 to 100 cm rather than a spike, as would be expected from an extraterrestrial source. The measured iridium values show an inverse correlation with $CaCO_3$, or presumably a direct correlation with clay content, much the same as the results at

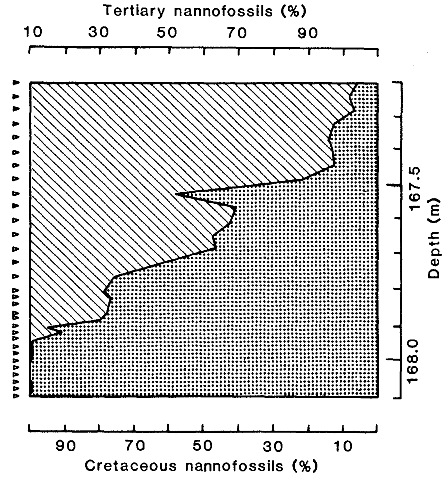


Fig. 3. Relative abundance of Cretaceous and Tertiary nannofossils versus core depth at DSDP site 384. [From Thierstein and Okada (*22*)]

site 465, which is discussed below. The iridium peak is not synchronous with the nannofossil transition but occurs at a depth of about 40 cm below the pronounced decrease in Cretaceous taxa.

DSDP site 356. Site 356 is located in the western South Atlantic on the São Paulo Plateau at 28°17′S, 41°05′W, where the water depth is 3175 m. Water depth during the late Cretaceous is estimated to have been about 1 km (*30*). The cores show a continuous sequence of slowly deposited calcareous oozes across the K-T transition, as might be expected if the calcite compensation depth were at a depth greater than 1 km rather than in the photic zone as suggested by Worsley (*31*).

The nannofossils show the same transition from Cretaceous to Tertiary as those at sites 384 and 524, with a $\delta t_n \geq 100$ ky (*29, 32*). No magnetic stratigraphy was reported for this site.

DSDP site 465. At site 465, in the northwestern Pacific at 33°49′N, 178°55′W and a water depth of 2161 m, an apparently complete section of calcareous oozes was recovered for the entire K-T transition (*33*). The sedimentation rates are estimated to be 1.8 cm ky^{-1} in the upper Maestrichtian and 0.4 cm ky^{-1} in the lower Paleocene. The foraminiferal assemblage places the transition at about 144 cm in core 3-3, and the transition is marked by an abrupt color change from white in the Maestrichtian to gray in the Danian (*33, 34*). Although the transition is apparently quite sharp, it is somewhat indeterminant in this core because of drilling disturbances and dissolution of the Cretaceous foraminifera; no estimate for δt_n is made. The nannofossil transition appears to occur later in the core at about 124 cm, which gives a ΔT_n value of 50 ky between the two transitions; the nannofossil δt_n value is around 10 ky.

Iridium determinations were also made at this site (*35*). The observed iridium peak is offset upward in the core by some 30 cm from the faunal K-T transition zone. From the estimated sedimentation rate of 0.3 cm ky^{-1} (*33*), this offset appears to be about 100 ky. The iridium is spread over an extended depth interval, in this case 60 cm, representing 150 ky; the distribution is not what would be expected for a spike input of iridium and normal bioturbation effects.

Iridium concentration is correlated directly with clay content and inversely with carbonate content, as shown in Fig. 4, with the exception of one observation. As mentioned, the same inverse Ir-$CaCO_3$ relation applies to the corresponding measurements at site 524. If the clay is of volcanic origin, Goldschmidt's comment (*5*) about a possible contribution of platinum metals to deep sea sediments from an explosion such as Krakatoa may be pertinent. Goldschmidt's second comment (*5*) about the enrichment of platinum metals under reducing conditions may also be pertinent. One cannot fail to note the presence of pyrite at the K-T transition in the core at site 465 (*36*). In the iridium analysis, the pyrite fraction at 119 and 143 cm in the core had iridium concentrations of 10.2 and 14.5 ppb, whereas the concentrations for the separated nonpyrite fractions at the same depths were 5.8 and 3.3 ppb (*35*).

