The Cretaceous-Tertiary (K-T) boundary interval, Raton Basin, Colorado and New Mexico, and its content of shock-metamorphosed minerals: Implications concerning the K-T boundary impact-extinction theory

by

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ABSTRACT

At 20 localities in the Raton Basin of Colorado and New Mexico and a few other Western North American Cretaceous-Tertiary (K-T) boundary sites, a pair of thin claystone layers, an iridium abundance anomaly, and a concentration of shock-metamorphosed minerals mark the paleontologic boundary. The lower unit is informally called the "K-T boundary claystone," and the upper unit is informally called the "K-T boundary impact layer." This stratigraphic couplet is generally overlain by a coal bed, informally called the "K-T boundary coal bed" that ranges from 4 to 16 cm in thickness.

In the Raton Basin, the boundary claystone is a 1- to 2-cm-thick tonstein composed of kaolinite and a small amount of illite/smectite (I/S) mixed-layer clay. Scanning electron microscopic (SEM) study shows that the claystone mainly consists of a polygonal boxwork filled with micrometer-sized kaolinite microspherules. The lower part of the claystone contains angular fragments of cryptocrystalline kaolinite nearly as large as 1.0 mm. Typically, small amounts of carbonaceous material including horizontal vitrinite laminae, subvertical carbonaceous structures, plant impressions, swirls of vitrinite, and millimeter-sized rounded fragments of cellular fusinite occur in the claystone. In Saskatchewan, Canada, the claystone contains inclusions of yellow amber, as much as 2.5 mm in diameter. The claystone has a trace-element content similar to average North American shale, except that it contains about 0.07 to 0.32 ppb iridium. An equivalent of the K-T boundary claystone seemingly does not occur at marine K-T boundary sites outside of North America.

Solid kaolinite and hollow and solid goyazite spherules, 0.05 to 1.2 mm in diameter, occur in the K-T boundary claystone but not in the overlying impact layer. Goyazite spherules are rare and widely scattered in the claystone in the Raton Basin, but are common in a 1-cm-thick goyazite-enriched layer at the top of the claystone at Wyoming K-T boundary sections. In Wyoming, they compose as much as 30 percent of the goyazite-enriched layer. SEM and thin section study of the goyazite and kaolinite spherules indicates that they formed authigenically and are not of impact origin.

The upper unit, the K-T boundary impact layer, is typically 5 mm thick in the Raton Basin and elsewhere in Western North America, whereas it is only about 1 mm thick at Caravaca, Spain, and Stevns Klint, Denmark. The claystone consists chiefly of kaolinite and various amounts of I/S mixed-layer clay. Typically the claystone is microlaminated and contains planar laminae of vitrinite and ubiquitous kaolinite pellets. The SEM texture of the claystone is, in general, similar to smectite clay and is different from the underlying microspherulitic boundary claystone. The contact between the impact layer and the underlying boundary claystone is generally sharp and records a significant change in depositional regime.

Of importance to the impact-extinction theory is the fact that the impact layer contains as much as 2 percent clastic mineral grains, 30 percent of which contain multiple intersecting sets of planar lamellae clearly of shock-metamorphic origin. Only one such concentration of shock-metamorphosed minerals has been found near the K-T boundary. Of the grains that contain shock lamellae, quartzite, metaquartzite, and chert constitute about 60 percent and quartz the remainder. Grains of shocked feldspar and granite-like mixtures of quartz and feldspar are rare. The abundance of unshocked quartzite, metaquartzite, and chert in the impact layer and their paucity in underlying rocks suggests they are target materials of impact origin. The shocked minerals and an unknown phase (perhaps originally an aerosol) that carried iridium are the only impact-related components in the claystone.

Two sources of information, one from the compositional types of K-T boundary shock-metamorphosed minerals and the second from the study of pumice fragments in silicic pyroclastic rocks eliminates the possibility that the shock-metamorphosed minerals are of volcanic origin.

The global size distribution of shock-metamorphosed mineral grains in the impact layer suggests that the K-T impact occurred in North America. At Western North American K-T boundary sites, the mean size of shocked quartz grains is as follows: Raton Basin, Colorado and New Mexico (0.20±0.06 mm to 0.16±0.06 mm), Teapot Dome, Wyoming (0.14±.04 mm), Brownie Butte, Montana (0.15±0.05 mm), Alberta, Canada (0.26±0.06 mm). Rare grains, as long as 0.50 to 0.64 mm, are found at all Western North American sites. The mean size of 100 shocked quartz grains in the impact layer at Caravaca, Spain, is 0.09±0.03 mm, and the range is 0.04-0.19 mm, considerably less than the mean size of shocked quartz grains at North American K-T boundary sites (0.14-0.26 mm). Forty shocked quartz grains from upper Eocene glassy impact sediment in DSDP site 612 off the New Jersey Coast range in size from 0.1 to 0.5 mm and average 0.26±0.1 mm.

The chemical composition of the impact layer and boundary claystones are similar, except that the former contains slightly more iron, potassium, barium, chromium, copper, lithium, vanadium, and zinc than the latter. Like the boundary claystone, the impact layer claystone has a low content of nickel and cobalt, only a few ppm. Radionuclide activation analyses of rocks that span the K-T boundary show that iridium is generally greatest in the K-T boundary impact layer; nevertheless, anomalously large values also can occur in carbonaceous-rich layers, particularly coal, either below or above the impact layer. Amounts of iridium vary from 1.2 to 14.6 ppb; these amounts are considerably more (5 to 66 times) than those in the underlying boundary claystone. The surface concentration of iridium varies considerably at localities only a few kilometers apart, from 8 to 120 ng/cm^2. Because shock-metamorphosed minerals are concentrated in the impact layer, the assumption was made that the iridium also was originally concentrated in this layer. During diagenesis of the sediments, some iridium was probably mobilized and transported away from the impact layer and concentrated in adjacent carbonaceous-rich sediments.

The origins of the K-T boundary claystone and K-T boundary impact layer are seemingly closely related because they form a stratigraphic couplet at numerous Western North American sites. However, considerable observational and chemical evidence suggests that the claystone matrix of the units is not altered volcanic or impact-generated material. The most important observational fact that bears on the origin of the K-T boundary claystone is that shock-metamorphosed minerals are restricted essentially to the K-T boundary impact layer and are not found in the underlying K-T boundary claystone. It is difficult to understand how the boundary claystone could originate from impact material that would be entirely free of clastic mineral grains, in particular, shock-metamorphosed minerals.

The Manson, Iowa, structure is proposed as the K-T impact site because of the mineralogic similarity of Manson subsurface rocks and shocked K-T boundary minerals, the large size (35 km) of the impact structure, the compatible isotopic age of shocked granitic rock from the Manson impact structure and the K-T boundary (66 Ma), and the proximity of the Manson impact structure to North American boundary sites that contain relatively abundant and large shocked minerals. Objections can be raised that the Manson, Iowa, impact structure is not the K-T boundary impact site because it is too small. Inherent in this reasoning are the assumptions that the composition, mass, strike-angle of the asteroid are known. None of these assumptions may be warranted, and, thus, the Manson impact structure is a viable candidate site for the K-T boundary impact. The killing mechanism for the K-T extinction event may not have been the lofting of large volumes of dust into the atmosphere and the ensuing global darkness as was proposed by the Alvarez team. Instead the killing mechanism may have been the generation of large volumes of nitrogen oxides and carbon dioxide produced in the atmosphere from the K-T impact. Acid rain generated during the impact would have a disastrous effect on plant life, particularly in the Western Interior of North America, and the biota of the photozone of the oceans. An increase in the atmosphere of impact-generated carbon dioxide may have caused a green house effect and resulting global warming.

INTRODUCTION

In 1979 and 1980, a team of scientists including L.W. Alvarez, Walter Alvarez, Frank Asaro, and H.V. Michel of the University of California reported finding trace amounts of the platinum-group metal iridium in a 1-cm-thick marine claystone bed at the paleontologic Cretaceous-Tertiary (K-T) boundary near Gubbio, Italy (Alvarez and others, 1979a, 1979b, 1980). The peak amount of iridium they measured, 9.1 parts per billion (ppb), in the claystone bed was about 30 times above that in subjacent Cretaceous and superjacent Paleocene limestone beds. Their discovery was significant because iridium generally occurs in minute amounts, less than 0.1 ppb, in most rocks that comprise the Earth's crust (Palme, 1982, fig. 1). It is present in much greater concentration (500-1,000 ppb) in primitive cosmic materials such as chondritic meteorites and presumably in the Earth's core. They also reported finding anomalously large amounts of iridium in marine sedimentary rocks at the K-T boundary at such widely scattered localities as Denmark and New Zealand. Their purpose in analyzing the claystone for the siderophile element iridium was to determine, if possible, the time interval during which the claystone bed was deposited (Alvarez and others, 1980, p. 1096). They apparently were aware of the suggestion of Pettersson and Rotschi (1952, p. 90) that anomalously large amounts of platinum-group metals, including iridium and osmium, might be found in deep ocean-floor sediments owing to the influx of meteoritic dust. If the influx rate of meteoritic dust into the ocean sediments were constant, it would provide a method for calculating the deposition rate of the sediments. Subsequently, Barker and Anders (1968, p. 627) measured trace amounts of osmium and iridium in sea-floor sediments by the instrumental neutron activation method and set an upper limit for the accumulation rate for these elements.

Because the iridium anomaly coincides exactly with the classic mass extinction of animals that occurred at the close of the Cretaceous (Loeblich and Tappan, 1964; Raup and Sepkoski, 1982), the Alvarez team proposed a theory, now called the "impact-extinction theory," linking these two factual observations. They speculated that a 10±4-km-sized Apollo asteroid struck Earth and formed a crater 200 km in diameter. The Alvarez team suggested that the impacting body was an asteroid rather than a comet. I also refer to the extraterrestrial object as an asteroid, but for discussion purposes only, because the exact nature of the impacting body is unknown. They proposed that enormous amounts of fine-grained material (60 times the asteroid mass) were lofted into the stratosphere, and some of the dust-sized impact ejecta stayed aloft and blocked sunlight for several years. As a result, global darkness followed, the level of photosynthesis markedly decreased, and the food chain collapsed, resulting in the extinction of about half of Cretaceous life forms including the dinosaurs. This extinction event has been used to define the end of Cretaceous time at 66 Ma. I prefer to use 66 Ma for the numerical isotopic age of the K-T boundary, rather than the commonly cited 65 Ma for reasons discussed elsewhere in this report.

The discovery of the iridium anomaly by the Alvarez team triggered a search by many scientists for evidence that would test the impact-extinction theory. Considerable debate has arisen, especially in regard to the fabric and timing of the mass extinction event, the composition and size of the extraterrestrial body, and the location and mechanics of formation of the impact crater. There has been an enormous growth in the number of papers reporting results of research since 1980. Reports have been published by astronomers, biologists, biostratigraphers, geochemists, geologists, geophysicists, oceanographers, mineralogists, paleontologists, physicists, planetary geologists, and experts in cratering mechanics. So many papers have been written that no attempt is made here to list the numerous reports generated as a result of the Alvarez theory. References can be found in papers published as a result of a conference titled "Large Body Impacts and Terrestrial Evolution--Geological, Climatological, and Biological Implications" held at Snowbird, Utah, in 1981 (Silver and Schultz, 1982). Paleontologic papers and information can be found in the proceedings of (1) a symposium "Cretaceous-Tertiary Boundary Events" (Birkelund and Bromley, 1979); (2) a workshop "Cretaceous-Tertiary Extinctions and Possible Terrestrial and Extraterrestrial Causes" (Russell and Rice, 1982); (3) a book "Catastrophes and Earth History" (Berggren and Van Couvering, 1984); and (4) the abstracts and papers of a conference on mass extinction events held in Bilbao, Spain, in 1987 "Conference on Paleontology and Evolution: Extinction Events" (Lamolda and Cearreta, 1987).

An important prediction of the impact-extinction theory is that an iridium anomaly should be found globally at complete K-T boundary stratigraphic sections. This prediction has been elegantly confirmed because more than 73 boundary sections containing an abundance anomaly of iridium (and other noble metals) have been reported (Smit and Hertogen, 1980; Ganapathy, 1980; Luck and Turekian, 1983; Alvarez and Montanari, 1985; Kyte and others, 1985; Lichte and others, 1986). Be this as it may, the origin of geochemical anomalies such as the K-T boundary iridium spike are open to multiple interpretations, and it is not surprising that a few earth scientists have suggested alternative ideas to explain its origin. Officer and Drake (1985), McLean (1985), and Officer and others (1987) proposed that the iridium is of terrestrial volcanic rather than of extraterrestrial impact origin, but Alvarez (1986) has argued cogently in defense of the impact origin of the iridium.

Although the origin of the K-T boundary geochemical iridium anomaly is controversial, there is directly observable <u>physical evidence</u> indicating that a large extraterrestrial object struck Earth at the close of the Cretaceous 66 Ma. In K-T boundary rocks in the Hell Creek area of eastern Montana, Bohor and others (1984) discovered quartz grains that contain an unusual type of microstructure consisting of multiple intersecting sets of closely spaced planar lamellae. They concluded that the planar lamellae are identical to those formed in quartz that has undergone high strain rates during hypervelocity shock in laboratory experiments and at meteor impact sites and nuclear and chemical explosion sites. For brevity, I use alternatively the term "shock lamellae" rather than the cumbersome term "multiple intersecting sets of planar lamellae."

Following the discovery by Bohor and others (1984) of shock-metamorphosed quartz grains in K-T boundary rocks in the Hell Creek area of eastern Montana, Izett and Pillmore (1985a, 1985b) reported finding not only grains of shock-metamorphosed quartz, but also grains of shock-metamorphosed oligoclase and a potassium-rich feldspar that is probably microcline. Of special interest was the discovery in the Raton Basin of Colorado and New Mexico of shock-metamorphosed composite grains that, in essence, are small lithic fragments of metaquartzite, quartzite, and granite-like material. On a global scale, shock-metamorphosed mineral grains have been found in K-T boundary rocks in Italy, Denmark, Spain, New Zealand, Canada, the North Pacific Ocean, Asia, and Austria (Bohor and Izett, 1986; Badjukov and others, 1986; Preisinger and others, 1986; Bohor and others, 1987a; Alexopoulos and others, 1987). These discoveries provide the principal <u>physical evidence</u> supporting the impact component of the impact-extinction theory of the Alvarez team.

Concerning the "extinction" component of the impact-extinction theory, many ideas have been advanced to explain major mass extinction events such as the one at the close of the Cretaceous. For example, Laubenfels (1956) suggested that the impact of an extra large meteorite caused the extinction of the dinosaurs. McLaren (1970) and Urey (1973) proposed an extraterrestrial cause for major mass extinction events in the Phanerozoic, but unlike the Alvarez team, they did not present any direct evidence to support the impact theory.

The proposal that the impact had a profound effect on Cretaceous life forms, as predicted by the impact-extinction theory, has not been accepted by many geologists, especially paleontologists that deal with marine fossils (see review by Hallam, 1987). Many paleontologists believe that the extinction event at the close of the Cretaceous is only a part of a broad pattern of extinctions that began in the Late Cretaceous in response to lowering of sea level. Kauffman (1984, p. 154) recently wrote that "Despite its classic status as a major catastrophe, therefore, the biological fabric of the terminal Cretaceous extinction is largely unknown." He (1984, p. 199) further wrote "To sum up, data on the Cretaceous-Tertiary extinction event are at best sparse, especially among the macrobiota, and reflect the relatively few sections in the world that preserve complete or nearly complete boundary sequences." He concluded (1984, p. 237) that "* * * diverse evidence strongly suggests that Terminal Cretaceous extinction was graded over a 1 to 5 myr in the marine realm * * * and was primarily the result of massive environmental deterioration." Alvarez and others (1984, p. 1135) observed that prior to 1980 "* * * the weight of informed paleontological opinion held that the terminal-Cretaceous extinction was gradual on a time scale of (1 to 10) X 10^6 years * * *." Surely, a complete evaluation of the effects of the impact on marine Cretaceous life forms will take more biostratigraphic data than are now available. More detailed biostratigraphic studies such as the one dealing with brachiopods near the K-T boundary (Johansen, 1987) are needed before the effects on Cretaceous life forms can be understood.

The influence of the asteroid impact on land plants was minimal according to Hickey (1984, p. 303). He stated that "* * * the relative continuity of the angiosperm record across the K/T boundary; the generally moderate and geographically variable levels of land-plant extinction; and the lack of synchroneity in plant-dinosaur extinctions make it highly unlikely that a universal biotic catastrophe occurred." On the other hand, Tschudy and others (1984) believed that a massive destruction of vegetation and the extinction of a few plant species occurred as a result of some catastrophic event at the close of the Cretaceous Period. More recently, Tschudy and Tschudy (1986, p. 667) stated that "No massive total extinction of plant genera at the end of the Cretaceous can be seen from the palynological record." They further pointed out that the flora of the Western Interior of North America underwent a profound shock at the exact K-T boundary. Apparently this floral break was limited to the Western Interior of North America.

The effect of the K-T boundary asteroid on land animals such as the dinosaurs has been a highly controversial subject in the last several years. Alvarez and others (1980) and Alvarez (1983) suggested that the impact of an asteroid caused the extinction of the dinosaurs. Concerning the extinction of the dinosaurs, Schopf (1982, p. 416) stated that "The major point of this and the previous paragraphs is that dinosaurs appear to be last known from beds older than latest Maastrichtian in all places except in northern Europe and in the western interior of North America." He (1982, p. 420) noted that "* * * with regard to dinosaurs, so far as is currently known, only a handful of species existed in the latest Cretaceous, and these may have been living under precarious ecologic conditions." Archibald and Clemens (1982, p. 380) observed that fossil dinosaur remains disappear from the stratigraphic section in the Hell Creek area of Montana about 3 m below a coal bed that marks the K-T boundary. In Alberta, Canada, Lerbekmo and St. Louis (1985) indicated that the stratigraphically highest dinosaur remains are 4.5 m below the paleontologic K-T boundary as marked by shock-metamorphosed minerals and an iridium abundance anomaly (see Izett, 1987a). In the Lance Creek area of Wyoming, Bohor and others (1987b, p. 899) found dinosaur remains 1 m below the K-T boundary as defined by the triple occurrence of a pollen break, a concentration of shock-metamorphosed minerals, and an iridium abundance anomaly. Sloan and others (1986) suggested that the extinction of the dinosaurs was a gradual process that began 7 million years before the end of the Cretaceous. Recently, Rigby and others (1987) asserted that dinosaur teeth ocxcur within alleged Paleocene channel-fill deposits in the Hell Creek area of Montana. Apparently, neither the K-T boundary claystone, shock-metamorphosed minerals, nor an iridium abundance anomaly have been found at their specific fossil localities. Nevertheless, they believe that the dinosaurs survived the K-T boundary impact event.

PURPOSE OF THE STUDY

Descriptions of some aspects of the stratigraphic location and mineralogic features of shock-metamorphosed minerals in K-T boundary rocks in Western North America have been reported (Bohor and others, 1984, 1987a; Izett and Pillmore, 1985a, 1985b; Izett and Bohor, 1986a). However, important stratigraphic and mineralogic details have not been reported previously, and these can be used to constrain models for the K-T boundary impact event. It is, therefore, the purpose of this study to (1) briefly describe the sedimentary rocks that constitute the K-T boundary interval chiefly in the Raton Basin of Colorado and New Mexico, but also at other complete K-T boundary sections in Western North America, (2) provide petrographic information pertinent to the shock-metamorphosed minerals in the K-T boundary impact layer, (3) present mineralogic evidence that indicates the minerals containing shock lamellae are of impact and not volcanic origin, and (4) suggest that the K-T boundary asteroid struck Earth in an area underlain by continental crust composed of metamorphic, sedimentary, and granitic rock, such as the area of the Manson, Iowa, impact structure.

It is important to keep in mind that the size, composition, density, velocity, and entry angle of the K-T boundary asteroid will never be known with assurance. Thus, the intensity and magnitude of the K-T boundary impact and impact processes, including the transportation and deposition of impact ejecta, are also largely unknown. For these and other reasons, some of the interpretations and conclusions given in this study are controversial and provocative.

Until recently, distal air-fall (or reworked air-fall) impact-ejecta deposits, other than those described in this study, have not been reported in the geologic literature. Gostin and others (1986) found a deposit of air-fall impact material in Proterozoic sedimentary rocks of South Australia, and Glass (1987) and Thein (1987) described a layer of coarse-grained glassy impact ejecta at the base of an 8- to 12-cm-thick bed in core of DSDP site 612 that lies on the continental slope off the coast of New Jersey.

K-T BOUNDARY INTERVAL IN THE RATON BASIN

In the Raton Basin, the K-T boundary occurs in a sequence of fine-grained, continental sedimentary rocks deposited in fluvial, lacustrine, and swamp environments following the retreat of the Western Interior Cretaceous seaway. The uppermost Cretaceous and lowermost Tertiary rocks include, from oldest to youngest, the marine Pierre Shale and Trinidad Sandstone and the continental Vermejo and Raton Formations. The Pierre is about 600 m and the Trinidad about 30 m thick. The Vermejo Formation, which varies from 0 to 115 m in thickness, overlies the Trinidad and consists of alternating beds of sandstone, siltstone, carbonaceous shale, mudstone, and coal. The Vermejo is overlain unconformably by the Raton Formation of Cretaceous and Paleocene age, and the contact is marked at some places by a conglomeratic sandstone at the base of the Raton Formation. The K-T boundary occurs in the lower part of the Raton Formation in a sequence of rocks similar to those of the underlying Vermejo Formation. The very generalized stratigraphic framework outlined above is a condensed version of descriptions given by Pillmore and Flores (1987).