DSDP site 464. Site 464, at 39°52′N, 173°53′E, is located in a water depth of 4637 m and has a monotonous sequence of brown clays extending from the Upper Cretaceous to the Miocene (*37*); this contrasts with the calcareous oozes at site 465. The ages and age ranges were determined from a study of ichthyoliths in the clays (*38*). The average sedimentation rate for this time span was 0.07 cm ky^{-1} (*37*). In contrast to site 465, the paleodepth at site 464 was below the calcite compensation depth. No magnetic stratigraphy was reported for either of these sites.

DSDP site 47. Site 47 is also located in the northwestern Pacific at 32°37′N, 157°43′E, in 2689 m of water. Here a sequence of calcareous oozes was sampled across the K-T transition (*39*), but the core appears to have been artificially disturbed. The average sedimentation rate in the vicinity of the transition is estimated to be 0.6 cm ky^{-1} (*40*). Both the nanno- and microfossil assemblages show a mixed Cretaceous and Paleocene constituency over an interval of 450 cm in this core (*40*), which, if real, would imply a δt_n of 750 ky. These two mixed zones are, in turn, offset from each other by about 600 cm, with the nannofossil zone occurring later than the microfossil zone for a ΔT_n difference of 1000 ky. It is possible that this offset is related to the drilling disturbance and subsequent core sampling, but it is difficult to understand how such a separation could occur if the two were originally synchronous in the core.

These six DSDP sites were selected for discussion because (i) the cores extended into the Cretaceous, (ii) core recovery across the K-T transition was good, (iii) the depositional sequence across the K-T transition appeared to be continuous, and (iv) there were detailed descriptions of some combination of the lithology, sedimentation rates, biostratigraphy, magnetic stratigraphy, and iridium content. There are many other DSDP sites that meet the first condition and some fewer that also meet the second condition. These latter sites have been summarized in Fig. 2, and the general lithology of the sites across the K-T transition has been divided into the four groups—a continuous calcareous sequence, a continuous pelagic clay sequence, a calcareous sequence with a boundary layer clay, and an extended hiatus. The extended hiatus, in all cases, is related to regional structural features. The normal sequence is the continuous calcareous ooze or pelagic clay, with two exceptions—site 524 and a site off Portugal. It would appear from this compilation that the K-T transition at many of the sites that have been examined in greatest detail (Gubbio, Belluno, Zumaya, Caravaca, Stevns Klint, and site 524) may represent regional rather than worldwide factors.

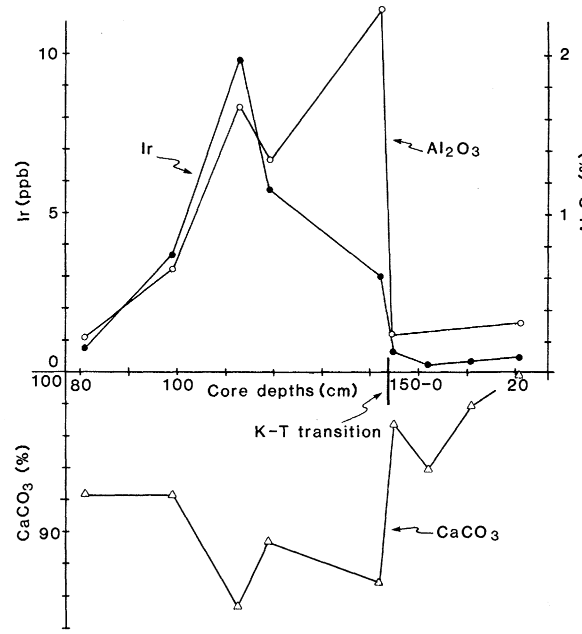


Fig. 4. Distribution of iridium, alumina, and calcite around the K-T transition at DSDP site 465.

Subaerial Outcrops of Marine K-T Transitions

Gubbio, Italy. The K-T section at Gubbio consists of a Cretaceous limestone, boundary layer shale about 1 cm thick, and Tertiary marlstone (*4, 13*). The average sedimentation rates have been estimated to be 0.8 to 1.3 cm ky^{-1} for the calcareous Upper Maestrichtian and 0.2 to 0.5 cm ky^{-1} for the shalier Paleocene (*13*). From the foraminiferal sequence (*14*), the K-T transition has been located in the boundary layer shale with a magnetic stratigraphy age of 29.3R, or $T_n = +140$ ky (*16*); and δt_n for the change in the individual microfossil taxa studied has been estimated to be < 10 ky (*41*). Evidence from burrows in the limestones suggest that bioturbation extended to a depth of about 5 cm in the compacted sediments (*13*).

The iridium content shows a distinct spike in the shale layer (*4*) with some extension into the overlying limestone, which can be attributed to bioturbation with a mixed layer thickness of 7 cm. The shale layer, itself, was deposited under reducing conditions (*4, 13, 42*).

Belluno, Italy. Additional magnetic stratigraphy age determinations were made for the foraminiferal K-T transition at four sections in the nearby Belluno Basin. The results were similar to those at Gubbio with a transition age of 29.3R (*43*).

Zumaya, Spain. The K-T section at Zumaya is similar to that at Gubbio; it consists of Cretaceous marlstone, boundary layer shale, and Tertiary limestone (*42*). The boundary shale, however, is 25 and 38 cm thick at the two sites investigated whereas it is about 1 cm at Gubbio. The shale is described as highly pyritic.

Both nanno- and microfossil transitions are described for the Zumaya sections. The sedimentation rate for the upper portion of the marlstone is estimated to be 4.0 cm ky^{-1} and that for the overlying limestone, 0.7 cm ky^{-1}; no sedimentation rate is given for the shale. The vanishing nannoplankton species disappear gradually from the last few centimeters in the marlstone through the lower half of the boundary shale; the new species become dominant in the middle of the shale. If a 0.7 cm ky^{-1} sedimentation rate is used as an estimate for the upper limit for shale deposition, then $\delta t_n \geq 30$ ky. The Cretaceous planktonic foraminifera, which show a different sequential development, gradually disappear in the marlstone some 20 ky before the nannofossil transition. The foraminiferal transition occurs across a depth range of about 120 cm, giving a δt_n value of 30 ky. The sequences suggest that the transition is not to be sought in some transitory event but in a more subtle and long-lasting modification of the biosphere (*42*).

Caravaca, Spain. The K-T section of Caravaca also resembles that at Gubbio with a 10-cm intermediate layer of clayey marl (*44*). The average sedimentation rates for the Maestrichtian and Paleocene are estimated to be 3.9 and 1.2 cm ky^{-1}, respectively; the time interval for the intermediate layer is estimated to be 6 to 25 ky. The dominant Cretaceous foraminifera disappear abruptly at the base of the intermediate layer, with an estimated $\delta t_n \leq 0.2$ ky. Some resistant Cretaceous taxa continue throughout the layer, and the Tertiary foraminifera begin abruptly at the top of the layer for an aggregate δt_n of between 6 and 25 ky.

There is an iridium spike in the basal portion of the clayey marl with a peak concentration of 25 ppb as compared with 6 ppb at Gubbio (*45*). Other trace elements and metals including osmium, arsenic, chromium, cobalt, nickel, selenium, and antimony also show enchanced concentrations.

Braggs, Alabama. The K-T section at Braggs is of shallow water origin and consists of a glauconitic calcareous sandstone overlain by a calcareous clay (*31*). The nannofossil transition is similar to that at DSDP sites 384 and 356 (*29*) and is estimated to have an interval time of $\delta t_n \geq 100$ ky. No magnetic stratigraphy was reported for the Zumaya, Caravaca, or Braggs sections.

Macrofossil K-T Transitions

The K-T macrofossil transitions have been studied for a longer period of time and, in general, are better known than the corresponding nanno- and microfossil transitions. Many of the macrofossil groups show a development that would place the K-T transition at or near the top of the Danian (*46*), reflecting the fact that the interval represented by the Danian was at one time considered to be uppermost Cretaceous. The length of the Danian is about 3400 ky, and it extends from magnetic anomaly 26R into anomaly 29R (*19*).

Denmark. In the well-studied sections in Denmark, echinoids (*47*), bryozoans (*48*), and echinoderms (*49*) show relatively little change during the K-T transition, particularly in comparison with corresponding faunal changes later in the Tertiary. The Danian taxa have as strong a biological affinity with the underlying Maestrichtian as with the overlying Tertiary. Although the bivalves in the Denmark sections show a change associated with the K-T transition, in terms of evolutionary perspective, all the new taxa were already present as important members of a complex fauna before the faunal break at the K-T transition (*50*).