The boundary interval is composed of a pair of thin claystone units informally called herein the "K-T boundary claystone bed" and "K-T boundary impact layer" and a thin, persistent coal bed informally called the "K-T boundary coal bed" (fig. 1). Because these and adjacent units were deposited in low-energy depositional regimes, they are especially well suited for study of fine-scale sedimentary structures that record in detail K-T boundary depositional events. Moreover, the lithologic succession that contains the boundary is similar at each locality in the Raton Basin, and this is surprising considering that the rocks were deposited in continental sedimentary environments commonly noted for their lithologic heterogeneity. More surprising, the same sequence of K-T boundary units that occurs in the Raton Basin has been identified by me in samples from boundary outcrops in the Lance Creek, Wyo., area discovered by Bohor and others (1987b), in the western Powder River Basin, Wyo., area discovered by J.A. Wolfe of the U.S. Geological Survey (Wolfe and Izett, 1987), and at Brownie Butte in the Hell Creek area of Montana discovered by Bohor and others (1984). Farther north, I have tentatively identified these same units in samples from the Morgan Creek, Saskatchewan (Nichols and others, 1986) and Red Deer Valley, Alberta, areas (Lerbekmo and St. Louis, 1985). At all of these Western North American sites, there is no record in the fine-grained sedimentary rocks that span the boundary of any pronounced disruption of the sedimentological regime (such as channels or coarse sediments) produced by the impact of the K-T boundary asteroid. The degree of uniformity and lateral extent of these fine-grained units over such a large area of Western North America is puzzling in view of the cataclysmic K-T impact scenarios envisioned by some (Alvarez and others, 1980; Wolbach and others, 1985; Prinn and Fegley, 1987).

A)

B)

Figure 1.--Polished slabs of the K-T boundary interval from the Raton Basin, Colo. The K-T boundary interval consists of carbonaceous shale, the boundary claystone, the impact layer, and the boundary coal bed. A, Clear Creek North locality, 10 km south of Trinidad, Colo. Two prominent carbonaceous subvertical root-like structures (shown at arrows) transect the boundary claystone. Clastic quartz and other mineral grains are found along such features but not in surrounding claystone. B, Berwind Canyon locality, about 20 km north of Trinidad, Colo. Carbonaceous shale below boundary claystone lost during preparation.

ROCKS DIRECTLY UNDERLYING THE K-T BOUNDARY CLAYSTONE

The rocks of the Raton Formation that directly underlie the K-T boundary claystone are chiefly composed of several tens of centimeters of mudstone, shale, coal, or sandstone. The clay minerals forming the mudstone and shale are composed of kaolinite and ordered and randomly ordered illite/smectite (I/S) mixed-layer clay (C.G. Whitney, U.S. Geological Survey, written commun., 1986). Where the rocks below the boundary claystone are shale and mudstone, they commonly contain a small amount of 0.1-mm-diameter detrital grains of quartz, feldspar, metaquartzite, quartzite, and chert. At most places, the mudstone and shale are dark brownish gray and carbonaceous; at other places, they are light brownish gray and less carbonaceous. For example, at the Raton, N. Mex., site, light-gray mudstone several tens of centimeters thick lies below the K-T boundary claystone bed. Fragments of macerated plant material and leaf impressions occur on many bedding surfaces of the carbonaceous shale. Locally, the shale and mudstone contain rounded blebs of cellular fusinite as much as 1.0 mm in diameter. The mudstone and carbonaceous shale that generally underlie the K-T boundary claystone are probably of fluvial and paludal origin as indicated by the laminated character and content of macerated plant debris. On a microscopic scale, the bedding consists of generally planar laminations whose presence is emphasized by discontinuous laminae of coaly carbonaceous material that is probably vitrinite (N.H. Bostick, U.S. Geological Survey, written commun., 1986). Discontinuous 3.0-mm-thick laminae and lenses of a jarosite-like mineral, probably derived from the alteration of pyrite, are also scattered throughout the unit along bedding planes or joints. The jarosite-like mineral, which contains small amounts of sodium, is referred to as jarosite elsewhere in this report. The mudstone and claystone must have been deposited along low-relief floodplains in an exceedingly low-energy regime as indicated by the finely laminated character of the shale and by the small amount and small size of included clastic mineral grains.

At two localities, Sugarite, N. Mex., and Madrid East, Colo. (fig. 2), the rocks below the K-T boundary claystone consist of coal and impure coal, respectively. At another locality, Clear Creek South, Colo., the rock unit below the boundary claystone is fine-grained sandstone rather than the typical carbonaceous shale.

Megascopically, the contact between dark-gray carbonaceous shale and mudstone and the overlying boundary claystone appears sharp and conformable (fig. 1A). However, microscopic study of sawed, polished surfaces of boundary interval specimens reveals that the contact is gradational and undulatory on a small scale. The undulatory nature of the contact probably resulted from original surface irregularities and differential soft-sediment compaction and deformation during diagenesis of the carbonaceous-rich sediment.

The K-T boundary in the continental sedimentary rocks of the Raton Basin and elsewhere in the Western Interior of North America is placed at the abrupt disappearance of several palynomorphs (Tschudy and Tschudy, 1986). Dinosaur remains have been used to locate the K-T boundary in the Western Interior, but they have not been found in the Raton Basin. Curiously, dinosaur remains occur in uppermost Cretaceous rocks to the south and north in the San Juan and Denver Basins. The interpretation that the pollen-defined K-T boundary in continental rocks of the Western Interior is coeval with the fossil-defined K-T boundary in marine rocks is based on the observations that both boundaries occur in a reverse polarity magnetozone, 29R, (Shoemaker and others, 1987, p. 148) and contain anomalously large amounts of platinum-group metals and a spike of shock-metamorphosed minerals (Bohor and Izett, 1986; Izett, 1987a).

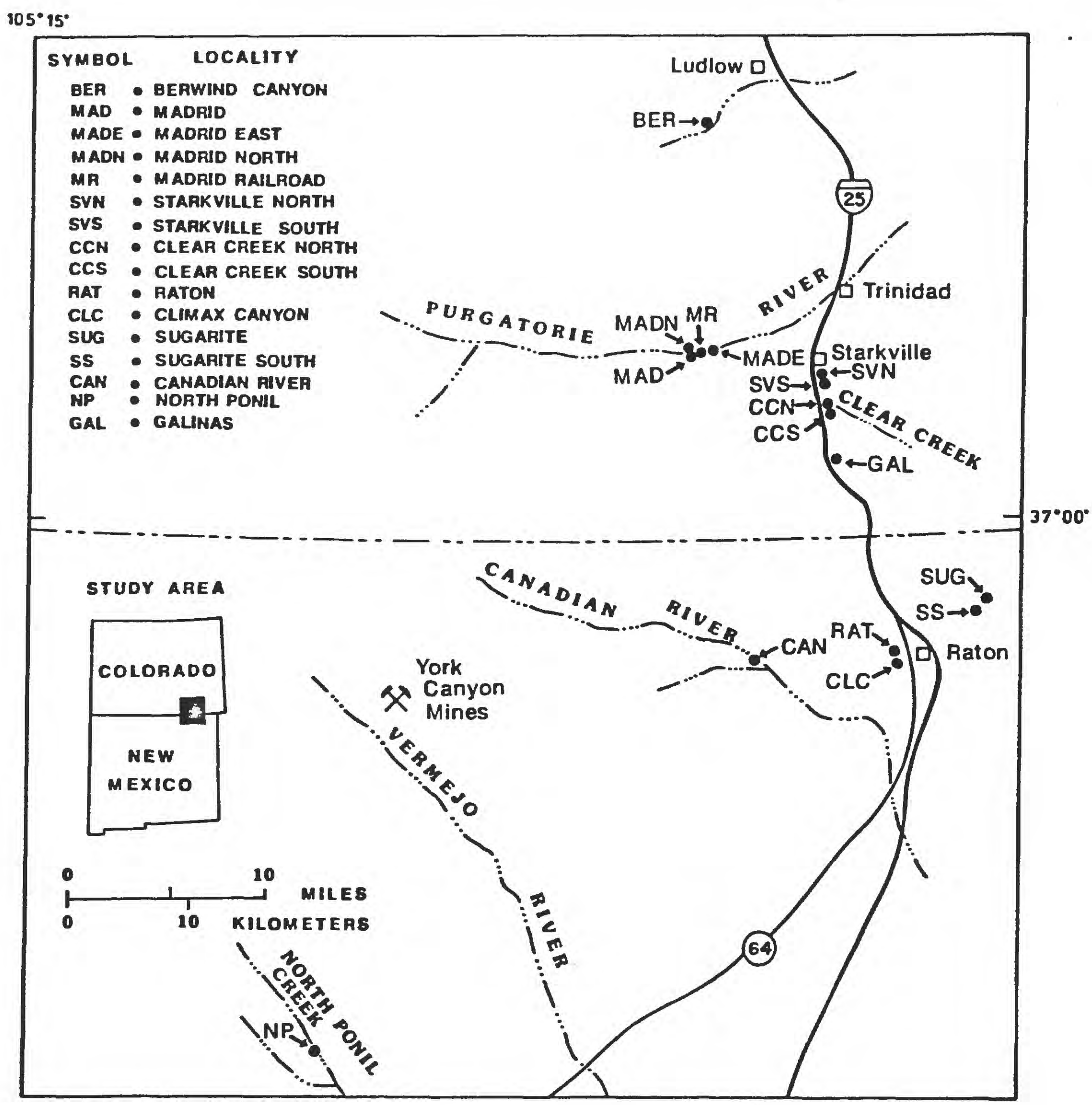


Figure 2.--Index map showing locations of K-T boundary sections in the Raton Basin mentioned in the text. A list of abbreviations used in tables 1-7 for these sections is given. Modified from Pillmore and Flores (1987).

Near Trinidad, Colo., in the Raton Basin, the paleontologic K-T boundary occurs about 20 m above the base of the Raton Formation (Pillmore and Tschudy, 1983). On the basis of the regional extinction of several palynomorph taxa (i.e., Proteacidites), the K-T boundary was placed by Tschudy and others (1984) and Tschudy and Tschudy (1986) at the top of the boundary claystone. This was an arbitrary choice because only rare palynomorphs have been recovered from the boundary claystone, and these might have been reworked into the unit from underlying Cretaceous sediments. The rocks just above the boundary claystone, which contain abundant sapropel and decaying vegetation, were termed the "barren-and-recovery interval" by Tschudy and Tschudy (1986, p. 668). An abrupt change in the ratio of fern spores to angiosperm pollen also occurs at this stratigraphic level.

K-T BOUNDARY CLAYSTONE BED

The most easily recognized microstratigraphic unit in the K-T boundary interval in the Raton Basin was called the "K-T boundary clay" by Pillmore and others (1984). Within the basin, its stratigraphic level has been traced for nearly 80 km and identified at 20 spot localities by C.L. Pillmore of the U.S. Geological Survey (Pillmore and Flores, 1987). The locations of K-T boundary sections in the Raton Basin are shown on figure 2. Because the boundary claystone forms a thin, discrete layer within a coal bed or directly below a coal bed, it is similar to claystone partings in coal sequences called "tonsteins" by European geologists (Williamson, 1970). It is also similar in its stratigraphic setting and certain lithic characteristics to claystone layers below coal beds termed "underclays" (Huddle and Patterson, 1961). Some geologists, including Spears and Rice (1973) and Bohor and Pillmore (1976), have maintained that tonstein layers are altered volcanic ash, and, indeed, some may be (Seiders, 1965). However, others have favored an origin involving either the biogenic accumulation of a silica-alumina gel or the accumulation and alteration of a previously formed clay. Williamson (1970) presented a summary of the various theories concerning the origin of tonstein layers.

The K-T boundary claystone is a discrete, thin layer of uniform, cryptocrystalline claystone that contrasts markedly with the adjacent sedimentary rocks. Its light-gray to grayish-brown color is in sharp contrast to subjacent dark-gray carbonaceous shale and superjacent black coal (fig. 1). It is normally 1-2 cm thick, and, for such a thin unit, it is astonishingly uniform in thickness. The claystone is moderately indurated and, when fractured, breaks into angular fragments and crumbs.

One of the chief characteristics of the claystone is its content of carbonaceous material. At most localities, the claystone contains centimeter-long horizontal vitrinite laminae and convoluted swirls of vitrinite. Another type of carbonaceous material forms peculiar structures that can be seen on sawed and polished slabs of the claystone cut normal to bedding (figs. 1 and 3). These features are subvertical to vertical and consist of poorly delineated, disjointed, and deformed trains of vitrinite. Some of these are funnel-shaped in cross section, that is, they appear to flare out at the contact with the overlying impact layer (fig. 3B). Others are smeared out horizontally at their upper ends within the claystone (fig. 3A). Where they flare out at the contact with the overlying impact layer, carbonaceous material may be concentrated. Some of these structures extend from the base to the top of the claystone bed. The carbonaceous structures have not been seen to cut the bedding of the overlying impact layer. In sections cut parallel to bedding, most of the structures are irregular in shape, but a few seem to be roughly tubular in shape. The origin of the structures is

A)

B)

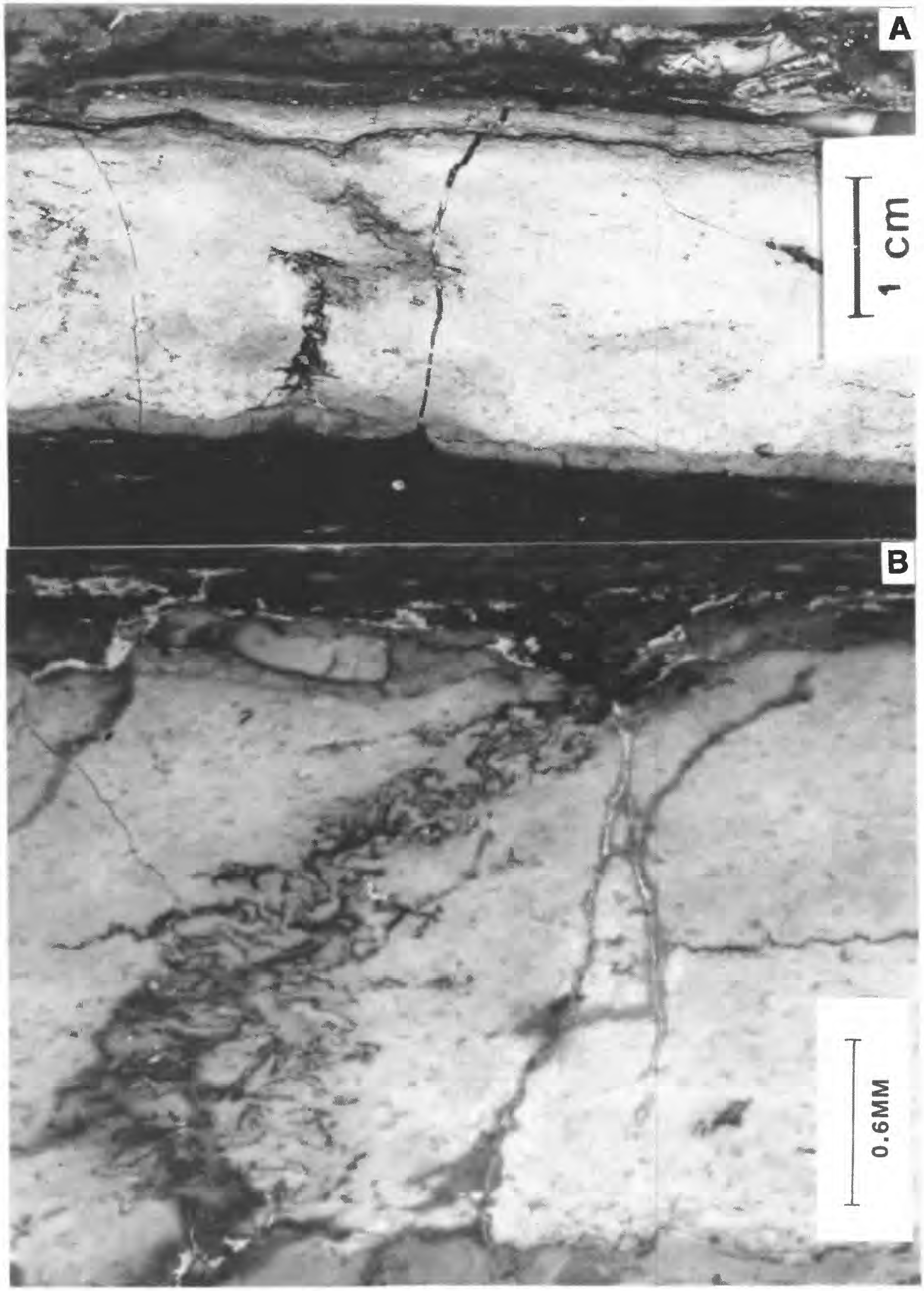


Figure 3.--Carbonaceous root-like structures in the K-T boundary claystone from the Clear Creek North, Colo., locality. A, Root-like structure in boundary claystone smeared out near middle of the bed. B, Root-like structure contains a concentration of carbonaceous material at the contact between the boundary claystone and impact layer. Clastic quartz and other mineral grains occur along such features but not in surrounding claystone.

problematic, but they may be the remains of roots of plants that grew on a surface at the top of the boundary claystone prior to the deposition of the overlying K-T boundary impact layer.

In addition to the carbonaceous structures just described, the boundary claystone contains several types of fossil plant material, including the impressions of plant debris. At many places in the Raton Basin, the boundary claystone contains millimeter-sized rounded fragments of cellular fusinite that generally occur in the lower part of the bed. At the Madrid East, Colo., locality, a centimeter-sized chunk of rounded carbonized, cellular, woody material occurs in the upper part of the claystone bed fig. 4). The claystone matrix is totally free of detrital quartz and feldspar grains. Within the boundary claystone at the Teapot Dome locality in Wyoming, I found fossil roots along bedding planes and the corm (nodule on a root) of a plant, possibly a water lily or sedge (identified by J.A. Wolfe, 1987). At the Brownie Butte, Mont., locality, the claystone contains fossil impressions of twigs as much as 2 cm in length.

In samples of the boundary claystone from the Morgan Creek locality in Saskatchewan, Canada (Nichols and others, 1986), I found blebs of yellow amber, as much as 2.5 mm in diameter. The blebs of amber occur at various stratigraphic positions in the claystone. They formed presumably from drops of conifer sap that sporadically fell directly into the accumulating sediment that now forms the boundary claystone. The occurrence of the blebs of amber in the boundary claystone is of considerable interest because Berner and Landis (1987) determined that the composition of gas in bubble inclusions in Eocene and Oligocene amber is similar to that of present-day air. If the blebs of amber in the boundary claystone contain bubble inclusions and entrapped gas, the gas might be of an unusual composition if, as some geologists believe, the boundary claystone formed during the K-T impact event.

The clay mineralogy of the boundary claystone was determined by X-ray diffraction analysis by Pollastro and Pillmore (1987) and as part of this study by C.G. Whitney (U.S. Geological Survey, written commun., 1986). Five samples from the Raton Basin and one from the Brownie Butte, Mont., locality were analyzed. The analyses show that the claystone consists of well crystallized kaolinite and a small amount of random and ordered I/S mixed-layer clay.

The pervasive groundmass texture of the boundary claystone is cryptocrystalline, and it differs in this respect from other tonsteins of the Raton Basin, which are coarse grained. The term "cryptotonstein" was applied by Retallack (1976, p. 390) to such cryptocrystalline tonsteins, and he suggested that they possibly form by biochemical processes in soils. The microscopic texture of the claystone is a micrometer-sized aggregate of kaolinite granules and diffuse lenticules of complexly arranged accordion-like kaolinite booklets. The uppermost part of the claystone contains angular patches, a few tenths of a millimeter in length, of a feathery-appearing clay mineral having a higher birefringence than the adjacent kaolinite groundmass. This mineral, probably I/S mixed-layer clay, may have formed in vugs because Fastovsky and McSweeney (1987, fig. 2) described and illustrated similar vugs in Cretaceous and Tertiary boundary paleosols. Scanning electron microscopic (SEM) study of the groundmass claystone reveals that it mainly consists of a polygonal boxwork filled with micrometer-sized kaolinite microspherules (fig. 5A). The arrangement of the spheres is similar to the hexagonal close packing lattice. The texture of this groundmass suggested to Pollastro and Pillmore (1987) that the precursor of the claystone was glass. This texture also occurs in claystone that fills cells of degraded plant