The ammonites of Western Europe show a continuous decline throughout the Cretaceous (*51*). The marked changes in ammonoid character are related to marine regressions, and their disappearance is related to the major K-T regression.

At Stevns Klint, Denmark, the K-T transition is also characterized by a boundary layer clay similar to those at the Italian and Spanish sections. The clay, which is a few centimeters thick in the region sampled, has an iridium spike of 42 ppb (*4*). This iridium anomaly has been confirmed through measurements by others (*52, 53*). The clay is pyritic (*4, 31*).

Western North America. The dinosaurs are the most spectacular of the fauna to become extinct near the K-T transition but are, perhaps, the least useful in providing reliable T_n and δt_n values because of the small statistical sample available and because they occur in terrestrial sequences in which the magnetic stratigraphic interpretations are complicated by hiatuses. Butler and Lindsay *et al.* (*54*) describe the terrestrial sequence for three sections in the San Juan Basin of New Mexico. The magnetic stratigraphy correlates well with the sea-floor anomaly pattern, particularly that for long interval 30N, the short interval 30R, and long interval 31N (Table 1). An ash layer near the top of interval 30N has a potassium-argon age 64.6 ± 3.0 million years. The K-T transition, as defined by the highest stratigraphic occurrence of dinosaurs, is roughly at the midpoint of a well-defined normal polarity zone that has been correlated with interval 29N, or $T_n \sim -300$ ky. Some investigators have questioned the presumed continuous nature of these sections (*55*).

Lerbeckmo *et al.* (*56*) describe the terrestrial sequence at Red Deer Valley in Alberta, Canada. The palynofloral K-T transition occurs over an interval of 200 cm, centered 150 cm above a 20-cm bentonite layer with a potassium-argon age of 63 ± 2 million years. The last occurrence of *Triceratops* occurs 600 cm below the palynofloral transition and those of *Torosaurus* and *Tyrannosaurus* occur somewhat lower. The average sedimentation rate is estimated to be 6.5 cm ky^{-1}. There is considerable scatter in the magnetic stratigraphy data, presumably because of later overprints; but Lerbeckmo *et al.* suggest that both the palynofloral transition and the last dinosaurs occur in a reversed magnetic sequence that they correlate with 29R, or $T_n \sim +100$ ky. For the palynofloral transition, $\delta t_n = 30$ ky. The difference in T_n values between the last occurrence of dinosaurs and the change in palynoflora is 90 ky.

Archibald *et al.* and Hickey (*57*) describe a similar investigation in northeastern Montana. The palynofloral change and the occurrence of dinosaurs are in the same approximate stratigraphic interval as measured over a few meters but with a distinct nonsynchroneity of plant and dinosaur extinctions, and both occur in a reversed magnetic polarity zone. The magnetic stratigraphy observations cover only a short interval of geologic time, but on the basis of faunal correlations with the San Juan Basin, we infer that the reversed interval is 28R, or $T_n \sim -700$ ky.

Table 2. Summary of Cretaceous-Tertiary transition times (T_n) and transition times intervals (t_n).

Location	Polarity zone	T_n (ky)	δt_n (ky)	Index fossils
DSDP 384	29.9N	−60	100	Nannofossils, aggregate
			< 20 to 50	Nannofossils, individual
DSDP 524	29.2R	+90	80	Nannofossils, aggregate
DSDP 356			≥ 100	Nannofossils, aggregate
DSDP 465			10	Nannofossils, individual
DSDP 47			< 750	Nannofossils, aggregate
			< 750	Microfossils, aggregate
Gubbio	29.3R	+140	< 10	Microfossils, individual
Belluno	29.3R	+140		Microfossils, individual
Zumaya			30	Nannofossils, aggregate
			≥ 30	Microfossils, aggregate
Caravaca			6 to 25	Microfossils, aggregate
			≤ 0.2	Microfossils, individual
Braggs			≥ 100	Nannofossils, aggregate
Denmark			3000	Macrofossils
San Juan Basin	29N	−300		Dinosaurs
Red Deer Valley	29R	+100		Dinosaurs
			30	Palynoflora
Montana	28R	−700		Dinosaurs

Discussion

Faunal transitions. A summary of the K-T transition times and transition time intervals (Table 2) shows that the T_n values for Gubbio, Belluno, and DSDP site 524 all occur in the early portion of reversed magnetic polarity zone 29R but that for DSDP site 384 in the western North Atlantic is different, occurring in the latter portion of the normal polarity zone 29N. Although the times for the last occurrence of dinosaurs are subject to some question because of the nature of the terrestrial sections, the definitive results in the San Juan Basin indicate an extinction time about 400 ky later than the Gubbio microfossil transition time.