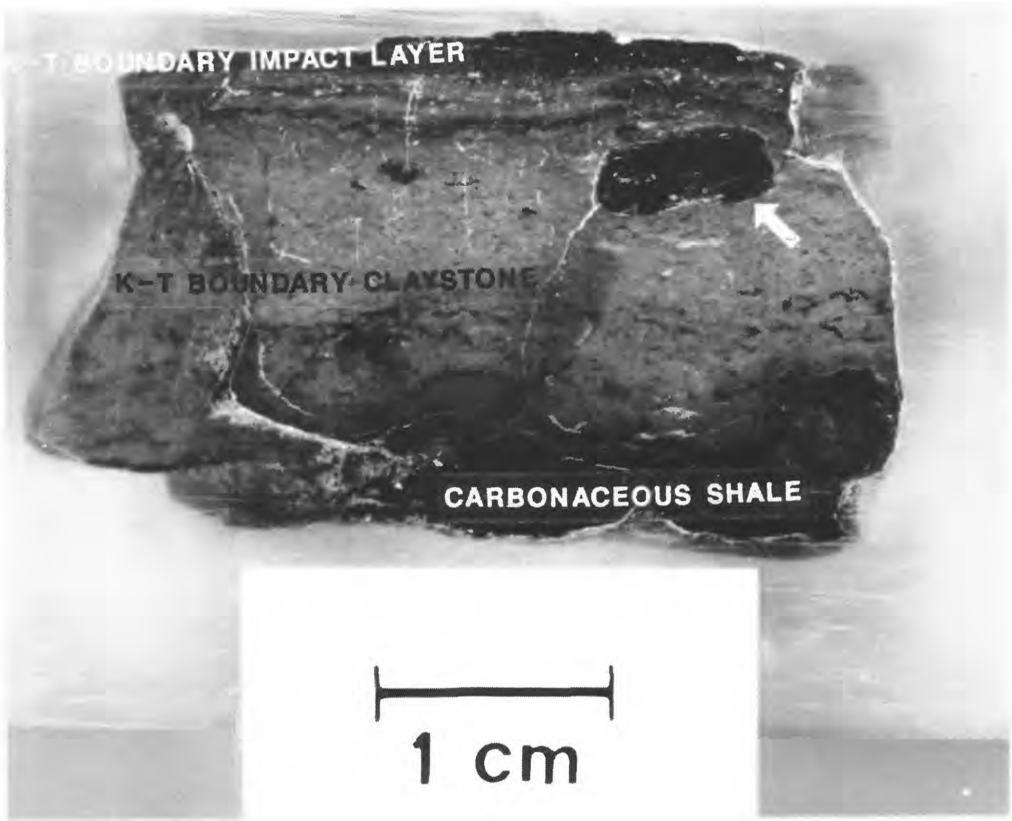
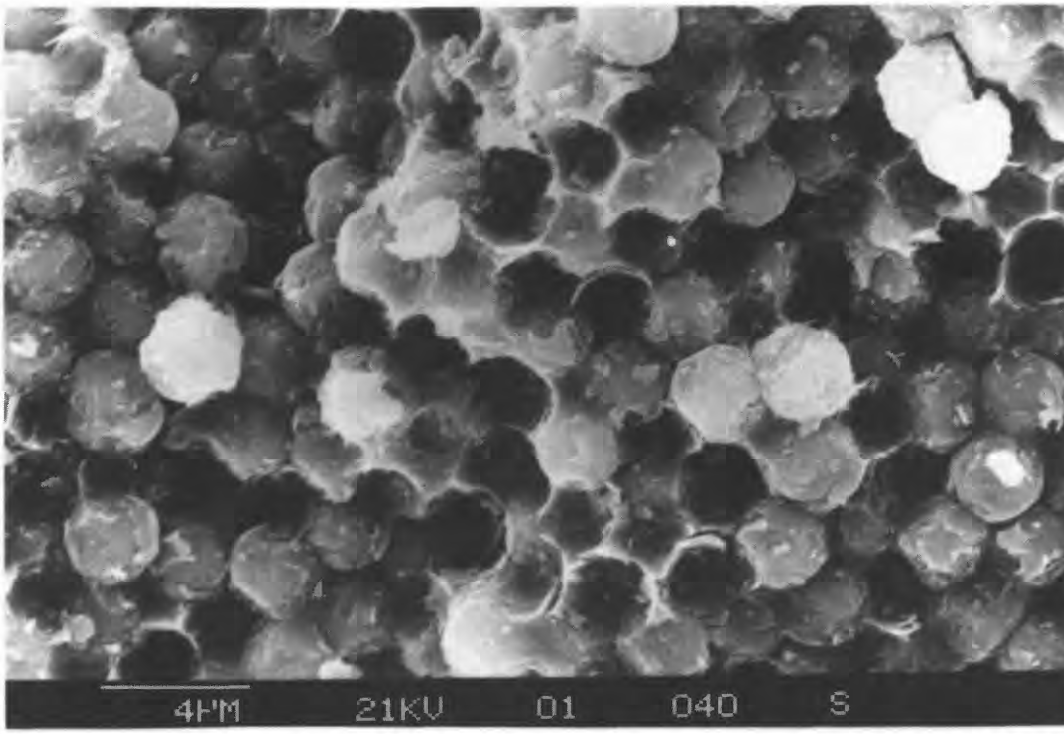


Figure 4.--Polished slab of the K-T boundary interval from the Madrid East, Colo., locality. The boundary interval consists of carbonaceous shale, the boundary claystone, and the impact layer. The boundary coal was lost during preparation. A rounded, coalified wood fragment shown at arrow. Bedding in boundary claystone warped over the fragment.

A)



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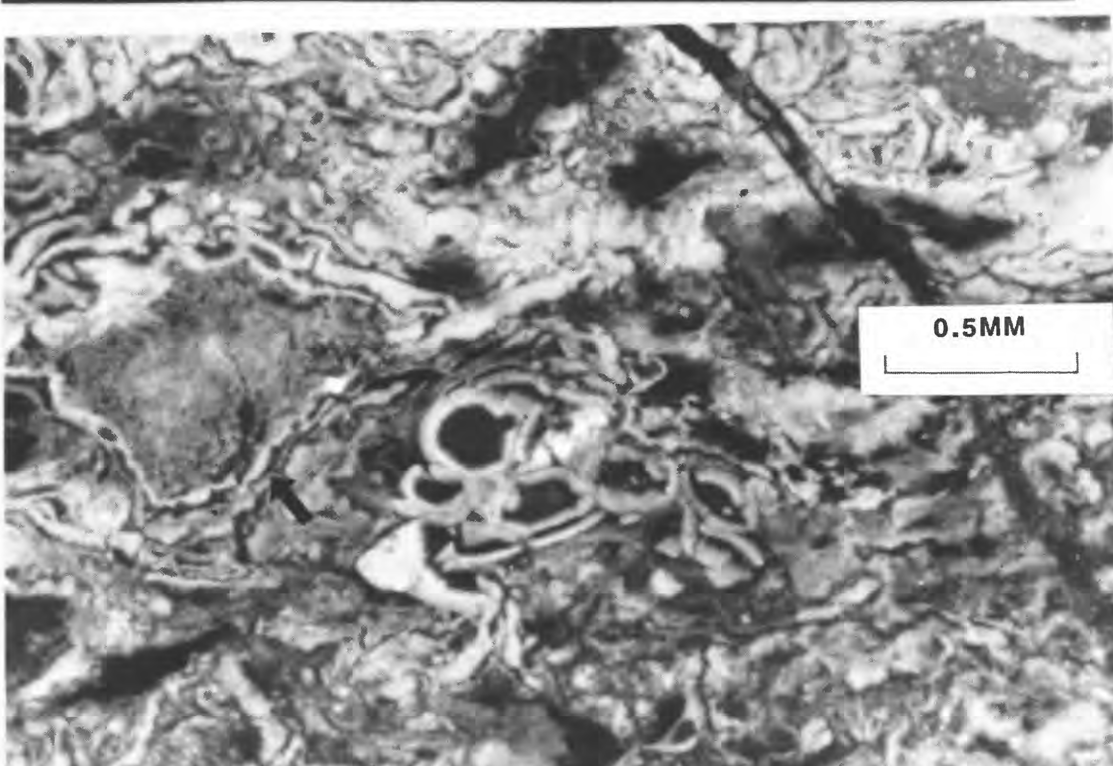


Figure 5.--Microscopic texture of the K-T boundary claystone. A, Texture as revealed by the SEM is a microspherulitic boxwork. B, Texture as seen optically with incident light is locally a compacted mass of crushed and uncrushed kaolinite spherules, some of which appear hollow. Arrow points to a clastic fragment of claystone that seemingly is rimmed by another generation of clay.

material and in cores of kaolinite spherules (Izett, 1987a), which indicates that it probably did not result from the in-situ alteration of glass.

Another textural type of claystone forms millimeter-sized layers and patches that consist of a compacted mass of kaolinite spherules and curved plate-like vermicules (fig. 5B). Single and groups of uncrushed kaolinite spherules occur near highly crushed masses, and this texture suggests that it formed during repeated episodes of spherule growth and compaction, possibly in a soil-forming environment. The curved kaolinite plates of crushed spherules, which form an integral part of the texture, typically contain a concentration along their convex sides of light-brown submicrometer-sized inclusions. The inclusions are rich in titanium as demonstrated using the X-ray fluorescence system of a scanning electron microscope. This texture was referred to as "imbricate" by Pollastro and Pillmore (1987) in reference to its similarity to the compacted fabric of welded tuffs.

The claystone has another textural feature that may have a bearing on its origin. This texture is fragmental, and it mainly occurs in the lower part of the claystone. Angular fragments of cryptocrystalline kaolinite nearly as large as 1.0 mm are set in the kaolinite matrix (fig. 5B). Some fragments have not been moved far because their margins can be matched with adjacent fragments. This fragmental texture is restricted to the boundary claystone and does not occur in the overlying impact layer, suggesting that the process responsible for the fragmentation of the boundary claystone occurred prior to the deposition time of the impact layer.

At most localities in the Raton Basin, the claystone is free of quartz grains and consists of macroscopically uniform material. At several places, however, it is impure and locally contains a large amount of unshocked quartz. At the Clear Creek South, Colo., locality (fig. 2), the claystone is contaminated with clastic mineral grains of quartz incorporated into the claystone from a directly underlying sandstone bed. At the south end of the outcrop, the claystone is as much as 5.0 cm thick and has been extensively reworked as evidenced by its content of unshocked mineral grains (as much as 20 percent). The sand grains are set in a wavy-bedded, clayey matrix composed mainly of kaolinite and I/S mixed-layer clay. At the middle of the outcrop, the boundary claystone is about 2.0 cm thick and consists of two distinct layers, apparently formed during two depositional stages (fig. 6).

The lower 1.5 cm of claystone formed during stage 1 and contains scattered quartz grains set in the claystone matrix. Among the thousands of quartz grains studied in oil immersion mounts were two to three that exhibited shock lamellae. In contrast, the impact layer that overlies the unit of reworked claystone contains a normal concentration of clastic mineral grains (as much as 2 percent), some of which contain shock lamellae. At such localities where the boundary claystone consists of reworked material, it typically and pervasively contains planar shreds of vitrinite preferentially oriented parallel to bedding.

The upper 0.5 cm of claystone deposited during stage 2 contains more disseminated planar shreds of vitrinite than the claystone deposited during stage 1, as shown by its dark-gray color (fig. 6). Study of oil immersion mounts of the clastic grains separated from the claystone deposited during stage 2 shows that shock-metamorphosed minerals are about five times as abundant as in the underlying claystone deposited during stage 1.

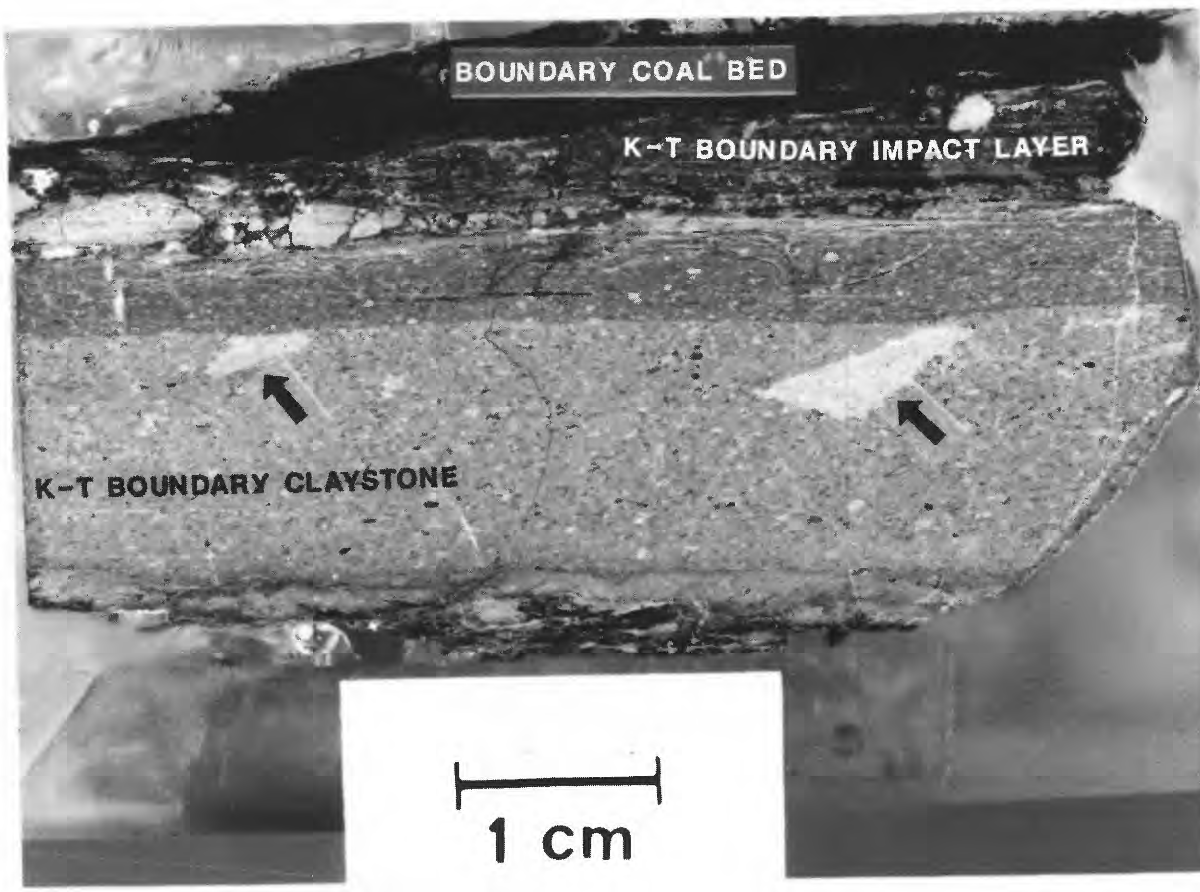


Figure 6.--Polished slab of the K-T boundary interval from the Clear Creek South, Colo., locality. The boundary claystone bed, impact layer, and boundary coal bed comprise the boundary interval. The boundary claystone consists of two subunits of reworked claystone that typically contain planar shreds of vitrinite and a considerable amount of unshocked quartz and feldspar grains. Rare quartz grains show multiple sets of shock lamellae. Two large angular clastic claystone fragments composed of unreworked boundary claystone shown at arrows. One clastic fragment is truncated by the overlying unit.

In addition to the clastic mineral grains, the claystone at the Clear Creek South, Colo., locality contains clastic angular claystone fragments (fig. 6), as large as 1.0 cm in length, that have a texture identical to that of the average (unreworked) boundary claystone. These fragments are truncated by an overlying erosion surface at the base of the overlying (stage 2) unit. Thus, the evidence indicates that the claystone fragments probably were derived by erosion of boundary claystone that had sufficient strength to survive erosion and transportation. The amount of time separating the formation of the claystone and its subsequent erosion and transportation is unknown, but it may have been appreciable. These relationships imply that the boundary claystone was deposited, lithified, and eroded prior to the arrival of shock-metamorphic minerals in the impact layer. At the Sugarite and Sugarite South, N. Mex., sites, the boundary claystone also was reworked, but because the subjacent sediment was coal-forming material, the claystone contains only rare quartz and feldspar grains.

At one place in the Raton Basin, the claystone appears macroscopically to be pure, homogeneous, and free of clastic mineral grains, but microscopic examination shows that the claystone may be slightly contaminated with clastic mineral grains that are concentrated along subvertical carbonaceous structures that are most probably paleoroots (fig. 1A). Such grains in these structures have only been seen at the Clear Creek North, Colo., locality. The grains are most common where clastic mineral grains in the overlying impact layer are most numerous. The grains were revealed by polishing the claystone with 6 micrometer diamond paste rubbed on the surface of the claystone with a soft tissue. Following polishing, the resistant grains stand in sharp relief to the surrounding soft claystone. The grains in the subvertical carbonaceous structures probably were introduced into the boundary claystone along roots of plants or possibly burrows or fractures following the depositional event that formed the overlying impact layer. Some of these grains display multiple sets of shock lamellae. Clastic mineral grains are restricted to the subvertical structures (probably paleoroots) and have not been seen in the horizontal carbonaceous laminae.

Rare shock-metamorphosed mineral grains were also identified in the hydrofluoric-acid-insoluble residue obtained from some pieces of carbonaceous shale, a few millimeters thick, that directly underlies the boundary claystone at the Clear Creek North, Colo., locality (fig. 1). However, at other localities the carbonaceous shale underlying the boundary claystone is free of shock-metamorphosed minerals. Such rare grains possibly were introduced along paleoroots. The rare shocked quartz grains were found by splitting pieces of carbonaceous shale from the base of the boundary claystone directly into a new plastic beaker. The shale was disaggregated and dissolved by ultrasonic scrubbing in a 5 percent solution of hydrofluoric acid. Using this method, the possibility of intralaboratory contamination was avoided.

In summary, the claystone that composes the K-T boundary claystone is generally free of clastic quartz and other mineral grains. However, rare clastic mineral grains were found in paleoroots at one locality in the Raton Basin. Locally, where the boundary claystone bed consists of reworked material, it contains as much as 20 percent clastic quartz and other mineral grains. At one locality in the Raton Basin, exceptionally rare clastic mineral grains occur in the carbonaceous claystone, a few millimeters thick, that underlies the boundary claystone.

Chemical Composition of the K-T Boundary Claystone

Chemical analyses of the boundary claystone were made by four different analytical methods to obtain as complete as possible a picture of its major- and trace-element compositional range: X-ray fluorescence (XRF), induction-coupled plasma atomic emission spectroscopy (ICP), instrumental neutron activation analysis (INAA), and radiochemical neutron activation analysis (RNAA). XRF analyses show that the major-element content of the claystone (about 95 percent SiO_2 and Al_2O_3 calculated on a water-free basis) closely resembles that of the clay mineral kaolinite. Small differences in the amounts of iron, magnesium, sodium, potassium, calcium, titanium, manganese, and phosphorus occur in claystone samples from place to place in the Raton Basin (table 1). The volatile content, mostly water, is about 15 percent. The amount of titanium, about 2.0 percent, calculated water-free as TiO_2, is slightly more than that of kaolinitic tonstein layers in coals of the Raton Basin.

Analyses of the boundary claystone from Brownie Butte, Mont. (Bohor and others, 1984) are listed in table 1. The claystone at the Montana locality is similar to the boundary claystone bed in the Raton Basin except that the Montana material contains more sodium and less potassium. This relation is compatible with X-ray diffraction analyses, which indicate that the claystone is more smectitic than the Raton Basin boundary claystone. The Montana claystone contains slightly less titanium than does the Raton Basin claystone.

Gilmore and others (1984) considered the amount of titanium in the boundary claystone to be an important tool for its identification. The titanium to aluminum ratio of 10 samples of the boundary claystone is 0.047±0.005 (1 sigma) as calculated from the data of table 1. Spears and Kanaris-Sotiriou (1979) assumed that this ratio remains constant during diagenesis of volcanic glass to clay. Huddle and Patterson (1961) pointed out that kaolinitic underclays below coal beds commonly contain from 0.5 to 1.5 weight percent TiO_2. The ratio of titanium to aluminum in the boundary claystone (0.047) is greater than that of most European tonsteins (0.004-0.04) considered by Spears and Kanaris-Sotiriou (1979, fig. 4) to be composed of altered silicic ash. Moreover, the titanium to aluminum ratio in the boundary claystone (0.047) is significantly less than that in European tonsteins considered by Spears and Kanaris-Sotiriou (1979, fig. 4) to be composed of altered basaltic ash (>0.16). If this is true, the boundary claystone is not altered silicic or basaltic volcanic ash because its titanium to aluminum ratio is markedly different than either of these compositional types. However, it is possible that the boundary claystone could have formed from intermediate (andesitic) composition volcanic glass.

The question of the relative mobility of major and trace elements during the diagenetic conversion of volcanic glass to clay was studied by Zielinski (1982). He compared the upper glassy part of a Miocene ash bed in the Troublesome Formation of Colorado with the lower part, which had been converted to clay (smectite). Relative to the glass of the ash, the smectite was enriched in magnesium, calcium, iron, titanium, manganese, phosphorous, and aluminum and depleted in potassium, sodium, and silicon during devitrification. The titanium to aluminum ratio remained nearly, but not exactly, constant. Among the trace elements, cesium, rubidium, uranium, and zirconium were depleted in the clay-altered ash (smectite) relative to the glass during diagenesis. Strontium, scandium, chromium, and cobalt were enriched in the smectite relative to the glass. In general, rare-earth elements were markedly leached from the glass during its conversion to clay.

Table 1.-- X-ray fluorescence analyses of the K-T boundary claystone bed from Colorado, New Mexico, and Montana.
[Analysts: A.J. Bartel, K.C. Stewart, and J.E. Taggart, U.S. Geological Survey; oxide elements in weight percent; Fe calculated as Fe_2O_3; LOI, lost on ignition at 900 ^{o}C. Localities of figure 2: SVS, Starkville South, Colo.; SVN, Starkville North, Colo.; CCN, Clear Creek North, Colo.; MADE, Madrid East, Colo.; RAT, Raton, N. Mex.; MR, Madrid Railroad, Colo.; BER, Berwind Canyon, Colo.; BB, Brownie Butte, Mont.]

	SVS	SVS	SVN	CCN	ME	RAT	MRR	BER	BB	BB
SiO_2	47.1	46.7	45.3	46.7	46.2	44.7	45.9	45.0	48.7	49.2
Al_2O_3	34.4	35.1	35.7	35.5	35.7	36.0	35.6	36.3	31.7	29.7
Fe_2O_3	0.62	0.67	1.23	0.68	0.60	1.61	1.37	1.17	0.88	1.20
MgO	.37	.37	0.53	.45	.37	0.44	0.29	0.40	.49	0.70
CaO	.18	.14	.30	.17	.20	.22	.12	.27	.06	.04
Na_2O	.17	<.15	<.15	<.15	<.15	<.15	.16	<0.15	.96	1.23
K_2O	.46	.44	.35	.28	.36	.25	.30	.27	.12	.17
TiO_2	1.76	1.86	1.80	1.68	1.92	1.64	1.59	1.91	1.37	1.65
P_2O_5	.06	.06	.11	.10	<.05	.11	.08	.10	<.05	<.05
MnO	<.02	<.02	<.02	<.02	<.02	<.02	<.02	<.02	<.02	<.02
LOI	15.0	14.9	14.9	14.5	15.1	15.4	15.1	15.0	15.0	15.7

Zielinski (1985) also compared chemical analyses of 979 silicic upper Cenozoic volcanic ash beds with analyses of a single Eocene kaolinitic tonstein at six localities in the Powder River Basin of Wyoming. He showed that, with the exception of aluminum and possibly titanium, the tonsteins were depleted during diagenesis of major rock-forming elements relative to the silicic ashes. Among the trace elements, rubidium and cesium were depleted, transition-series metals were moderately depleted, and gallium, zirconium, and hafnium were relatively immobile. Some elements, including thorium, tantalum, niobium, yttrium, and rare-earth elements (generally considered immobile during low-temperature diagenesis) were depleted.