In general, the transition time intervals, δt_n, for the nanno- and microfossil transitions have measurable values in the range of 10 to 100 ky. The macrofossil data from Denmark would indicate little to no change across the K-T transition.

If the animals had been affected by environmental influences at K-T time, their transition times are measured in a few million years.

The results from Zumaya show that the difference in transition times between the micro- and the nannofossils is 20 ky and similar, although less definitive results from DSDP sites 465 and 47 show an offset in the same direction. This information would appear to raise the question of whether the nanno- and microfossil transitions are necessarily synchronous or whether the microfossils disappear first.

All the results indicate that if there has been an extraterrestrial event, the faunal transitions have not occurred instantaneously in response to it but rather have occurred at different times and over different time intervals. There appears to be a spectrum of T_n and δt_n values for the K-T transition and these depend both on the different fossils and the site location. Perhaps, as Percival and Fischer (*42*), Schopf (*58*), and Archibald and Clemens (*59*) suggest, the transition is not to be sought in some transitory event but in a more long lasting modification of the environment.

Iridium. A summary of iridium measurements in the vicinity of the K-T transition are presented (Table 3) along with measurements from a coal seam located at a K-T palynofloral transition in the Raton Basin, New Mexico. If the iridium concentrations are directly related to an extraterrestrial source with an instantaneous input, or spike, as measured in real time, then distributions in the sediment column after bioturbation at most sites should resemble that found at Gubbio. The measurements at DSDP site 524 and the more detailed measurements at DSDP site 465 do not follow such a pattern, but rather the distribution extends over about 60 cm. In addition, the iridium peaks at both DSDP sites occur at depths different from the fossil transitions, in one case before and in the other after the transition.

Peak iridium concentrations cannot be compared among sites because of probable differences in effects of bioturbation, but the total flux can be calculated for each site. The flux is simply the integral with respect to depth of concentrations of iridium above background (see Table 3). If the source of the iridium in the sediments is related to a worldwide atmospheric fallout, it might be anticipated as a first approximation that the fluxes would be about the same from one location to the next. Although the fluxes for Gubbio, Stevns Klint, and DSDP site 524 are similar, those for Caravaca and

Table 3. Summary of iridium measurements in the vicinity of the Cretaceous-Tertiary transition. An iridium anomaly in a marine K-T section in New Zealand has also been reported (*4*). An iridium anomaly was originally reported for a terrestrial section in Montana (*66*) and subsequently denied (*67*); more recent observations (*60*) at the same locality show levels of 0.07 to 0.17 ppb.

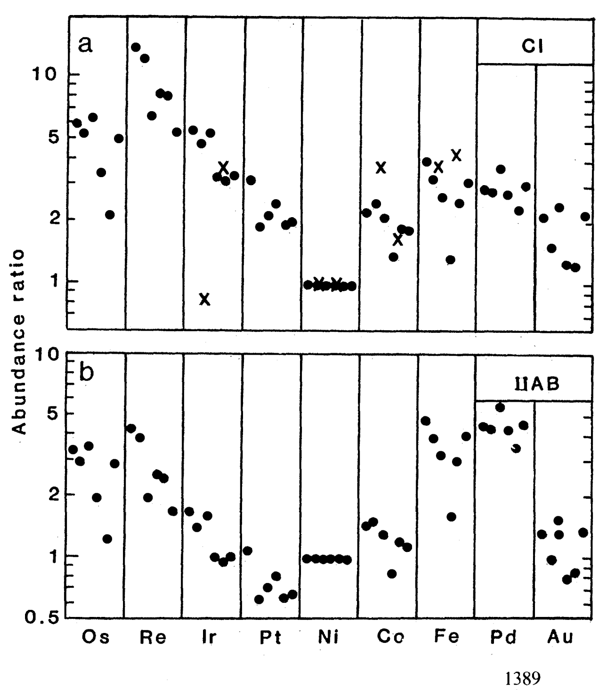
Location	Peak concentration (ppb)	Flux (ng/cm^2)	Distribution	Lithology	Comments
Gubbio (*4*)	6	90	Spike	Clay (1 cm)	Reducing
Stevns Klint (*53*)	41	90	Spike	Clay (3 cm)	Pyrite
Caravaca (*45*)	26	260	Spike	Clay (10 cm)	
Zumaya (*42*)				Clay (30 cm)	Pyrite
DSDP 524 (*11*)	3	110	60 cm	Clay (3 cm)	
DSDP 465 (*35*)	10	630	60 cm	Calcareous ooze	Pyrite
Raton Basin (*65*)	6	60	Spike	Coal (12 cm)	Reducing