Amounts of rare-earth elements, as determined by the INAA method, in the boundary claystone from the Raton Basin and Montana are listed in table 2. Their amounts were normalized to chondrites and plotted on figure 7. The plots show that the rare-earth-element content is low; lanthanum is only enriched 10 times and lutetium about twice relative to chondrites. Both patterns show a small negative europium anomaly. The shape of the rare-earth-element pattern of the boundary claystone from the Raton Basin is similar to that of clay in the K-T boundary interval at Stevns Klint, Denmark, and Gubbio, Italy, presented by Kastner and others (1984), except that the pattern of the boundary claystone from the Raton Basin and Montana have significantly lower chondrite normalized values (fig. 7). The rare-earth-element content of the boundary claystone from Western North American sites is an order of magnitude less than the average North American shale reported by Haskin and others (1968). However, the shape of the rare-earth-element pattern of the boundary claystones is similar to that of average North American shale and to the shapes of kaolinite (fig. 7) and smectite rare-earth-element patterns published by Cullers and others (1975).

Zielinski (1982) showed that rare-earth elements were, in general, leached from silicic volcanic ash during its alteration to smectite, but the general shape of the chondrite-normalized, rare-earth-element pattern remained similar to the parent glassy ash. In contrast, the chondrite-normalized, rare-earth-element patterns of some altered silicic volcanic ash beds are markedly different than those of normal silicic volcanic ash beds (G.A. Izett, unpublished data). Kendrick (1985) also showed that rare-earth elements are mobile during the alteration of volcanic glass to kaolinitic tonsteins.

Chemical analyses by the ICP method for 17 trace elements in the boundary claystone are listed in table 3. For comparison, analyses by the same analytical method of shale, claystone, coal, and jarosite from the K-T boundary interval are listed in table 4. The analyses (tables 3 and 4) show that the composition of the boundary claystone for the elements determined is similar to the "average shale" of the Earth's crust (Turekian and Wedepohl, 1961). The amounts of the rare-earth elements in the boundary claystone and carbonaceous claystone from one centimeter above the K-T impact layer are similar. Both have lanthanum contents only about 10 times chondrites. Kendrick (1985) showed that ash samples of coal also have very low rare-earth element contents, about 10-100 times chondrites.

Gilmore and others (1984) observed that the chemical composition of the boundary claystone is different from tonstein layers in coal beds of the Raton Basin, and that idea is confirmed by analyses in table 3 of this report. Gilmore and others (1984, p. 4) emphasized that the "enhanced" amount of chromium and scandium in the boundary claystone suggests that it is formed of altered basaltic ash. This conclusion may not be correct because the amounts of these elements are no different than that of average shale. Furthermore, the claystone lacks calcic plagioclase that commonly occurs in basaltic ash as

Table 2.--Instrumental neutron activation analyses of the K-T boundary claystone bed from Colorado, New Mexico, and Montana.
[Analysts: J.R. Budahn and R.J. Knight, U.S. Geological Survey; Fe in weight percent; other elements in parts per million; localities of figure 2: SVS, Starkville South, Colo., locality; BB, Hell Creek area, Brownie Butte, Mont.]

Element	SVS	BB
Cs	0.76	0.27
Rb	12.25	4.53
Sr	82.9	40.3
Th	3.75	1.39
U	2.01	2.60
La	3.35	3.30
Ce	7.74	5.04
Nd	1.76	1.75
Sm	.57	.48
Eu	.15	.10
Tb	.11	.08
Yb	.32	.34
Lu	.05	.06
Ta	.73	.68
Hf	2.83	1.77
Sb	3.54	1.27
Sc	19.8	14.4
Mn	8.41	26.32
Fe	.44	.64
Cr	59.4	43.6
Co	1.72	2.93

Table 3.--Induction-coupled atomic emission spectroscopic analyses of the K-T boundary claystone bed from Colorado, New Mexico, and Montana.
[Analyst: M.J. Malcolm, U.S. Geological Survey; Ti in weight percent; other elements in parts per million. Localities of figure 2: SVN, Starkville North, Colo.; SVS, Starkville South, Colo.; CCN, Clear Creek North, Colo.; RAT, Raton, N. Mex.; MAD, Madrid, Colo.; MADE, Madrid East, Colo.; BER, Berwind Canyon, Colo.; BB, Brownie Butte, Mont.]

	SVN	SVS	CCN	RAT	MAD	MADE	BER	BB
Ti	1.0	1.1	1.0	1.1	0.9	1.2	1.1	0.8
Mn	5	4	<17	15	5	<4	5	25
As	50	<10	20	120	40	40	20	50
Ba	210	150	180	370	650	320	260	120
Ce	78	17	62	80	81	7	76	10
Co	2	1	1	1	1	2	<1	2
Cr	65	70	51	49	39	52	68	53
Cu	12	3	6	19	7	11	69	9
Ga	30	31	30	32	31	35	32	<2
La	37	6	27	43	39	2	40	4
Li	220	280	240	240	250	200	180	200
Mo	7	8	9	12	6	13	6	2
Nd	38	5	26	28	40	<4	32	<4
Ni	4	3	4	4	<2	5	2	15
Sc	19	20	17	19	15	20	23	13
V	120	140	110	130	86	110	130	82
Zn	17	15	28	17	12	17	12	26

Table 4.--Induction-coupled atomic emission spectroscopic analyses of rocks from the K-T boundary interval from Colorado and New Mexico. [Analyst: M.J. Malcolm, U.S. Geological Survey; Ti in weight percent; other elements in parts per million. SHALE-1, millimeter-thick dark flaky shale directly overlying K-T boundary impact layer and underlying coal bed; SHALE-2, carbonaceous claystone about 1 cm above K-T boundary impact layer; CLY, claystone about 4 cm below K-T boundary claystone; COAL, K-T boundary coal bed; JAROSITE, jarosite laminae in K-T boundary impact layer; CRUST, analysis of average shale of Earth's crust reported by Turekian and Wedepohl (1961). Localities of figure 2: SVS, Starkville South, Colo.; BER, Berwind Canyon, Colo.; CCN, Clear Creek North, Colo.]

	SVS SHALE-1	SVS SHALE-2	BER CLY	BER COAL	CCN JAROSITE	CRUST AVG.SHALE
Ti	0.8	1.2	0.50	0.04	0.3	0.46
Mn	17	14	40	8	20	850
As	80	20	10	10	1500	13
Ba	380	360	380	47	150	580
Ce	29	9	82	19	62	60
Co	9	11	4	9	4	19
Cr	240	230	68	23	120	90
Cu	23	28	62	24	37	45
Ga	24	21	29	4	7	19
La	10	4	46	9	33	30
Li	25	21	35	2	13	66
Mo	6	4	<2	<4	<4	11
Nd	12	5	35	12	24	28
Ni	13	15	10	68	<2	68
Sc	28	37	21	16	10	13
V	220	180	170	90	360	130
Zn	58	61	42	78	45	95

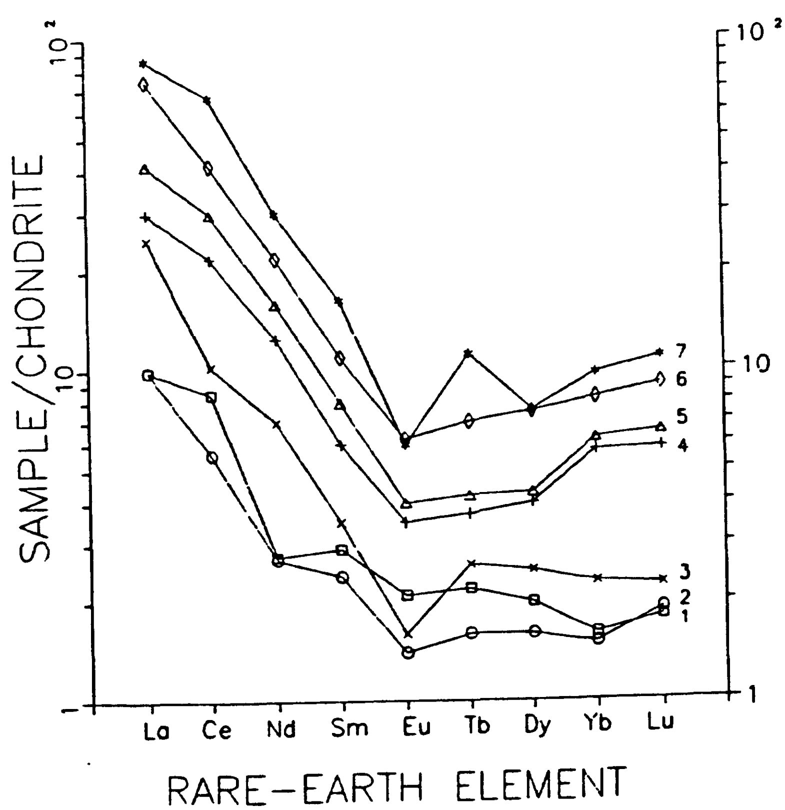


Figure 7.--Rare-earth-element diagram of K-T boundary interval rocks from North America and Europe. A typical kaolinite pattern from Georgia is also shown. Element concentrations normalized to amounts in chondritic meteorites plotted against REE atomic number. 1, Boundary claystone from the Starkville South, Colo.; 2, Boundary claystone at Brownie Butte, Mont.; 3, Boundary claystone residue from Stevns Klint, Denmark (Kastner and others, 1984); 4, Insoluble residue of Tertiary clay directly overlying K-T boundary at Stevns Klint, Denmark (Kastner and others, 1984); 5, Insoluble residue of Cretaceous clay directly underlying K-T boundary at Stevns Klint, Denmark (Kastner and others, 1984); 6, Insoluble residue from K-T boundary claystone at Gubbio, Italy (Kastner and others, 1984); 7, Macon, Georgia, kaolinite of Cullers and others (1975).

phenocrysts and microlites. The chemical composition of the boundary claystone may be slightly different than that of Cretaceous tonsteins of the Raton Basin, but that is not to say that it is different in its chemical composition (except platinum-group metals) from the average crustal shale.

Of interest is the low nickel (2-5 ppm) and cobalt (1-2 ppm) content of the K-T boundary claystone (table 3) in the Raton Basin. This is surprising because some (Pollastro and Pillmore, 1987) believe that the boundary claystone consists of altered impact ejecta, the asteroidal component of which should contain significant amounts of nickel and cobalt. In contrast to the low content of these elements (< 5ppm) in the boundary claystone in continental K-T boundary rocks of the Raton Basin, the K-T boundary impact layer in marine rocks at Caravaca, Spain, and Stevns Klint, Denmark, contains 2,580 (ppm) and 1,370 (ppm) nickel and 626 (ppm) and 146 (ppm) cobalt, respectively (Kyte and others, 1985).

The iridium, silver, and gold contents of the boundary claystone from four localities in the Raton Basin were determined by H.T. Millard, Jr., of the U.S. Geological Survey, by the RNAA method, and the data are listed in table 5 and illustrated on the stratigraphic sections of figure 8. In general, the iridium data are similar to those reported by Orth and others (1981), Pillmore and others (1984), and Gilmore and others (1984). However, the iridium data sets are not strictly comparable because the workers cited above did not recognize or sample the impact layer as a separate stratigraphic unit; accordingly, some of their samples may have been mixtures of the impact layer and the boundary claystone. The analyses of Millard clearly indicate that the boundary claystone contains only a small amount of iridium, generally less than 1.0 ppb. In contrast, the impact layer contains as much as 66 times more iridium than the boundary claystone (fig. 8).

Table 5.--Radiochemical neutron activation analyses for iridium, gold, and silver in the K-T boundary claystone and impact layer from Colorado and New Mexico.
[Analyst: H.T. Millard Jr., U.S. Geological Survey; elements in parts per billion; K-T, K-T boundary claystone bed; MAG, K-T boundary impact layer. Localities of figures 2 and 8: SVS, Starkville South, Colo.; CCN, Clear Creek North, Colo.; RAT, Raton, N. Mex.; MADE, Madrid East, Colo.]

	SVS K-T	CCN K-T	RAT K-T	MADE K-T	SVS MA	CCN MAG	RAT MAG	MADE MAG
Ir	0.27	0.22	0.07	0.32	1.21	14.6	1.2	7.5
Au	1.1	2.0	4.4	5.1	1.7	7.3	2.5	9.4
Ag	31	74	100	56	50	280	63	70

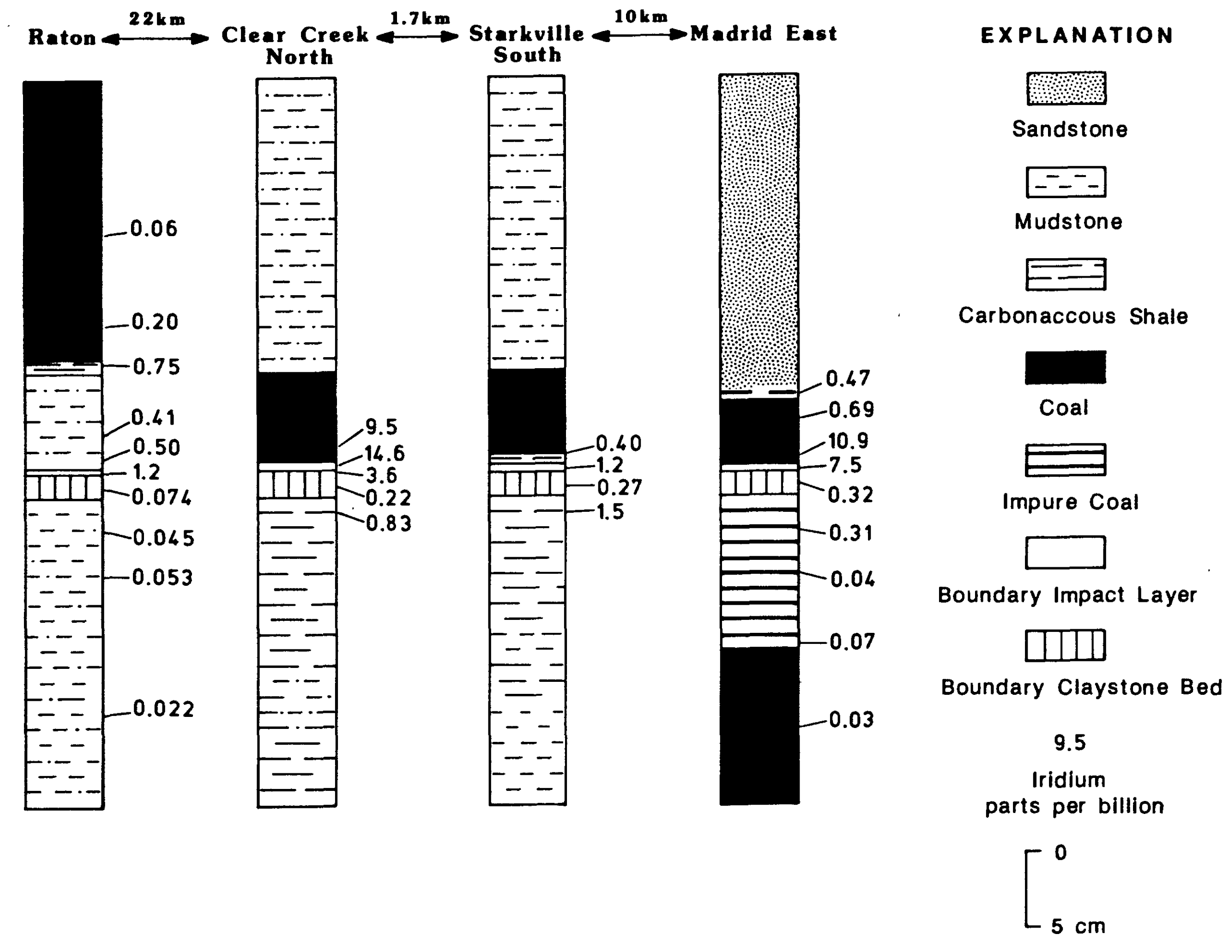


Figure 8.--Stratigraphic diagram showing amount of iridium (ppb) in K-T boundary rocks of the Raton basin. Locations of stratigraphic sections shown on figure 2. Iridium analyses made by H.T. Millard, Jr., of the U.S. Geological Survey following methods he described in 1987.

Origin of the K-T Boundary Claystone

Although the boundary claystone is a useful geological mapping tool and stratigraphic guide for identifying and locating the K-T lithologic and paleontologic boundary, its origin is controversial and a paradox. On the one hand, the claystone has many similarities to tonsteins and cryptotonsteins in coal beds, but the origin of these, either volcanic, biogenic, or diagenetic, has been debated for years (Williamson, 1970). Some tonstein layers are surely of volcanic origin, but not all can be ascribed to this origin. If an iridium anomaly and shock-metamorphosed minerals had not been found associated with the boundary claystone, a volcanic origin for it would be assigned without question by many geologists. Pollastro and Pillmore (1987) emphasized that the boundary claystone (1) forms a thin, discrete unit compatible with the type of unit they visualized for a distal impact ejecta fallout layer, (2) constitutes a widespread stratigraphic marker that occurs from New Mexico to near Edmonton, Canada (Bohor and Izett, 1986), and (3) contains above-background amounts of platinum-group elements such as iridium. On the other hand, information gathered during this study, including stratigraphic relations, lithologic observations, petrographic evidence, and chemical data, suggests that the boundary claystone may not be of impact origin.

Alvarez and others (1980) proposed that the boundary claystone in marine rocks of Europe consists of "pulverized rock" of impact origin. If the boundary claystone were originally composed of pulverized shock-metamorphosed rock, its component clastic minerals should have survived diagenesis. This conclusion is based on the fact that quartz and feldspar in the boundary claystone and impact layer are not altered. Accordingly, the boundary claystone should contain considerably more quartz and feldspar than it does.

Alvarez (1983) speculated that the marine boundary claystone bed is of volcanic origin. Gilmore and others (1984) and Pollastro and Pillmore (1987) suggested that the claystone in the Raton Basin formed by alteration of glassy impact ejecta. Pillmore and others (1984) suggested that the boundary claystone in the Raton Basin is of impact or possibly a combination of volcanic and impact origin.

Information supporting the contention that the boundary claystone is fallout material from either a volcanic or impact event is based mainly on two observations, its geometry and texture. Concerning the first observation, the geometry of the bed, Gilmore and others (1984, p. 228) maintained that "The uniformity in thickness of the K-T kaolinitic clay bed over a distance of 1,200 km indicates that the fallout material came from a common source at a considerable distance." Their argument is persuasive, but the geometry of the bed does not conclusively prove that it formed from air-fall material of either impact or volcanic origin. Moreover, it should be emphasized that the boundary claystone may not form a continuous blanket-like layer in K-T boundary rocks of Western North America. The boundary claystone has been found only at spot localities in a fine-grained sequence of rocks deposited in coal swamps under similar environmental conditions.

Concerning the second observation, Pollastro and Pillmore (1987) referred to the texture of the claystone as "imbricate" and claimed it was inherited from a very coarse, fragmental impact fallout material consisting of vesicular glass shards. Although, superficially, the texture resembles that of pyroclastic rocks, it lacks angular shard shapes (see fig. 5B) and, more important, lacks broken crystals typical of these rocks (Ross and Smith, 1961, figs. 16-23).

Alvarez (1983) and Pillmore and others (1984) proposed that the boundary claystone formed, at least in part, from the alteration of volcanic glass shards. If this were true, it should contain components of an assemblage of primary volcanic minerals typical of voluminous, widely dispersed, volcanic ash beds. On the contrary, it does not. The claystone is notable for its general lack of quartz and feldspar grains. Studies of numerous volcanic ash beds have shown that most regional silicic ash beds contain, as a minimum, either quartz, zircon, apatite, or ilmenite phenocrysts of uniform chemical composition (Izett, 1981). It would be a rare ash that did not contain one of these stable mineral species as a primary phase. Surely, some of these minerals would survive diagenesis of ash to kaolinite (Ryer and others, 1980; Spears and Rice, 1973). The trace-element chemical composition of the claystone is also incompatible with the idea that it consists of altered silicic volcanic ash, as pointed out by Gilmore and others (1984). Moreover, the low TiO_2 to Al_2O_3 ratio (0.047) of numerous analyzed samples of the claystone (table 1) indicates that, if the precursor material were volcanic glass, it could not have been of basaltic composition (see Spears and Kanaris-Sotiriou, 1979, fig. 4), which is the only type of volcanic material that possibly would contain iridium in amounts comparable to those in K-T boundary rocks.

Information supporting the contention that the boundary claystone is not fallout impact or volcanic material consists of stratigraphic relations, lithologic observations, and chemical data. An important mineralogic-stratigraphic relation was found during my study of Raton Basin K-T boundary sediments. Shock-metamorphosed materials appear abruptly, and are concentrated (fig. 9), in the thin layer (herein called the K-T boundary impact layer) that rests sharply on the K-T boundary claystone bed. The facts that shocked minerals are (1) concentrated in the impact layer and (2) generally absent in the boundary claystone suggest that the boundary claystone may not be composed of impact material. If it were, it is difficult to envision a single-event, air-fall-depositional model that would allow large shock-metamorphosed mineral grains to be concentrated in the upper rather than in the lower unit of the boundary interval couplet. It is also difficult to understand how the boundary claystone could originate from an impact glass free of shock-metamorphosed minerals. Stöffler (1984, p. 466) and Grieve (1987, p. 253) pointed out that an important feature of impact glasses is their content of shocked mineral grains and lithic clasts.

A second stratigraphic relation pointing to a non-impact origin for the boundary claystone is its apparent absence in marine K-T boundary rocks in Europe. At nonmarine depositional sites in Western North America, the impact layer records the abrupt appearance of shock-metamorphosed minerals. At marine depositional sites, such as Caravaca, Spain, and Stevns Klint, Denmark, shock-metamorphosed minerals first appear in a 1- to 3-mm-thick clay unit that marks the K-T boundary (Izett, 1987a). This clay unit also contains an iridium concentration spike and marks a significant mass extinction event (Smit and Hertogen, 1980). There is no equivalent of the Western North American K-T boundary claystone below the impact layer at Caravaca, Spain, or Stevns Klint, Denmark (fig. 10). Thus, the impact layer of Western North America and Europe records the first arrival of shock-metamorphosed minerals in the boundary interval. Then, seemingly, the boundary claystone, which underlies the impact layer in Western North America but not in Europe, may not be impact related. For if it were, one would expect to find its equivalent in European K-T boundary sections.