DSDP site 465 are much larger and that for the Raton Basin smaller. For sections where there is a boundary layer clay, the fluxes for sites with 1 to 3 cm of clay are about the same but at Caravaca where the clay is 10 cm thick, the flux is much greater.

Kyte *et al.* (*53*) include measurements of other related elements from the Stevns Klint clay (Fig. 5). There are considerable variations in the nickel-normalized abundance ratios with respect to chondrites and to a lesser degree with respect to iron meteorites. Unless fractionation was apparent in the sedimentation and diagenic processes, such variations would not be anticipated for an extraterrestrial source of these elements. As Kyte *et al.* state, "perhaps the key message is that relative fractionation of siderophiles has occurred and that the resulting patterns differ from location to location." Perhaps the relative abundances of these elements are not related to a single extraterrestrial event but rather to a terrestrial source or to changes in the depositional environmental conditions, as suggested by Goldschmidt (*5*).

Rucklidge *et al.* (*60*), using mass accelerator spectrometry techniques, made observations of platinum and iridium in the various constituents of the Danish boundary layer clay at two locations. They report that the largest concentrations occur in the pyrite fraction and suggest that "some marine geochemical process has been at work to achieve this." In particular, the platinum and iridium concentrations are 308 and 51 ppb, respectively, in the heavy fraction of the pyrite-rich clay layer, 51 and 11 ppb in the light fraction of the same layer, and 3 and 3, 61 and 6, 2 and 3 ppb in the underlying gray clay layer and overlying black and brown clay layers. It would seem possible, on the basis of observations of pyrite in the K-T boundary clays, that a reducing environment at the time of deposition was the responsi-

Fig. 5. Abundance ratios normalized to nickel for nine siderophile elements relative to CI chondrites and IIAB iron meteorites. Stevns Klint samples (●) and DSDP site 465 samples (X). [From Kyte *et al.* (*53*)]



ble agent for the observed iridium concentrations.

Further analyses of the clay mineralogy of samples of the K-T boundary layer from Stevns Klint, Gubbio, Caravaca, and El Kef in Tunisia suggest that the base of the clay is neither mineralogically exotic nor distinct from clays above and below the boundary (*61*). The clays at the various localities contain a large smectite component that is probably volcanic in origin.

Finally, Wezel *et al.* (*62*) have found iridium concentrations at the Gubbio section both above and below the K-T boundary layer clay. In particular, the iridium concentration in a 1-m thick cherty black shale at 240 m below the K-T transition has an iridium anomaly twice that of the boundary clay; they infer that the enrichments are of volcanic origin.

It is not clear that the high iridium concentrations are necessarily related to an extraterrestrial event. The observations of a distributed iridium pattern as a function of depth at some sites, the differences in the flux inputs, the variations in noble metal abundances normalized to extraterrestrial concentrations, the depositional effects that might be anticipated in a reducing environment, and the clay mineralogy of the boundary layer clays argue against such a conclusion.

Conclusions

The data that support our conclusions have been presented above. Let us briefly summarize our results.

1) If there has been an extraterrestrial event, then the faunal transitions did not occur instantaneously in response to it; there is a range of transition times and transition time intervals depending on the fossil descriptor and site location.

2) The iridium distribution pattern as a function of depth at some sites, the differences in the flux inputs, the variations in noble metal abundances normalized to extraterrestrial concentrations, the possible effects of a reducing environment, and the clay mineralogy of the boundary layer clays bring into question a large extraterrestrial event as the cause of the iridium anomaly in the vicinity of the K-T transition.

3) The evidence does not support a worldwide shallowing of the calcite compensation depth to the photic zone during K-T transition time. The boundary layer clay is not typical of K-T transition lithology in marine sections but is restricted to Western Europe and the eastern South Atlantic sections.

Investigation of the environmental effects of terrestrial events including extensive volcanism, major regression and transgression of sea level, paleoclimatic, and paleoceanographic changes, and the significance of geochemical changes such as oxidation-reduction conditions and oxygen and carbon isotopic variations may be useful to an understanding of the K-T faunal transitions. The effects of a single environmental change may not be applicable to the entire range of faunal transitions from nannofossils to dinosaurs. Another concept that is important, but difficult, to include is the documentation of environmental effects on speciation, such as that given by Wiedmann (*51*) for Cretaceous ammonites, by Williamson (*63*) for Cenozoic molluscs, and by Dungan *et al.* (*64*) for recent asteroids.

References and Notes

1. B. P. Glass and J. R. Crosbie, *Bull. Am. Assoc. Pet. Geol.* **66**, 471 (1982).
2. W. A. Berggren, *Lethaia* **5**, 195 (1972).
3. R. Ganapathy, *Science* **216**, 885 (1982).
4. L. W. Alvarez, W. Alvarez, F. Asaro, H. V. Michel, *ibid.* **208**, 1095 (1980).
5. V. M. Goldschmidt, *Geochemistry* (Oxford Univ. Press, London, 1951), p. 695.
6. D. M. McLean, *Earth Planet. Sci. Lett.* **53**, 103 (1981).
7. D. I. Axelrod, *Geol. Soc. Am. Spec. Pap. 185* (1981).
8. P. H. Nadeau and R. C. Reynolds, *Nature (London)* **294**, 72 (1981).
9. T. R. Worsley, *ibid.* **230**, 318 (1971).
10. B. E. Tucholke and P. R. Vogt, *Init. Rept. Deep Sea Drill Proj.* **43**, 791 (1979).
11. K. J. Hsü *et al.*, *Science* **216**, 249 (1982).
12. C. Emiliani, *Eos* **61**, 505 (1980).
13. M. A. Arthur and A. G. Fischer, *Geol. Soc. Am. Bull.* **88**, 367 (1977).
14. I. Premoli Silva, *ibid.*, p. 371.
15. W. Lowrie and W. Alvarez, *ibid.*, p. 374.
16. W. M. Roggenthen and G. Napoleone, *ibid.*, p. 378.
17. W. Alvarez *et al.*, *ibid.*, p. 383.
18. D. H. Tarling and J. G. Mitchell, *Geology* **4**, 133 (1976).
19. J. L. LaBrecque *et al.*, *ibid.* **5**, 330 (1977).
20. G. Ness *et al.*, *Rev. Geophys. Space Phys.* **18**, 753 (1980).
21. Shipboard Scientific Party, *Init. Rept. Deep Sea Drill Proj.* **43**, 107 (1979).
22. H. R. Thierstein and H. Okada, *ibid.*, p. 601.
23. E. D. Goldberg and M. Koide, *Geochim. Cosmochim. Acta* **26**, 417 (1962); W. H. Berger and G. R. Heath, *J. Mar. Res.* **26**, 134 (1968); C. B. Officer, *Mar. Geol.* **46**, 261 (1982).
24. N. L. Guinasso and D. R. Schink, *J. Geophys. Res.* **80**, 3032 (1975).
25. C. B. Officer and D. R. Lynch, *Earth Planet. Sci. Lett.* **61**, 55 (1982).
26. Y. Nozaki *et al.*, *ibid.* **34**, 167 (1977); T.-H. Peng *et al.*, in N. R. Anderson and A. Malahoff, Eds., *The Fate of Fossil CO_2 in the Oceans* (Plenum, New York, 1977), p. 355; *Quat. Res. (N.Y.)* **11**, 141 (1979).