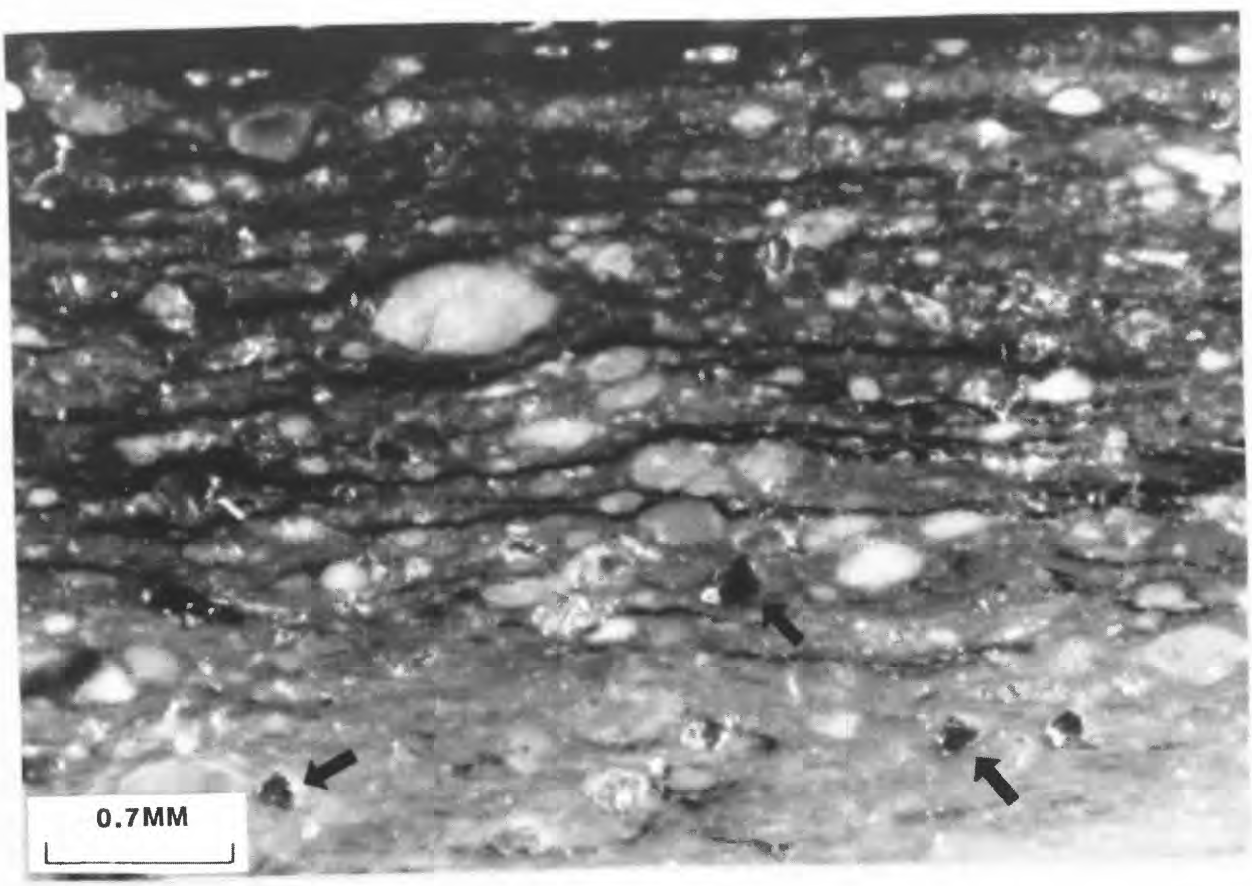


Figure 9.--Photomicrograph of the K-T boundary impact layer at the Berwind Canyon, Colo., locality. Impact layer typically contains laminae of black vitrinite and white almond-shaped kaolinite pellets. Scattered small clastic quartz grains shown at arrows. Some quartz grains contain shock lamellae. Photographed with incident light.

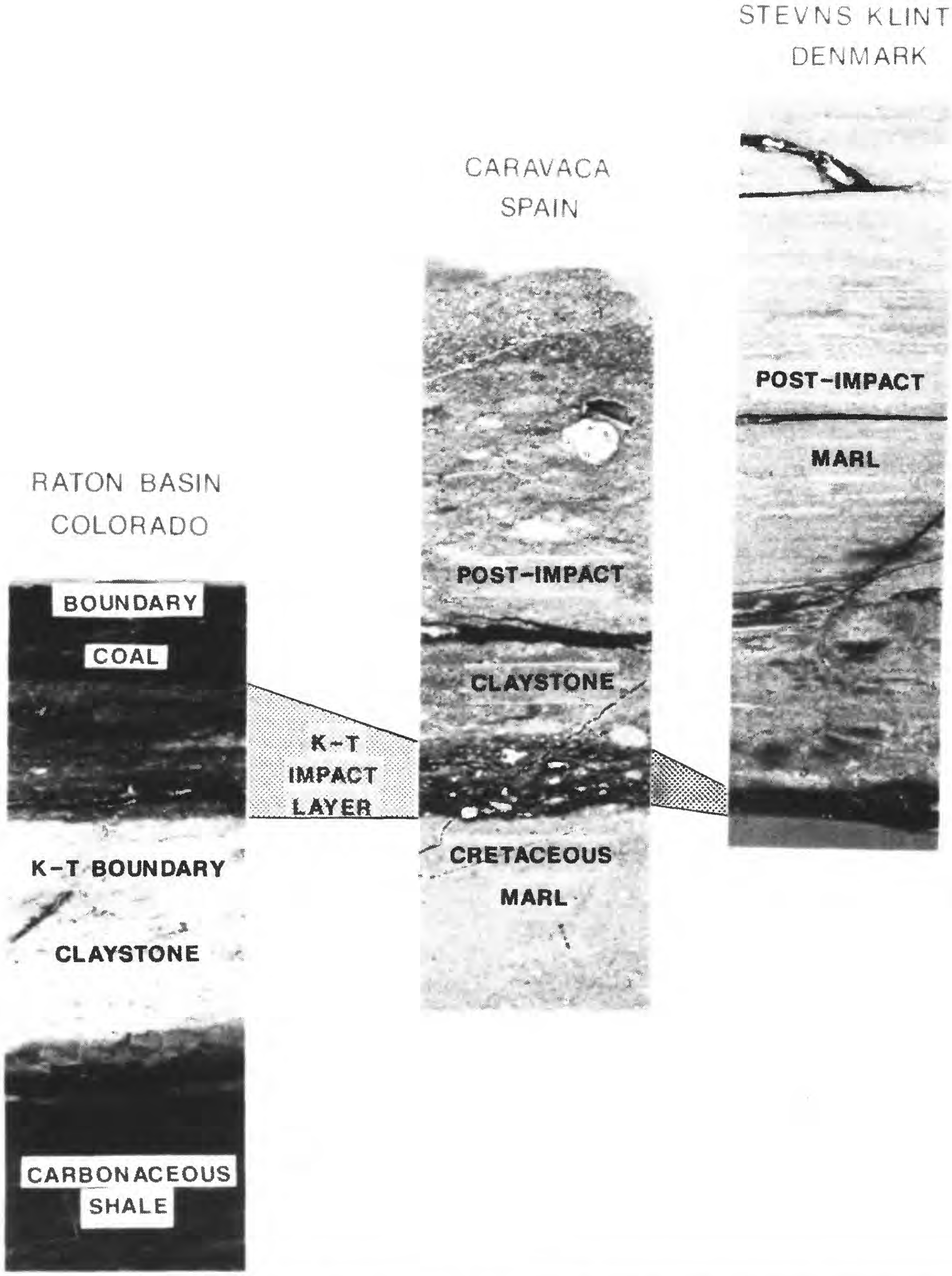


Figure 10.--Comparison of K-T boundary interval rocks deposited in continental and marine environments from North America and Europe. Polished slab of continental K-T boundary rocks from the Clear Creek North, Colo., locality, Raton Basin. Impact layer is 6.0 mm thick. K-T boundary interval marine rocks from the Caravaca, Spain, locality. Impact layer, 1.0 mm thick, is underlain by Cretaceous marl. K-T boundary interval marine rocks from the Stevns Klint, Denmark, locality. Impact layer is 1.0 mm thick. Marl underlies the impact layer, but it was lost during preparation of sample. The K-T boundary impact layer records the abrupt appearance of shock-metamorphosed minerals and an iridium abundance anomaly. No equivalent of the K-T boundary claystone bed of the Western Interior North America occurs at marine K-T boundary sites in Europe.

A third stratigraphic relation bears on the origin of the K-T boundary claystone. At most localities in the Raton Basin, the boundary claystone is sharply overlain by the impact layer (fig. 1). The contact between these units separates dissimilar lithologies and marks an abrupt change in depositional regime. At some places, however, the contact appears gradational, but only within a narrow interval (1 mm). Within this 1-mm-thick interval the claystone below appears billowy and seems to have been degraded into proto-pellets (fig. 11). Thus, the evidence seems to suggest that the top of the boundary claystone was a paleosurface upon which plants grew (figs. 1 and 3) and the impact layer was deposited.

Lithologic information bears on the origin of the claystone. The contact between the boundary claystone and the impact layer separates relatively homogeneous claystone containing swirls of vitrinite from overlying laminated claystone (fig. 1). This observation suggests that a pre-impact deformational process locally contorted presumably horizontal vitrinite laminae in the boundary claystone prior to the deposition of the impact layer. Furthermore, at the Clear Creek North, Colo., locality, the impact layer rests unconformably on carbonaceous shale that contains lenses of the boundary claystone (fig. 12), indicating that the rocks below the contact have been reworked prior to deposition of the impact layer.

Other lithologic information suggests that the boundary claystone bed accumulated slowly by normal sedimentary processes rather than rapidly by fallout of impact material. Its content of centimeter-long, horizontal vitrinite laminae, blebs of amber, degraded plant material, fossil twigs, fossil roots, angular and rounded claystone fragments, rounded fragments of woody material (fig. 4), and carbonaceous structures that flare out at the contact with the impact layer (fig. 3B) all argue for a complex origin for the boundary claystone.

Moreover, there is evidence that the K-T boundary claystone could not be altered tektite glass. A fact not appreciated by most geologists is that tektite glass is markedly different than ordinary volcanic glass. Tektite glass is essentially anhydrous and has a relatively stable structure, and thus it does not hydrate. Adams (1984, p. 202) stated that "The extremely low water content of tektites makes them virtually impregnable to hydration." For example, tektites and microtektites of upper Eocene age have survived in pristine condition for about 34 Ma in sediments of DSDP site 612 on the continental slope off New Jersey (Glass, 1987; Thein, 1987). It is entirely possible that tektite glass formed during the K-T event would still be glass, and would not have altered to kaolinite as some (Pillmore and others, 1984; Pollastro and Pillmore 1987) believe. In contrast, volcanic glasses contain a relatively large amount (a few tenths of one percent) of volatiles (see Ross and Smith, 1955). Because tektite glass does not hydrate under ambient conditions, it does not alter to clay or zeolites as does volcanic glass. Instead, in an aqueous environment, tektite glass dissolves. Typically, tektites have a corroded, rough surface.

If the claystone bed were altered impact ejecta, it should contain a major iridium anomaly. It quite simply does not. Data of Gilmore and others (1984, p. 226), Pillmore and others (1984), and Pillmore and Flores (1987) and data of this report illustrate the fact that amounts of iridium are generally highest in the impact layer and adjacent carbonaceous-rich rocks and lowest in the boundary claystone. The exact origin of the claystone remains a problem, but some of its microscopic textural features and chemical features suggest that it has a complicated diagenetic history. It is by definition a tonstein, but one that is different texturally and slightly different chemically from other tonsteins in the Raton Basin.

A)

B)

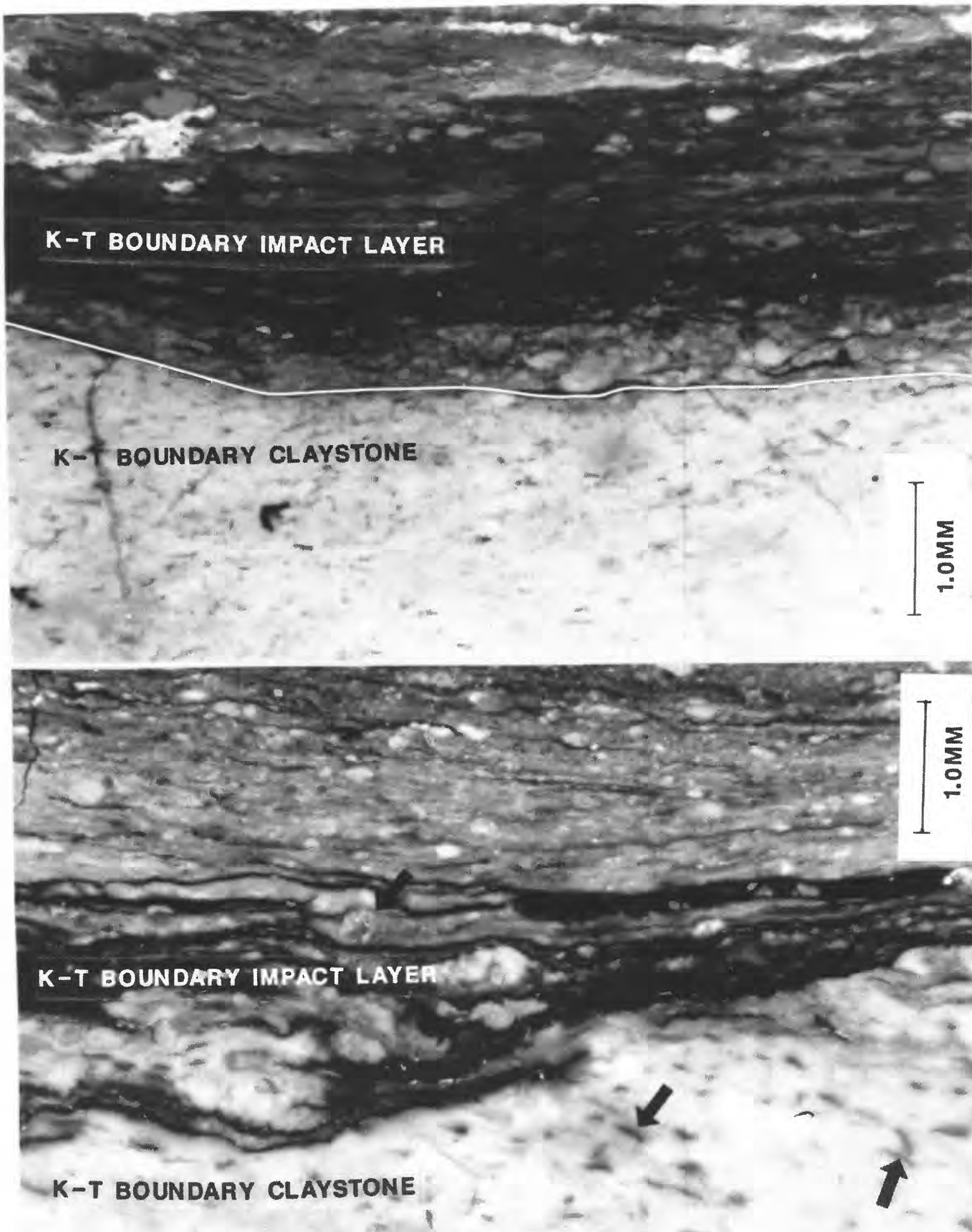


Figure 11.--Photographs showing the contact between the K-T boundary claystone and impact layer. A, Billowy nature of the uppermost part of the boundary claystone in specimen from the Clear Creek North, Colo., locality. B, Dark patches of I/S mixed-layer clay in boundary claystone and large shock-metamorphosed quartz grain (shown at arrows) in impact layer in specimen from the Berwind Canyon, Colo., locality.

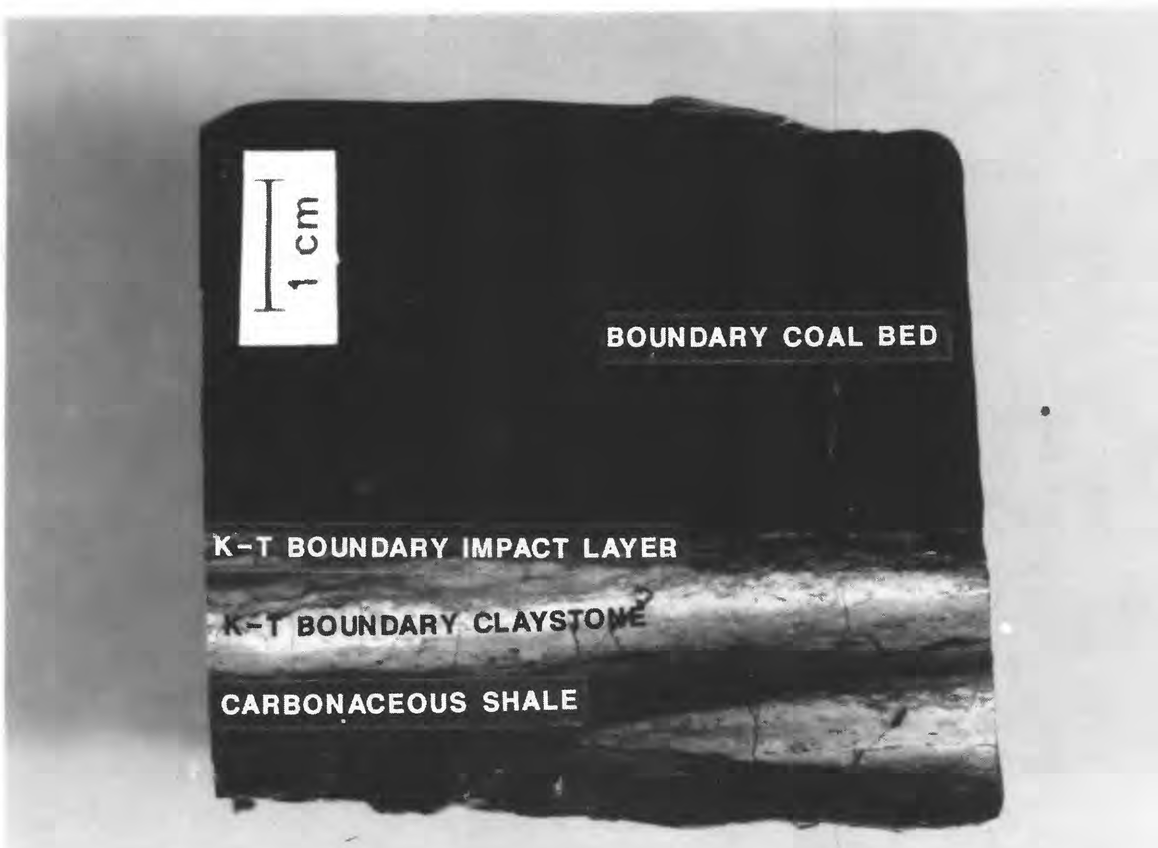


Figure 12.--Photograph of a polished surface of the K-T boundary interval from the Clear Creek North, Colo., locality. Microstratigraphic units shown, in ascending order, are (1) carbonaceous shale containing a lens of boundary claystone lithology overlain by another layer of boundary claystone lithology, (2) the K-T boundary impact layer, and (3) the K-T boundary coal bed.

In conclusion, the contention that the boundary claystone of the Western Interior of North America is of impact origin is based mainly on two observations: (1) it forms a thin, widespread bed in continental sedimentary rocks, and (2) it forms a stratigraphic couplet with the overlying K-T boundary impact layer at nearly all K-T boundary localities that contain anomalous amounts of iridium. The first observation, that the boundary claystone is a thin, widespread unit, may have been misinterpreted by those who favor an impact origin. It is true that the boundary claystone is a thin, widespread unit, but that is not to say that it formed as a continuous blanket-like layer. It is important to remember that the thin beds that make up the boundary interval (boundary claystone, impact layer, and boundary coal bed) in the Western Interior of North America have been found at only about 25 localities, and the rocks at each of these localities formed in low-energy, coal-swamp environments that existed in a series of north-trending Rocky Mountain coal basins. The boundary claystone has not been found outside of these coal basins.

The second observation that suggests that the boundary claystone is of impact origin is based on the stratigraphic association at nearly all localities of the boundary claystone and impact layer. This may be considered an argument for a genetic relationship between the two units. However, it should be remembered that the impact layer is overlain at most localities in Western North American K-T boundary sites by a thin coal bed that certainly is not of fallout, impact origin.

If the K-T boundary claystone is not of impact origin, as suggested by some observational evidence, reasonable questions are what is its origin and how could such a thin bed of claystone form over such a broad area by normal sedimentary processes? The answers to these questions are not presently known, but the boundary claystone has several features suggesting that its upper boundary is a paleosurface and that the claystone underwent pedogenic processes. The claystone should be studied by soil scientists familiar with microscopic characteristics of pedogenic claystone formed in tropical soil-forming environments (see Retallack, 1976). To form a claystone over such a broad area would require amazingly uniform depositional conditions, and indeed such conditions may have existed during the Cretaceous. Kauffman (1987, p. 150) pointed out that climatic conditions at the close of the Cretaceous were much more equable than at present and characterized by lack of polar ice, low seasonality, and broad tropical-temperate climate zones.

The most important observational fact that bears on the origin of the K-T boundary claystone is that shock-metamorphosed minerals are restricted essentially to the K-T boundary impact layer and are not found in the underlying K-T boundary claystone. As pointed out previously, it is difficult to understand how the boundary claystone could originate from impact material that would be entirely free of clastic mineral grains, in particular, shock-metamorphosed minerals.

AUTHIGENIC SPHERULES

In the Raton Basin, the K-T boundary claystone contains four compositional types of authigenic spherules including (1) kaolinite, (2) goyazite, (3) composite spherules consisting of a goyazite shell and kaolinite core (fig. 13A), and (4) jarosite. The spherules were studied on the surface of polished slabs, in thin sections, using a scanning electron microscope, and in oil immersion mounts. Because the spherules occur in the boundary claystone, which is thought by some to be of impact origin, they could be interpreted (Pollastro and others, 1983) as altered impact melt droplets analogous to microtektites. However, I will present evidence that indicates that the spherules are authigenic and not of impact origin.

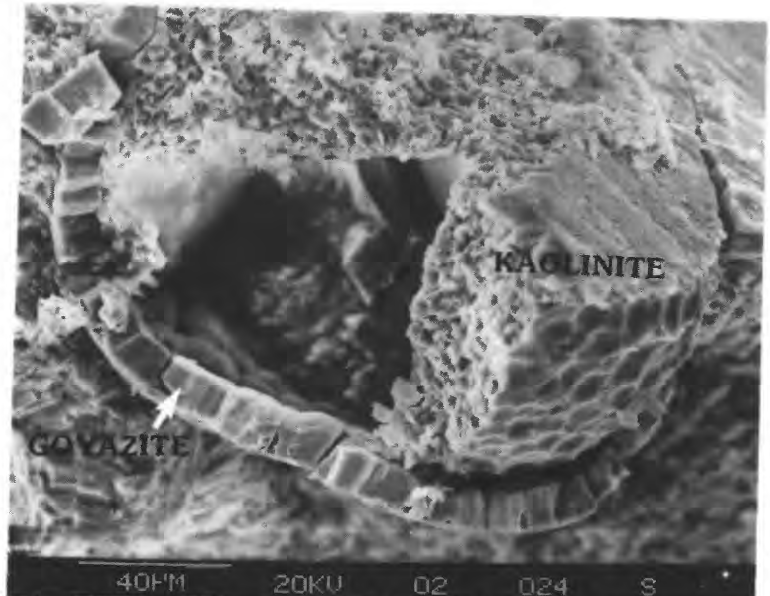
In the Raton Basin, kaolinite spherules are generally solid and range from about 0.1 to 0.2 mm in diameter. They consist of a kaolinite core encased by a thin shell of kaolinite having a different habit than the core. The shell, 5-10 micrometers thick, is composed of columnar kaolinite containing in its outermost region a concentration of minute brown inclusions having significantly more titanium and barium than the kaolinite of the core. The surface of the shells of the spherules is finely pitted reflecting the microspherulitic texture of the surrounding kaolinite host. The texture of the cores of the spherules, as seen microscopically, is a mosaic of kaolinite granules, a few micrometers in diameter. The SEM shows that the texture is a polygonal boxwork filled with kaolinite spherules 2-4 microns in diameter, similar to that of the kaolinitic matrix of boundary claystone.

The second compositional type of spherule in the boundary claystone is composed of a goyazite-like mineral (herein referred to as goyazite) of the plumbogummite series. Goyazite is an alumino-phosphate that contains small amounts of strontium, cerium, calcium, barium, and lanthanum. This alumino-phosphate mineral has been reported to occur in kaolinitic tonstein layers (Triplehorn and Bohor, 1983). Price and Duff (1969) pointed out that this mineral occurs in certain tropical soils, particularly laterites. The outermost part of the goyazite shell is consistently enriched in barium. In the Raton Basin, these spherules are scattered randomly through the boundary claystone and are not concentrated at its base.

Most spherules are light brown, opaque, and waxy appearing, and others are colorless to light reddish brown and translucent. They range in diameter from about 50 to 150 micrometers, but most are about 80 micrometers in diameter. As pointed out by Izett (1987a, fig. 9), many are spherical to subspherical; however, some were seen that are egg shaped and others are compound and form intergrown groups and mammillary masses. Most are hollow and have thick rims as pointed out by Izett (1987a); a few are solid and consist of radially oriented sheaf-like crystal masses. SEM study of the goyazite spherules shows that many are smooth, but some have a finely pitted surface reflecting the microspherulitic texture of the surrounding kaolinite host. The internal surface of the shell of hollow spherules is commonly colloform. The exterior surface of some spherules shows impressions of another unidentifiable coauthigenic mineral, possibly pyrite or goyazite.

The third compositional type of spherule, as large as 0.2 mm, is composed of a microcrystalline core of kaolinite surrounded by a 10-micrometer-thick shell of goyazite (fig. 13A). The interior surface of the goyazite shell is mammillary, and growth impressions of the goyazite are on the outer surface of the kaolinite core. This geometry suggests that the goyazite formed last and grew leaving a mammillary imprint on the pre-existing kaolinite groundmass. If the goyazite shell had formed first, it seemingly would not be possible for the microspherulitic kaolinite to fill the spherical void.

A)



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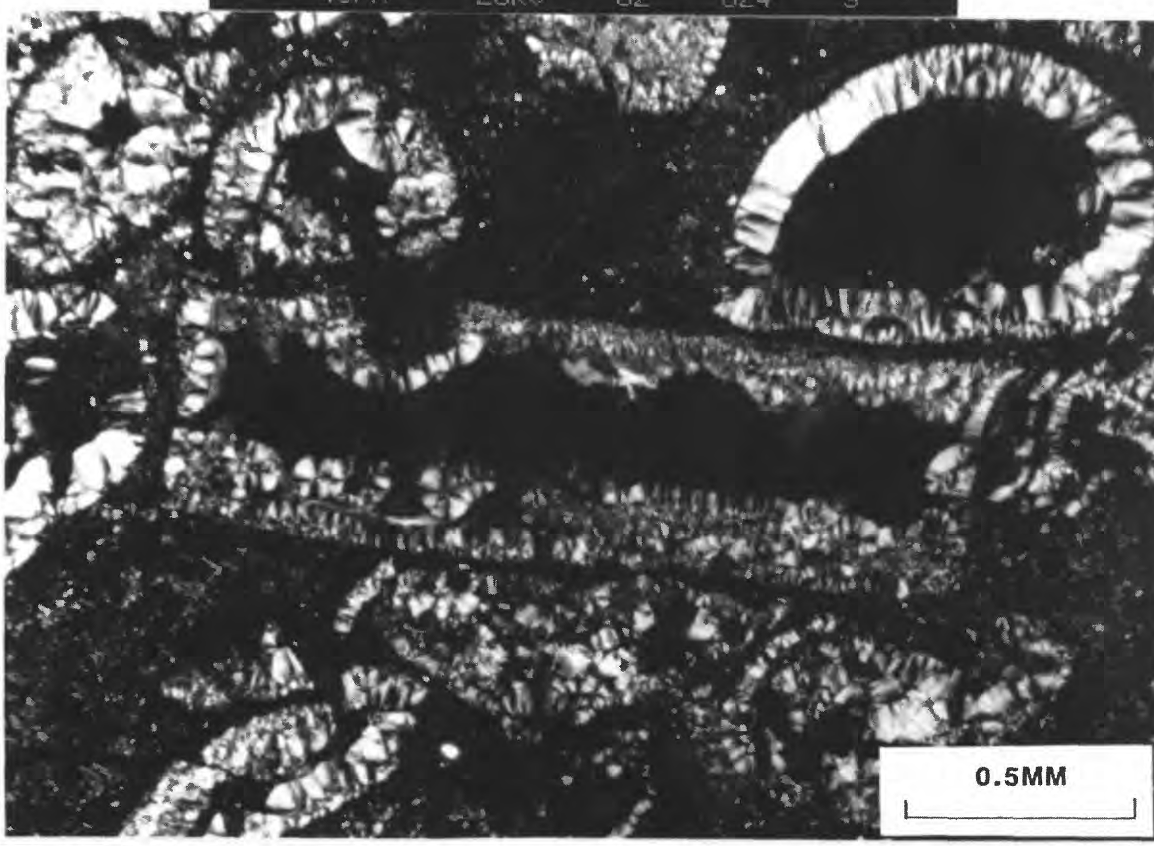


Figure 13.--Goyazite and goyazite-kaolinite "spherules" from the K-T boundary claystone bed. A, SEM photomicrograph of "spherule" from the Madrid, Colo., locality. Shell of goyazite surrounds core of microspherulitic kaolinite. Outer surface of the kaolinite core has growth impressions of the mammillary inner surface of the goyazite shell. B, Photomicrograph of thin section of the boundary claystone from Dogie Creek, Wyo. (Bohor and others, 1987b). Goyazite occurs in crushed masses, deformed spherules, and a rectangular cavity filling.

The fourth compositional type of spherule found in the claystone consists of authigenic, drusy, and spherical masses 10-50 micrometers in diameter composed of jarosite that probably formed from framboidal pyrite. The jarosite spherules occur singly or in planar lenticular spheroidal clusters in or near concentrations of carbonaceous material.

The abundance of the spherules in the Raton Basin is difficult to assess owing to their small numbers, but they comprise far less than 0.1 percent of the claystone. Goyazite spherules are relatively abundant at the Raton, N. Mex., Madrid, Colo., and Berwind Canyon, Colo., localities (fig. 2); in contrast, they are extremely rare at the Clear Creek North, Colo., locality. Kaolinite spherules are relatively abundant at the Clear Creek North, Colo., locality, but rare at the Berwind Canyon, Colo., locality (fig. 2). It is important to stress the fact that all four types of spherules can occur in kaolinitic claystone within a few millimeters of each other.

In contrast to the paucity of spherules in the K-T boundary claystone in the Raton Basin, hollow and solid goyazite spherules are abundant in the boundary claystone at two localities in Wyoming, along Dogie Creek in the Lance Creek area (Bohor and others, 1987b) and on the Allemand Ranch near Teapot Dome (Wolfe and Izett, 1987). At the Brownie Butte, Mont., locality 500 km north of the Lance Creek area, I have not seen any goyazite spherules in thin sections or polished slabs of the boundary claystone. Spherules also have not been found in the boundary claystone at sites in Alberta and Saskatchewan, Canada. The origin of the goyazite spherules is controversial, and because of their abundance, considerable time was spent in studying specimens from the Wyoming K-T boundary localities.

At the Dogie Creek and Teapot Dome, Wyo., K-T boundary localities, dark reddish-brown to colorless goyazite spherules are concentrated in a discrete, planar claystone layer as much as 1.0 cm thick that occurs at the top of the K-T boundary claystone (fig. 14). In addition, goyazite spherules occur in centimeter-long lenses within the claystone at the Dogie Creek site. This remarkable distribution of spherules within a discrete stratigraphic layer is markedly different than the distribution of spherules in the K-T boundary claystone in the Raton Basin. There the spherules are widely disseminated in the boundary claystone (Izett, 1987b). Goyazite spherules constitute about 30 percent of the 1-cm-thick goyazite-enriched layer at the Dogie Creek site, whereas at the Teapot Dome site, they form about 10 percent of the claystone. Locally, the goyazite-enriched layer contains kaolinite pellets that are similar to the normal kaolinite of the boundary claystone (fig. 14). Some of these pellets are penetrated by veins of goyazite, and some contain rare, small, goyazite spherules. Locally, the goyazite spherule layer abruptly terminates against kaolinite pellets.

The diameter of goyazite spherules at the Teapot Dome and Dogie Creek localities ranges from about 40 to 1,200 micrometers (0.04-1.2 mm). Most of the largest spherules are uniformly about 0.7-1.0 mm in diameter, and thus, the size distribution is markedly truncated at about 1.0 mm. Measurement of goyazite spherule diameters in oil immersion mounts and thin sections demonstrates that the largest goyazite spherules are generally hollow and thin walled (or filled with gypsum), and the smallest ones are thick walled or solid. Of considerable interest is the fact that the largest goyazite spherules at Wyoming K-T boundary sites are considerably larger (about 10 times) than the largest goyazite spherules at boundary sites in the Raton Basin.

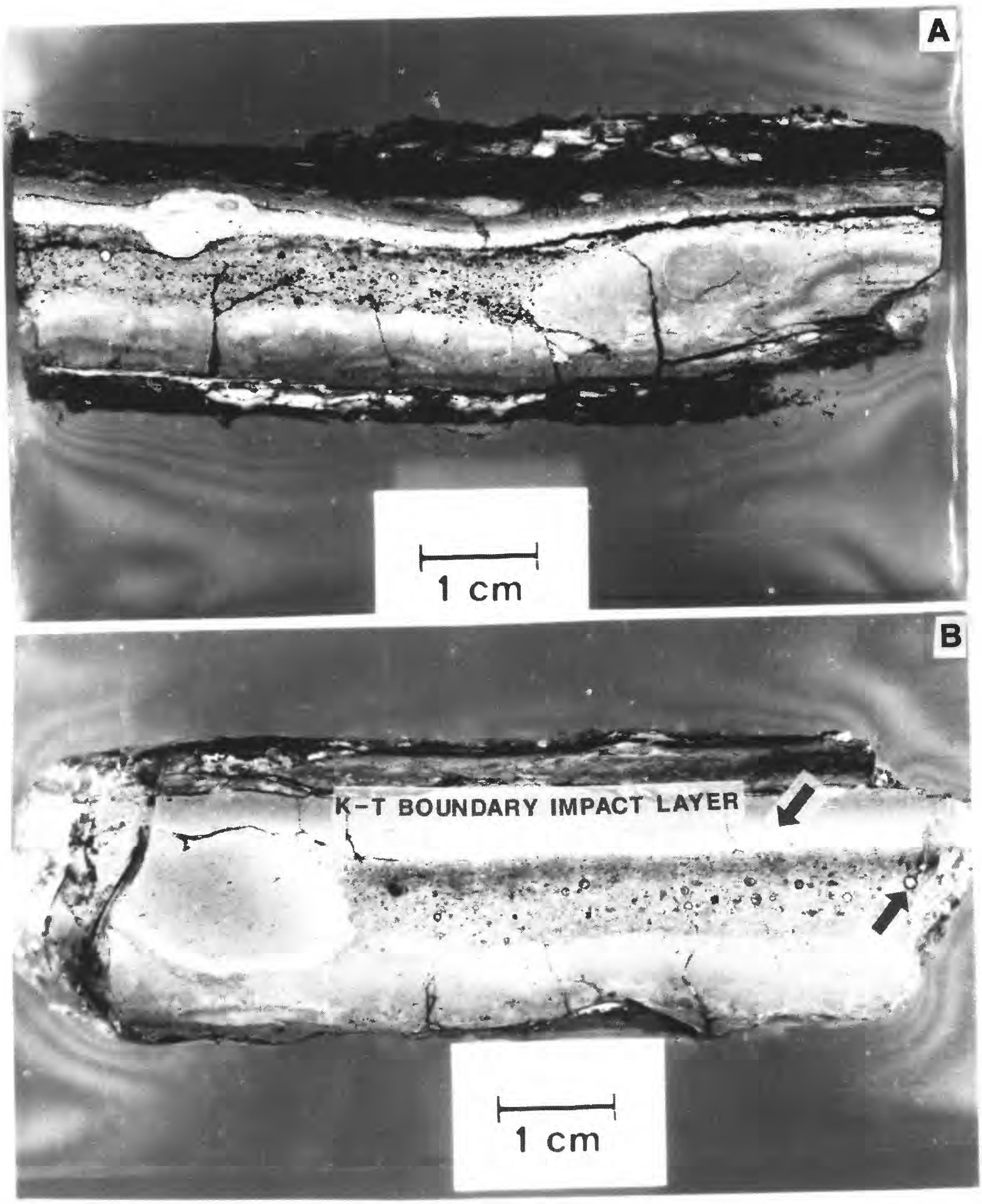


Figure 14.--Polished slabs of the K-T boundary interval from the Teapot Dome, Wyo., locality. A, Prominent dark layer which contains goyazite spherules, occurs at the top of the K-T boundary claystone bed. First appearance of shock-metamorphosed mineral grains occurs on top of the goyazite layer. B, Prominent layer of goyazite-enriched claystone abuts a pellet of kaolinitic claystone that is free of goyazite. Arrows point to K-T boundary impact layer and a large goyazite spherule.

Thin sections of the goyazite-enriched claystone show that the shells or walls of the goyazite spherules are colloform (fig. 15); that is, the goyazite shells consist of laminae upon laminae of concentrically oriented fibrous goyazite (fig. 15B). Some laminae are probably slightly different in chemical composition from their neighbors. This conclusion is based on the observation that some goyazite laminae have a slightly different color as viewed in plane-polarized light. Spherical growth began at point sources along the surface of the spherical cavity in which the goyazite was deposited as a gel, probably from a phosphate-rich colloidal fluid. One of the chief characteristics of the spherules is that the goyazite walls of the spherules have a colloform inner surface reflecting the spherical growth habit of the goyazite. Spheres and partial spheres of goyazite project into the central spherical cavity. The spherules, thus, have the appearance of hollow geodes.

Some hollow goyazite spherules were deformed or crushed; however, spherules filled with gypsum were, for the most part, not deformed or crushed. A few crushed goyazite spherules contain gypsum in their central areas. In these crushed spherules, broken shell fragments are surrounded by gypsum. Crushed and deformed hollow spherules are flattened parallel to bedding and occur adjacent to uncrushed ones. Lenses, stringers ,and irregular-shaped masses of the broken walls of spherules occur among uncrushed spherules. These relationships suggest that several growth generations of spherules occur in the layer.

Thin sections also show that goyazite forms not only the walls of spherules but also subspherical cavity fillings that have, in detail, scalloped borders (fig. 16B). The walls of the spherules consist of microlaminated goyazite that was precipitated laminae upon laminae in a void in the claystone. Goyazite also forms nonspherical objects (plant material?) such as the rectangular shaped one shown in figure 13B. This goyazite object occurs within the claystone near numerous crushed and uncrushed goyazite spherules.

In the goyazite-rich claystone layer, goyazite occurs not only as spherules but also as ellipsoid-, spindle-, teardrop-, and egg-shaped objects. A few intergrown double spherules of goyazite (fig. 16A) were also found in the goyazite enriched layer. Goyazite also occurs as disseminated irregular-shaped masses, laminae, and veins. Moreover, goyazite masses locally have been concentrated at the top of the spherule layer. Rarely, plant material (twigs) have been replaced by goyazite. Rare goyazite spherules also occur in the boundary claystone below the goyazite-enriched layer, but only in claystone that lacks disseminated, carbonaceous shreds and flecks. Such carbonaceous material is most common at the base of the boundary claystone. Many samples of the boundary claystone from the Dogie Creek site show that it consists of reworked claystone, as reflected by its appreciable content of clastic quartz grains.

Many of the large goyazite spherules are hollow, and some are filled with granular to coarsely crystalline gypsum. The gypsum that fills the cavities contains tiny spherules and broken rim fragments of goyazite. It can be demonstrated microscopically in thin sections and macroscopically with a steel probe that the goyazite spherules in the gypsum are completely encased by the gypsum. This evidence and other evidence previously cited indicates that goyazite spherules originally could not have been glass, but instead the goyazite filled spherical and nonspherical cavities at several stages within the goyazite-enriched layer at the top of the K-T boundary claystone.

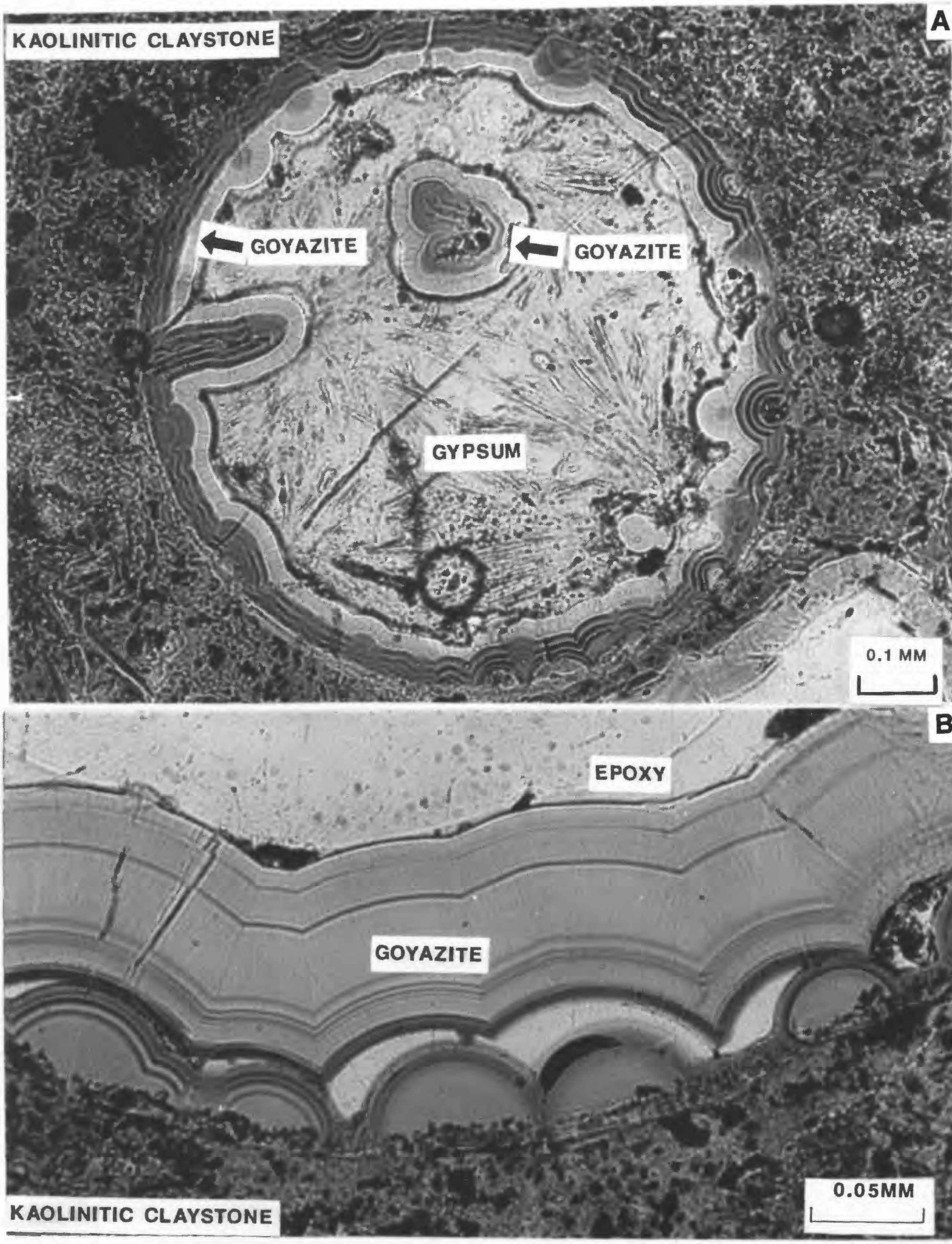


Figure 15.--Photographs of thin sections containing goyazite spherules from the Dogie Creek, Wyo., K-T boundary locality. A, Shell of spherule composed of colloform microlaminated goyazite. Center of spherule filled with gypsum. Goyazite inclusion in upper part of spherule. B, Details of the structure of the shell of a goyazite spherule. Colloform, fibrous goyazite filled a spherical cavity in the kaolinitic claystone of the boundary claystone.