27. Shipboard Scientific Party, *Init. Rept. Deep Sea Drill Proj.* **43**, 1049 (1979).
28. P. A. Larson and N. D. Opdyke, *ibid.*, p. 785.
29. H. Thierstein, *Soc. Econ. Paleontol. Mineral. Spec. Publ.* **32**, 355 (1981).
30. K. Perch-Nielsen *et al.*, *Init. Rept. Deep Sea Drill Proj.* **39**, 141 (1977).
31. T. Worsley, *Soc. Econ. Palentol. Mineral. Spec. Publ.* **20**, 94 (1974).
32. K. Perch-Nielsen, *Init. Rept. Deep Sea Drill Proj.* **39**, 699 (1977).
33. Shipboard Scientific Party, *ibid.* **62**, 199 (1981).
34. A. Boersma, *ibid.*, p. 377.
35. H. V. Michel *et al.*, *ibid.*, p. 847.
36. P. Giblin, *ibid.*, p. 851.
37. Shipboard Scientific Party, *ibid.*, p. 157.
38. P. S. Doyle and W. R. Riedel, *ibid.*, p. 491.
39. Shipboard Scientific Party, *ibid.* **6**, 67 (1971).
40. R. G. Douglas, *ibid.*, p. 1027.
41. D. V. Kent, *Geology* **5**, 769 (1977).
42. S. F. Percival and A. G. Fischer, *Evol. Theory* **2**, 1 (1977).
43. J. E. T. Channel and F. Medizza, *Earth Planet. Sci. Lett.* **55**, 419 (1981).
44. J. Smit, in *Cretaceous-Tertiary Boundary Events*, W. G. Christensen and T. Birkelund, Eds. (University of Copenhagen, Copenhagen, 1979). vol. 2, p. 156.
45. J. Smit and J. Hertogen, *Nature (London)* **285**, 198 (1980).
46. E. Voigt, *Newsl. Stratigr.* **10**, 92 (1981).
47. P. Gravesen, in *Cretaceous-Tertiary Boundary Events*, T. Birkelund and R. G. Bromley, Eds. (University of Copenhagen, Copenhagen, 1979), vol. 1, p. 72.
48. E. Hakansson and E. Thomsen, in *ibid.*, vol. 1, p. 78.
49. H. W. Rasmussen, in *ibid.*, vol. 1, p. 65.
50. C. Heinberg, in *ibid.*, vol. 1, p. 58.
51. J. Wiedmann, *Biol. Rev. Cambridge Philos. Soc.* **44**, 563 (1969).
52. R. Ganapathy, *Science* **209**, 921 (1980).
53. F. T. Kyte *et al.*, *Nature (London)* **288**, 651 (1980).
54. R. F. Butler, E. H. Lindsay, L. L. Jacobs, N. M. Johnson, *ibid.* **267**, 318 (1969); E. H. Lindsay *et al.*, *Geology* **6**, 425 (1978); *Am. J. Sci.* **281**, 390 (1981).
55. W. Alvarez and D. W. Vann, *Geology* **7**, 66 (1979); J. E. Fassett, *ibid.*, p. 69.
56. J. F. Lerbeckmo *et al.*, *Nature (London)* **279**, 26 (1979).
57. J. D. Archibald *et al.*, *Geology* **10**, 153 (1982); L. J. Hickey, *Nature (London)* **292**, 529 (1981).
58. T. J. M. Schopf, *Science* **211**, 571 (1981).
59. J. D. Archibald and W. A. Clemens, *Am. Sci.* **70**, 377 (1982).
60. J. C. Rucklidge *et al.*, *Proc. Third Am. Paleontol. Conv.* **2**, 455 (1982).
61. M. Rampino and R. C. Reynolds, *Science* **219**, 495 (1983).
62. F. C. Wezel *et al.*, *C.R. Acad. Sci. Paris* **293**, 837 (1981); F. C. Wezel, *Nature (London)* **294**, 248 (1981).
63. P. G. Williamson, *Nature (London)* **293**, 437 (1981).
64. M. L. Dungan, T. E. Miller, D. A. Thomson, *Science* **216**, 989 (1982).
65. C. J. Orth, J. S. Gilmore, J. D. Knight, C. L. Pillmore, R. H. Tschudy, J. E. Fassett, *ibid.* **214**, 1341 (1981).
66. L. W. Alvarez, W. Alvarez, F. Asaro, H. V. Michel, *ibid.* **211**, 654 (1981).
67. W. Alvarez, F. Asaro, H. V. Michel, L. W. Alvarez, *ibid.* **216**, 886 (1982).