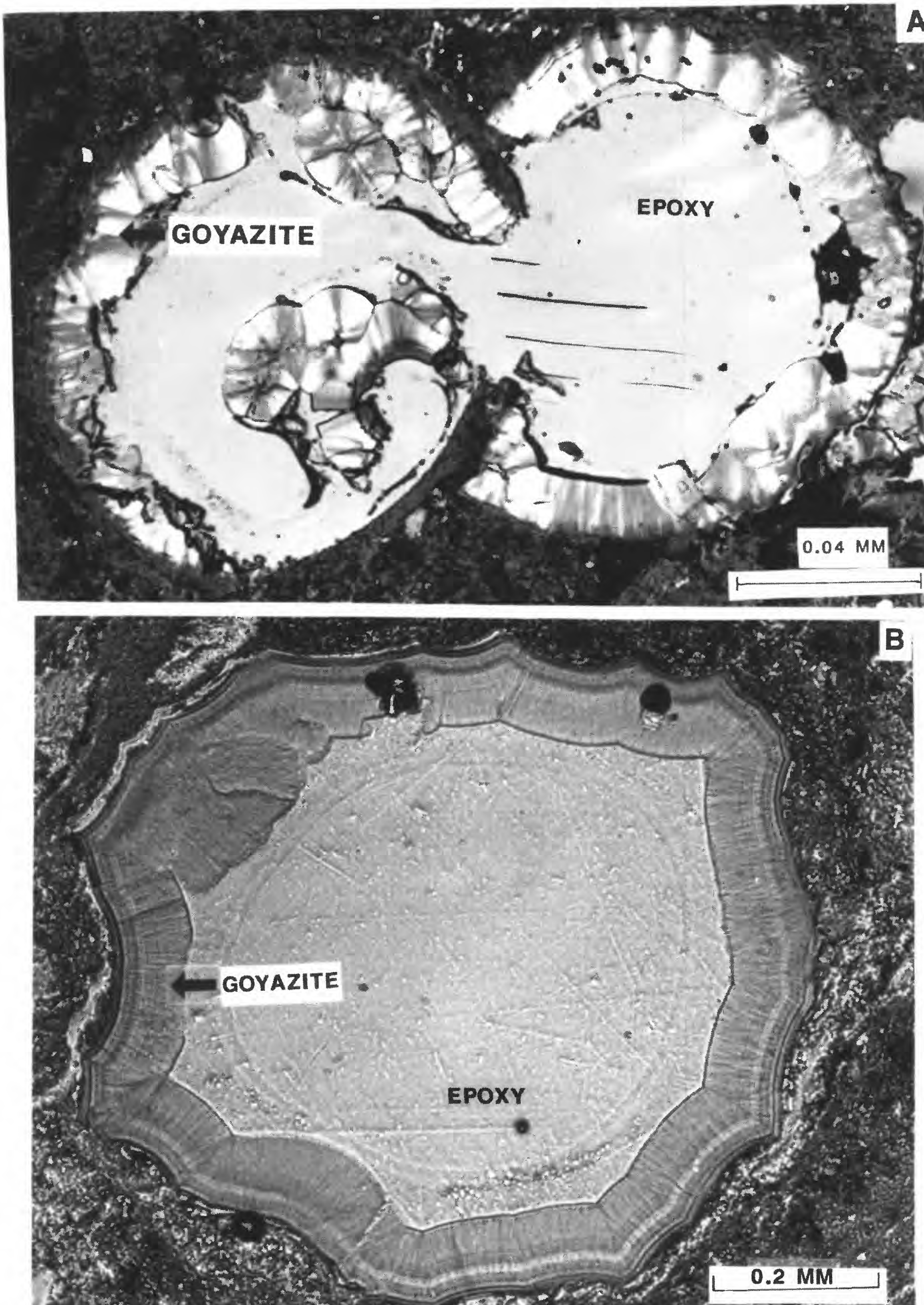


Figure 16.--Photographs of hollow goyazite objects in the K-T boundary claystone from the Dogie Creek, Wyo., K-T boundary locality. A, Double spherule composed of colloform microlaminated goyazite. Spherule on right formed first because its goyazite shell is cut off by the wall of the spherule to the left. B, Goyazite object showing a scalloped border. Wall of object composed of colloform goyazite.

Origin of Spherules

The origin of the spherules in the K-T boundary claystone in the Western Interior of North America is a controversial subject. Smit (1984), Pollastro and others (1983), Smit and Romein (1985), Bohor and others (1987b), Pollastro and Pillmore (1987), and Pillmore and Flores (1987) have either noted the occurrence or described spherules in K-T boundary rocks in the Raton Basin or Wyoming. Smit (1984) referred to them as being "microtektite-like" because of their spherical shape and their occurrence in a bed thought by him to consist of altered impact ejecta.

The kaolinite and goyazite spherules in the K-T boundary claystone in the Raton Basin are thought by some to be analogous to particles (described as spherules) composed of sanidine discovered by Smit and Klaver (1981) in marine K-T boundary rocks at Barranco del Gredero, Caravaca, Spain. They speculated that the spherules in Spain formed by solidification of melt droplets and "* * * were derived from the impacting body." Montanari and others (1984) found spherules of potassium feldspar (sanidine of Smit and Klaver, 1981) and mixtures of glauconite, magnetite, quartz, and pyrite at K-T boundary sites in northern Italy and in a deep-sea core from hole 465A in the central Pacific Ocean (see also Smit and Kyte, 1984; Kyte and Smit, 1986). Montanari and others (1984) studied the potassium feldspar spherules and, based on their internal texture, suggested that they formed diagenetically from calcic plagioclase melt droplets. They speculated that they formed by condensation of vaporized rock in the high-temperature fireball and cloud produced by the K-T impact. Thus, the potassium feldspar spherules are thought by some to be analogous to microtektites produced by asteroid impacts.

Interpretation by Shaw and Wasserburg (1982, p. 175) of isotopic data for the potassium feldspar spherules led them to suggest that they formed as a result of "* * * an oceanic impact in which both basaltic oceanic crust and overlying sediments or seawater were involved." They indicated that the isotopic results are compatible with a volcanic origin (authigenic growth from volcanic detritus) for the potassium feldspar. However, a continental source for the sanidine was ruled out by them. They concluded that a major problem still exists in that it is difficult to understand how to make potassium-rich sanidine that is extremely poor in strontium from any reasonable oceanic crust progenitor by an impact process. Isotopic analyses performed by DePaolo and others (1983) suggested to them that the present chemical composition of the potassium feldspar "* * * is attributable to authigenic growth."

Others (Hansen and others, 1986) have studied potassium feldspar and iron-oxide spherules in marine K-T boundary rocks in Europe and concluded that their morphology and structure indicates that they are not of impact origin. Instead, the spherules are the diagenetic infill of marine green algae (Prasinophyta). According to Hansen and others (1986, p. 870), these algae have been described as a disaster species. Their numbers increase explosively if the normal plankton population markedly declines. Izett (1987a) also studied K-T boundary samples from Caravaca, Spain, that contain the spherules, and he concluded that they may not be melt droplets of impact derivation. The potassium feldspar spherules are probably not altered basaltic melt droplets because their shapes are highly irregular and not compatible with the shape expected of melt droplets. Moreover, the spherules occur in claystone units stratigraphically below and above the impact layer. The open-meshwork texture of the potassium feldspar particles also suggests that they did not form by diagenetic alteration of melt droplets having an igneous quench texture as proposed by Montanari and others (1984).

In summary, two different types of impact origin have been suggested for the kaolinite and goyazite spherules in the K-T boundary claystone of the Western Interior of North America. One suggestion has been that the spherules were originally microtektites; that is they were originally solid glass spherules that probably formed from crustal target rocks during the earliest phase of the K-T impact event. The second suggestion (Bohor and Triplehorn, 1987; Bohor and others, 1987b) is that the hollow goyazite spherules, so common at the Wyoming K-T boundary sites, formed during the K-T boundary impact as hollow glass spherules derived from the "* * * depressurized rocks exposed at the impact-crater floor." Pillmore and Flores (1987, p. 119) concluded that the spherules in the Raton Basin "* * * may result from the alteration of material blown out during the K-T event, but they probably represent only authigenesis of some phosphatic constituent in the K-T boundary clay layer." On the basis of data presented elsewhere in this report, the kaolinite and goyazite spherules in the K-T boundary claystone in the Raton Basin may not be altered microtektites for the following reasons.

1. Spherules only occur in the K-T boundary claystone and not in the overlying K-T boundary impact layer, which is the chief stratigraphic location of an iridium abundance anomaly and shock-metamorphosed minerals. Seemingly the spherules (alleged microtektites) and the iridium anomaly should occur in the same layer composed, in part, of early-time, high-velocity impact ejecta.
2. Spherules are rare (far less than 1 percent) in the Raton Basin, but considerably more common (as much as 30 percent) at Wyoming K-T boundary sites. Such a variable abundance for the spherules (supposed microtektites) between relatively closely spaced localities, although not impossible, is difficult to reconcile with an impact origin.
3. Some spherules are indeed microtektite-like in their external shape, but many are not spherical and form peculiar-shaped objects (Izett, 1987a, fig. 9) unlike microtektites illustrated by Glass (1984, fig. 5).
4. Many spherules are hollow and unlike most microtektites which are solid (see Glass, 1984)
5. The surfaces of many of the goyazite spherules are smooth and unlike the pitted and corroded surfaces of tektites and microtektites (see Glass, 1984, p. 338).
6. If some of the spherules were originally solid microtektites that fell into standing water, then, because of their size and density, they seemingly would be concentrated at the base of the claystone bed. However, they are not concentrated at its base, but are widely distributed throughout the bed.
7. The occurrence of spherules of several different compositions at the same locality in the Raton Basin within a few millimeters of each other makes an impact origin unlikely. If the original material for each of these was a compositionally uniform impact glass, it is difficult to explain how one glassy spherule would alter to goyazite and an adjacent one alter to kaolinite.

Hollow goyazite spherules at the Dogie Creek, Wyo., K-T boundary site (figs. 15 and 16) were studied by Bohor and Triplehorn (1987) and Bohor and others (1987b), and they concluded that the spherules were produced during the K-T boundary impact event. Bohor and Triplehorn (1987) suggested that the hollow goyazite spherules are analogous to hollow flyash spherules formed during burning of coal at power plants. They drew this analogy because smaller goyazite spherules occur in clusters attached to the internal walls of the large hollow goyazite spherules. Bohor and others (1987b) speculated that the spherules formed during the K-T boundary impact as hollow glass spherules derived from the "* * * depressurized rocks exposed at the impact-crater floor." Bohor and Triplehorn (1987) proposed that the spherules "* * * along with finely divided glassy material * * *" were transported through the lower atmosphere in a "volcanic-like" surge cloud from the impact point to the deposition site in Wyoming. The existence of such surge clouds associated with large-scale impacts has not been established. Furthermore, the surge cloud must have traveled for at least 800 km from the nearest possible site for the K-T impact, the Manson, Iowa, impact structure. For a "volcanic-like" surge cloud to travel such a great distance is unlikely.

Bohor and Triplehorn (1987) further suggested that during diagenesis of the enclosing sediments the glass shells or walls of the spherules were devitrified to an unspecified silicate phase. Apparently this phase was subsequently replaced by the alumino-phosphate mineral goyazite. According to them, the devitrification and replacement processes exactly preserved the original morphology of the glass spherules. It should be pointed out, however, that no evidence of the former glass walls of the spherules can be seen in thin sections of the spherules (figs. 15 and 16). The walls now consist of laminae upon laminae of colloform goyazite. If the alleged glass walls had been completely dissolved away, it is unlikely that the goyazite that now fills the spherical cavities would mimic the shape of the proposed hollow glass flyash-like spherules of Bohor and Triplehorn (1987). Surprisingly, Bohor and Triplehorn (1987) proposed that the hollow glass spherules were converted to goyazite, but, that the "finely divided glassy material" that they envisioned as the precursor of the boundary claystone altered to kaolinite.

The following points bear on the origin of the hollow goyazite spherules in the K-T boundary claystone at the Dogie Creek and Teapot Dome sites.

1. Spherules only occur in the K-T boundary claystone and not in the overlying K-T boundary impact layer, which is the chief stratigraphic location of an iridium abundance anomaly (and shock-metamorphosed minerals).
2. At Wyoming K-T boundary sites, hollow goyazite spherules occur within a discrete stratigraphic layer at the top of the K-T boundary claystone whereas in the Raton basin they are widely disseminated in the boundary claystone.
3. Hollow goyazite spherules are common at K-T boundary sites in Wyoming and are rare or absent at boundary sites in the Raton Basin, Colo., and N. Mex., and Brownie Butte, Mont. Such a pattern for a supposed regional, air-fall, impact-generated deposit is anomalous.
4. Not all hollow goyazite objects are spherical. Some are highly irregular in overall shape and have scalloped margins. Goyazite also occurs as disseminated, irregular shaped masses, laminae, and veins in the claystone and locally replaces plant material.
5. Minute goyazite spherules occur within the gypsum that fills many of the hollow goyazite spherules.
6. Thin section study of rare double intergrown goyazite spherules indicates that they formed by sequential filling of adjacent cavities not as fused glass spherules.
7. Goyazite spherules are as much as 1.2 mm in diameter at Wyoming K-T boundary localities and about 0.05 mm at Raton Basin K-T boundary localities. Such a size-distribution pattern for a supposed regional, air-fall, impact-generated deposit is anomalous.

8. Most of the largest goyazite spherules are uniformly about 0.7-1.0 mm in diameter, and thus, the size distribution is markedly truncated at about 1.0 mm. This truncated size distribution of the spherules is difficult to reconcile with an impact origin.
9. Compacted masses of crushed spherules form lenses that occur adjacent to uncrushed hollow spherules. Seemingly, the spherules filled cavities during repeated episodes of growth.
10. If the goyazite spherules originally were hollow glass objects deposited in sediments of a pond, such as at the Teapot Dome locality, it is hard to visualize a depositional process that would concentrate the spherules in a discrete layer or in lenses within pond deposits. Seemingly, the hollow glass bubbles would float on the pond surface and would not be deposited as a discrete layer or lenses within the pond deposits.
11. Rare goyazite spherules occur in kaolinitic claystone below the goyazite-enriched layer layer. This suggests that the spherules do not comprise a primary air-fall deposit.
12. Rare goyazite spherules occur in kaolinitic claystone pellets that occur within the goyazite layer or within the impact layer.
13. No relict of the proposed glass walls of the hollow goyazite spherules can be seen in thin sections or oil immersion mounts of the spherules.
14. It is difficult to understand how the spherules could form late in the impact process from depressurized rocks exposed at the impact-crater floor as visualized by Bohor and others (1987b). If the hollow spherules formed late, they seemingly would be concentrated at the top of the K-T boundary impact layer rather than at the top of the K-T boundary claystone as they are.
15. Cratering studies (Roddy and others, 1987; Grieve, 1987, fig. 5) suggest that transient craters form early in the impact cratering process, but they are modified and filled rapidly with target materials. If this is correct, it is difficult to understand how the spherules could form late in the impact process from depressurized rocks exposed at the impact-crater floor as visualized by Bohor and others (1987b).
16. It seems unlikely that crustal crystalline rocks of the impact crater floor could provide a source for the large amount of volatiles needed to sustain a pyroclastic-type eruption and to form the spherules.

The combined weight of the evidence given above suggests that the spherules in the K-T boundary claystone may not be of impact origin; however, their exact origin remains a mystery. The spherules are clearly authigenic, and many formed by the filling of spherical and nonspherical cavities in the claystone. The spherules possibly are the infill of algae (see Hansen and others, 1986) or some unknown organic material.

K-T BOUNDARY IMPACT LAYER

A thin microstratigraphic unit of microlaminated claystone overlies the K-T boundary claystone in the Raton Basin (fig. 1) and other boundary sections in Western North America. The most surprising observation I made during study of K-T boundary rocks is that this unit, rather than the underlying boundary claystone, contains a concentration of shock-metamorphosed mineral grains (fig. 11). This observation has important implications concerning dispersal-deposition mechanics and arrival time of impact ejecta in the stratigraphic section in Western North America. The precise stratigraphic location of the shock-metamorphosed materials was determined by microscopic study of about 100 polished slabs of K-T boundary interval specimens similar to those shown in figure 1. It should be emphasized that the magnitude of the shock-metamorphic mineral anomaly is enormous because rocks below the K-T boundary interval are devoid of shocked minerals. This spike of shocked mineral grains occurs at each site in the Raton Basin in the same microstratigraphic sequence and unit. The shock-metamorphosed grains are not obviously concentrated at the base of the unit as would be expected if the impact layer were primary air-fall impact material. Because the unit always contains a concentration of shocked mineral grains and generally registers the peak iridium abundance anomaly in the boundary interval, I have informally named it the K-T boundary impact layer.

The thickness of the impact layer in the Raton Basin, as determined on 46 polished slabs from 12 localities, averages 5.2±1.1 mm and ranges from 4.0 to 8.0 mm. At the Clear Creek North, Colo., locality, the layer is at its thickest, 8.0 mm, because of numerous jarosite laminations. At other Western North American K-T boundary sites, such as the Teapot Dome, Wyo., and Brownie Butte, Mont., localities, the impact layer is not as thick, on average, as it is in the Raton Basin. For example, the thickness of the impact layer at the Teapot Dome, Wyo., locality ranges from 3.0 to 5.0 mm. Thus, the impact layer is considerably thinner than the 1-cm-thickness that was visualized for the global fallout layer by Alvarez and others (1980). At European K-T boundary localities the impact layer is only about 1 mm thick (Izett, 1987b).

The impact layer is composed mainly of kaolinite and I/S mixed-layer clay containing as much as 50 percent illite layers according to C.G. Whitney (U.S. Geological Survey, written commun., 1986). The claystone of the impact layer differs from that of the boundary claystone as shown by X-ray diffractograms of glycolated samples heated to 550 ^{o}C. For an unknown reason, the structure of the kaolinite and I/S mixed-layer clay of the boundary impact layer collapses and forms a sharp peak on diffractograms at about 8.8^{o} 2θ. In contrast, the structure of the kaolinite and mixed-layer clay of the boundary claystone collapses in the same region to nearly a flat trace.

The texture of the claystone as seen in thin sections is different from that of the boundary claystone. It consists of alternating brown stained laminae of kaolinite and feathery masses that probably are I/S mixed-layer clay because they have a markedly higher birefringence as compared to the kaolinite (fig. 17). The orientation of the clay minerals in the feathery masses is roughly parallel to bedding. SEM analysis shows that locally the claystone has a texture (fig. 18B) similar to smectite clay and different from the microspherulitic texture of the underlying boundary claystone (fig. 5A). Pollastro and Pillmore (1987) thought that the microspherulitic texture resulted from the alteration of glass.

At most localities, the claystone is randomly speckled with minute (10-50 micrometers in diameter) single jarosite spherules that probably formed from framboidal pyrite. Locally, jarosite spherules form planar groups and lenses. As measured by the energy dispersive system of an SEM, some perfectly formed jarosite crystals contain a trace amount of titanium and sodium. The titanium and sodium in the jarosite may account for its slightly different X-ray diffraction pattern as compared to that of typical jarosite. ICP analysis of the jarosite shows that it contains 0.3 percent titanium and 1,500, 120, and 360 ppm of arsenic, chromium, and vanadium, respectively (table 4).

Several lithic features make the impact layer a distinctive unit readily identifiable microscopically. The chief diagnostic feature of the claystone is its microlaminated nature that contrasts markedly with the generally massive underlying boundary claystone (fig. 11). Microlaminations of claystone are warped over and under clastic mineral grains reflecting soft

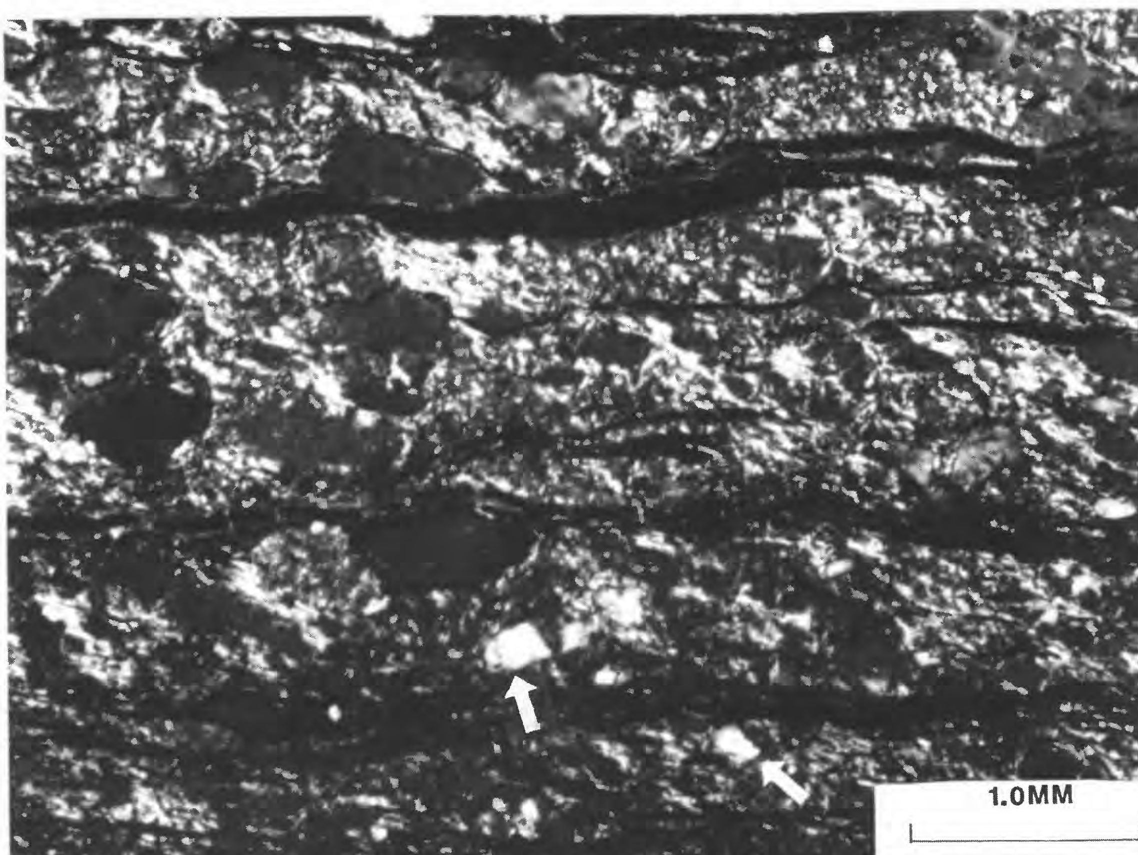


Figure 17.--Photomicrograph of a thin section of the K-T boundary impact layer. Specimen from the Berwind Canyon, Colo., locality shows typical feathery texture of the claystone, laminae of black vitrinite, and gray almond-shaped kaolinite pellets. Scattered clastic grains of quartz and other minerals shown at arrows. Some of these grains contain multiple intersecting sets of shock lamellae. Photographed with cross-polarized light.

A)

B)

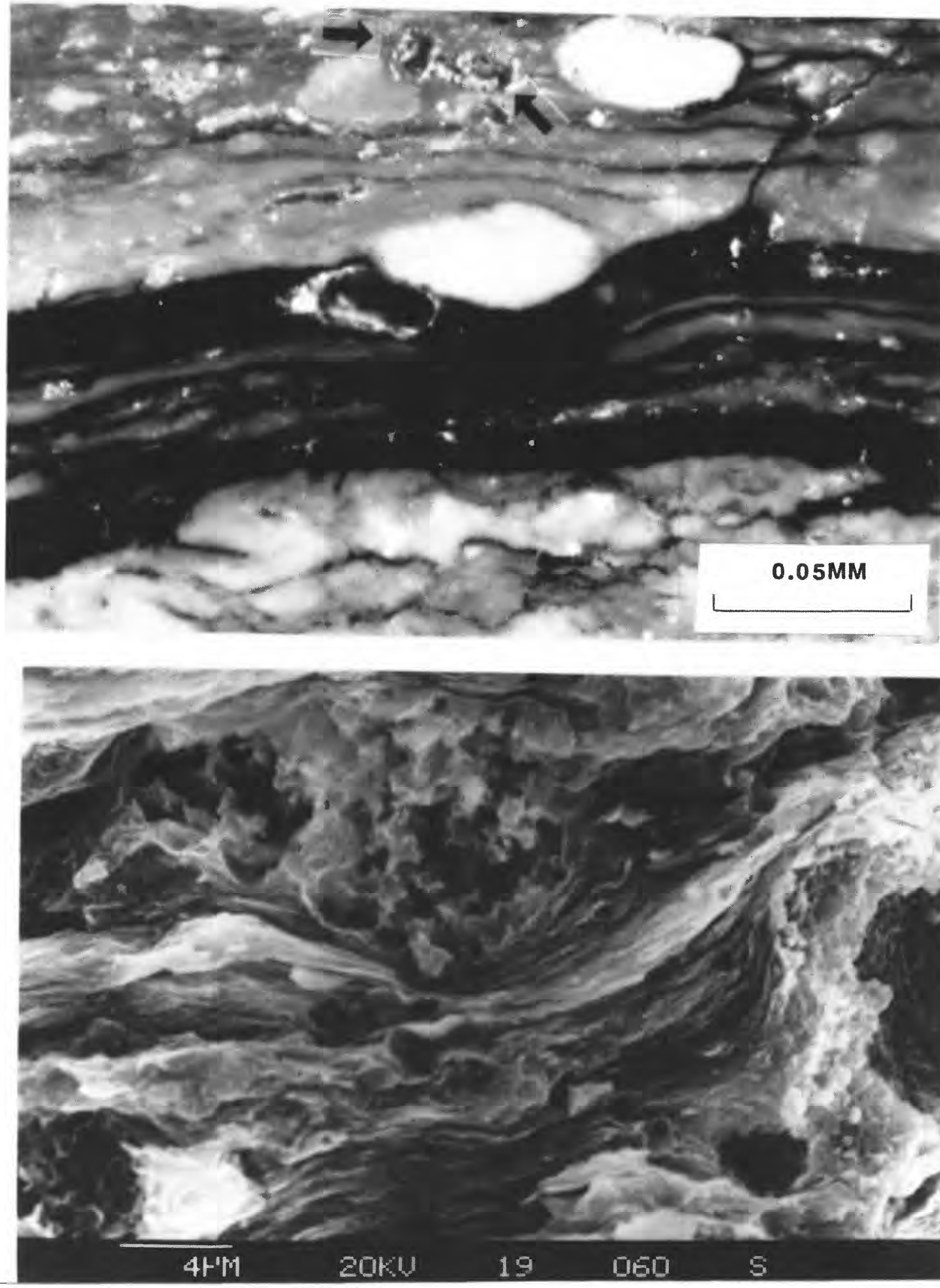


Figure 18.--Photographs of the K-T boundary impact layer. A, Polished surface of specimen from the Clear Creek North, Colo., locality showing laminae of black vitrinite and white almond-shaped kaolinite pellets. Clastic quartz grains shown at arrows. Vitrinite laminae deformed below kaolinite pellet. B, SEM photomicrograph of specimen from the Madrid East, Colo., locality showing the texture of the claystone. Note pits in claystone that were framboidal pyrite.

sediment compaction. The surfaces of the microlaminae display the impressions of macerated plant material, suggesting slow accumulation during normal depositional processes. Not only is the claystone microlaminated, but it also contains planar vitrinite laminae a few tens of micrometers thick. These vitrinite laminae may have formed from leaves and other plant materials.

Another diagnostic lithic feature of the impact layer is its ubiquitous content of pellets of white, cryptocrystalline kaolinite similar to "graupen" of tonstein layers in coal beds (Williamson, 1970; Burger, 1985). In cross section they are eye-shaped (fig. 18A) and in the third dimension they are almond-shaped. Their major axes are parallel to bedding, and the aspect ratio is about 2, but some are more flattened and have a ratio of about 3. In general, planar laminae within the impact layer are deformed around many pellets, suggesting that the laminae were deformed by compaction around the more competent pellets. In some cases, vitrinite laminae encase the pellets. The claystone of many pellets appears massive and lacks any detectable structure such as a concentric or radial pattern. SEM examination of the pellets shows that some are finer grained than the host claystone.

In addition to the ubiquitous almond-shaped pellets of kaolinite, the impact layer locally contains another type of pellet (0.2-0.5 mm long) that has an irregular cross-sectional outline (fig. 19). Some of these pellets contain wispy, macerated, carbonaceous material and have a microscopic texture identical to that of kaolinite of the boundary claystone. Rare fragments contain minute clastic quartz grains. Such pellets are probably rip-up clasts of the underlying boundary claystone that were fairly well indurated prior to their erosion and redeposition.

Allen and Nichols (1945) and Allen (1946, figs. 2 and 5) described and illustrated kaolinite pellets similar to those of the impact layer. They occur in the continental Calapooya Formation of Eocene age in Lane County, Oregon. Because they lack a radial or concentric structure, occur in a clay matrix, and are associated with charcoal, lignitic material, and diatoms, Allen and Nichols (1945) proposed that they formed by a sedimentary fluvial process whereby flakes of claystone were ripped up from the surface of Eocene floodplains and redeposited in the sediments that now form the Calapooya Formation. The pellets of the impact layer have a similar setting to those in the Eocene Calapooya Formation in that they occur in a clayey matrix and are associated with carbonaceous material. Because the pellets are similar to the kaolinite of the underlying boundary claystone, they may have formed during an episode of gentle, winnowing fluvial activity that followed deposition of the boundary claystone. Kaolinite pillows that had formed during a period of desiccation and wetting of the uppermost surface of the boundary claystone could have been ripped up during the fluvial episode and incorporated in the impact layer.

Exceedingly rare, minute (20-40 micrometers), perfectly formed crystals of a spinel-group mineral (magnesioferrite) have been found in K-T boundary marine sediments in Europe (Bohor and others, 1986; Kyte and Smit, 1986). Such crystals are thought by these authors to have formed during the K-T boundary impact event. Izett (1987a) also studied magnesioferrite crystals in the K-T boundary impact layer at Caravaca, Spain, and pointed out several problems with an impact origin for the crystals. Large samples of the K-T boundary impact layer were collected from several localities in the Raton Basin with the purpose of searching for magnesioferrite crystals. Not a single crystal of this mineral was found. This is puzzling in view of the rare but widespread occurrence of this mineral in marine K-T boundary rocks (Kyte and Smit, 1986).

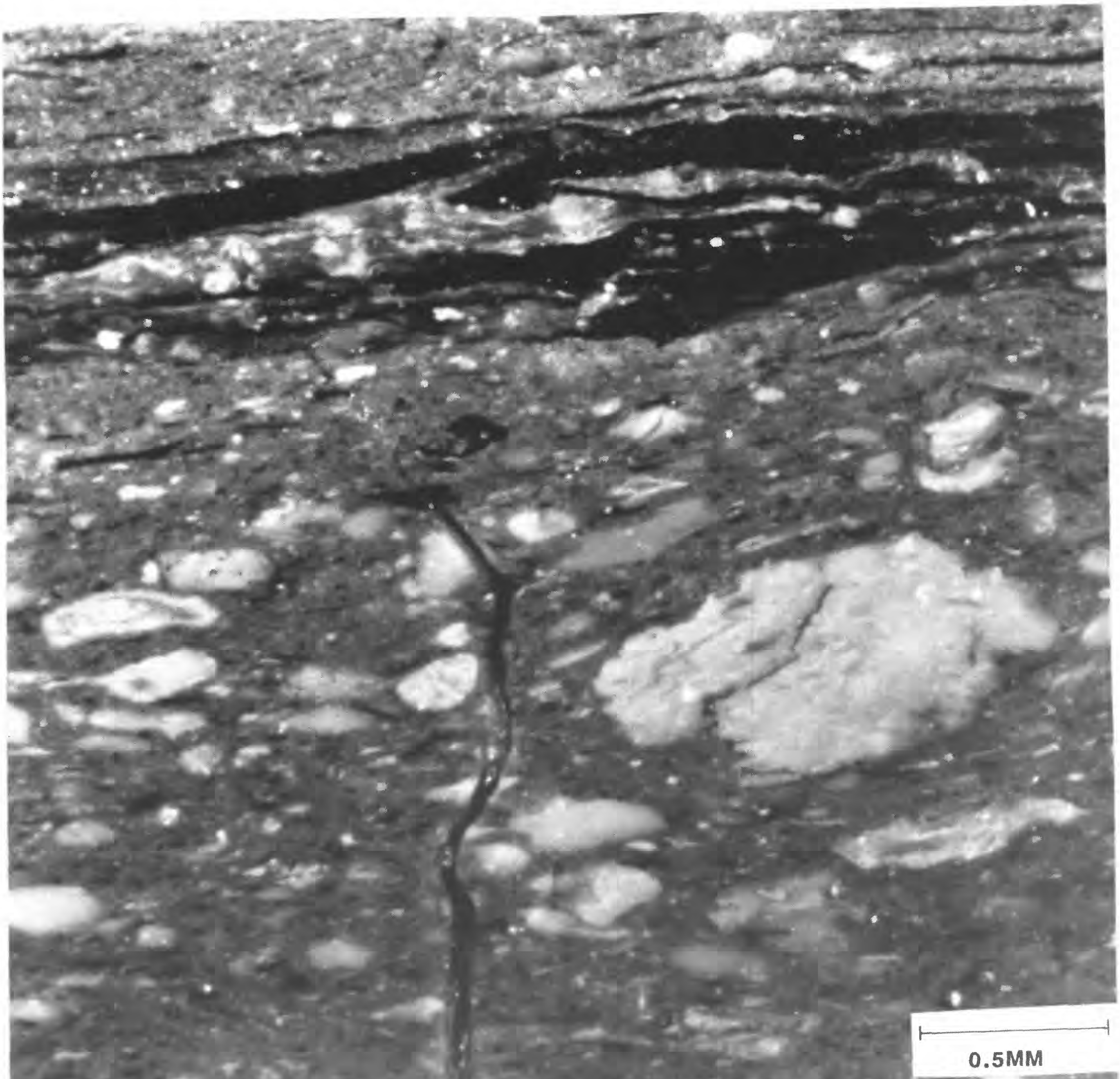


Figure 19.--Photograph of the K-T boundary impact layer at the Berwind Canyon, Colo., locality. Typical laminae of black vitrinite and light-gray kaolinite pellets called "graupen" are shown. Large pellet at arrow has an irregular outline suggesting it is a rip-up clast of the underlying boundary claystone.

One possibility that may account for the lack of magnesioferrite crystals in the continental K-T boundary rocks of the Raton Basin and their presence in marine K-T boundary rocks is that they were unstable during diagenesis. Another possibility that may account for their occurrence in marine K-T rocks is that the magnesioferrite crystals are not of impact origin, but rather are derived from alteration of cosmic glassy spherules. Such spherules are known to accumulate in slowly deposited marine sediments (Brownlee, 1981, p. 743).

Chemical Composition of the K-T Boundary Impact Layer

The major-element chemical composition of four samples of the impact layer from the Raton Basin was determined by XRF analysis, and the results are listed in table 6. The analyses show that the impact layer is similar to the underlying boundary claystone (table 1), except that the former contains considerably more iron and potassium than the latter, reflecting its jarosite (potassium-bearing iron sulfate) content. In addition, the claystone of the impact layer has a slightly larger Si to Al ratio than that of the boundary claystone.

The trace-element content of the impact layer claystone (including jarosite laminae) was determined by the ICP method (table 7), and its chemical composition is similar to that of the subjacent boundary claystone (table 3) except that it contains less lithium and slightly more barium, chromium, copper, nickel, vanadium, and zinc. The trace-element content of the impact-layer claystone is also similar to a thin bed of dark carbonaceous flaky shale that overlies the impact layer at the Starkville South, Colo., locality (table 4). Moreover, the trace-element composition of the impact layer is similar, in general, to the average shale of the Earth's crust (table 4).

The K-T boundary impact layer in the Raton Basin has a low content of nickel (6-16 ppm) and cobalt (3-7 ppm) (table 7). For comparison, a mudstone that occurs about 4 cm below the K-T boundary claystone layer contains as much nickel as does the impact layer (tables 4 and 7). The low content of nickel and cobalt suggests that claystone of the impact ejecta does not contain a large component of asteroidal material, which should contain significant amounts of these elements. In contrast to the low content (< 5ppm) of these elements in the impact layer claystone in continental K-T boundary rocks of the Raton Basin, the K-T boundary impact layer in marine rocks at Caravaca, Spain, and Stevns Klint, Denmark, contains 2,580 (ppm) and 1,370 (ppm) nickel and 626 (ppm) and 146 (ppm) cobalt, respectively (Kyte and others, 1985).

In the Raton Basin, Orth and others (1981) found and reported above-background amounts of iridium in continental sedimentary rocks at the precise palynological K-T boundary. Prior to this discovery, iridium abundance anomalies had only been found in marine rocks, and, thus, the discovery provided important support to the impact-extinction theory. Finding anomalously large amounts of iridium in continental rocks seemingly undercut suggestions by some that the K-T boundary iridium spike resulted from normal marine and geochemical processes including (1) slow deposition and incorporation of iridium-bearing meteoritic dust or preferential removal of sediment at the expense of meteoritic material, thereby increasing the iridium content (Kent, 1981), and (2) precipitation of iridium from seawater owing to a change in ocean chemistry caused by the latest Cretaceous mass extinction event (Kyte and others, 1981). If these suggestions were correct, iridium abundance anomalies should be found at numerous levels in the stratigraphic column. However, a methodical search for significant iridium abundance anomalies, other than the one at the K-T boundary, generally has been unsuccessful (Orth and others, 1985; Kyte, 1986).

Table 6.--X-ray fluorescence analyses of the K-T boundary impact layer from Colorado and New Mexico.
[Analysts: A.J. Bartel, K.C. Stewart, and J.E. Taggart, U.S. Geological Survey; oxide elements in weight percent; Fe calculated as Fe_2O_3; LOI, lost on ignition at 900 °C. Localities of figure 2: BER, Berwind Canyon, Colo.; SVS, Starkville South, Colo.; MADE, Madrid East, Colo.; MR, Madrid Railroad, Colo.]

	BER	SVS	MADE	MR
SiO_2	42.2	46.8	45.7	42.7
Al_2O_3	26.0	25.3	28.9	27.5
Fe_2O_3	6.81	4.13	3.99	9.14
MgO	0.79	0.82	0.59	0.51
CaO	.51	.26	.28	.30
Na_2O	.29	.29	.25	.15
K_2O	2.42	2.47	1.38	1.24
TiO_2	1.42	1.29	2.27	1.81
P_2O_5	.15	<.05	.06	.17
MnO	<.02	<.02	<.02	<.02
LOI	18.0	18.3	16.7	16.9

Table 7.--Induction-coupled atomic emission spectroscopic analyses of the K-T boundary impact layer from Colorado and New Mexico. [Analyst: M.J. Malcolm, U.S. Geological Survey; Ti in weight percent; other elements in parts per million. Localities of figure 2: BER, Berwind Canyon, Colo.; MR, Madrid Railroad, Colo.; MADE, Madrid East, Colo.; SVS, Starkville South, Colo.; CCN, Clear Creek North, Colo.]

	BER	BER	MR	MR	MADE
Ti	1.3	1.0	0.4	1.1	1.5
Mn	14	14	23	22	10
As	100	170	60	110	80
Ba	830	700	760	1200	330
Ce	78	58	180	52	20
Co	3	5	7	6	6
Cr	230	310	110	130	160
Cu	68	69	50	50	32
Ga	30	30	28	27	28
La	42	30	79	27	12
Li	89	51	16	97	81
Mo	8	6	10	4	8
Nd	31	23	82	26	9
Ni	6	6	16	9	6
Sc	40	43	24	24	30
V	280	320	190	170	60
Zn	45	52	73	2	37

In their search for an iridium abundance anomaly in the Raton Basin, Orth and his colleagues were able to concentrate their effort in a narrow stratigraphic interval because of the prior palynological work of Tschudy (1973) and Pillmore (1976). Tschudy (1973) had identified a floral break in the lower part of the Raton Formation, as well as at other places in the Western Interior of North America, and judged that the break marks the K-T boundary.

As a result of the work of Orth and others (1981) and Pillmore and others (1984, fig. 1), iridium abundance anomalies were found at eight sites in sedimentary rocks of the Raton Basin at the palynological K-T boundary. Gilmore and others (1984, fig. 2) showed that the peak concentration of iridium occurs in coal or carbonaceous material slightly above the K-T boundary claystone bed at three of four localities they studied.

Because previous workers did not recognize or sample the impact layer as a discrete unit, I made collections across the K-T boundary at four localities in the Raton Basin to determine the iridium, silver, and gold content. RNAA analyses of the samples were made by H.T. Millard, Jr., of the U.S. Geological Survey, following procedures he developed (Millard, 1987). The data are listed in table 5 and shown graphically on figure 8. Millard's analyses show that the iridium content of the impact layer varies from 1.2 to 14.6 ppb; this amount is considerably more than that in the underlying boundary claystone and subjacent Cretaceous rocks. The detection limit for iridium, as determined by Millard, was about 0.004-0.07 ppb. In general, the iridium data are similar to those reported by Orth and others (1981), Pillmore and others (1984), and Gilmore and others (1984), but Pillmore and Flores (1987) reported a considerably larger iridium spike (56 ppb) at the Starkville South, Colo., locality than that measured by Millard (1.2 ppb) at the same locality. However, the data sets are not strictly comparable because the impact layer was not recognized or sampled by the above cited authors.

As shown on figure 8, the iridium concentration generally peaks in the K-T boundary impact layer; nevertheless, large values also can occur in carbonaceous-rich layers either below or above the impact layer. For example, significant iridium anomalies have been measured below the K-T boundary claystone at the Clear Creek North, Colo., and Starkville South, Colo., localities (0.83 ppb and 1.5 ppb, respectively). At the Madrid East, Colo., locality, a significant iridium anomaly (0.47 ppb) was found in carbonaceous shale at the top of the boundary coal bed.

The surface concentration of iridium in rocks that span the K-T boundary varies considerably at localities only a few kilometers apart. Orth and others (1981) and Gilmore and others (1984) reported values that average 50 ng/cm^2 and range from 8 to 120 ng/cm^2 for six localities in the Raton Basin. Because shock-metamorphosed minerals are concentrated in the impact layer, I made the assumption that the impact-generated iridium also was originally concentrated in this layer. During diagenesis of the sediments, some iridium was probably mobilized and transported in solution away from the impact layer and concentrated in adjacent carbonaceous-rich sediments. Bowles (1986) showed that platinum-group metals can be mobilized and transported in acidic low-temperature aqueous solutions and deposited in carbonaceous-rich sediment during weathering and diagenesis of rocks (see also Westland, 1981).

Luck and Turekian (1983) determined the ratio of osmium isotopes ($^{187}Os/^{186}Os$) in a sample (presumably of the impact-layer claystone) collected at the Starkville South, Colo., locality by C.J. Orth, of the Los Alamos National Laboratory. Orth reportedly (Luck and Turekian, 1983, p. 615) sampled material that contains the highest iridium abundance in the boundary interval (56 ppb, according to Pillmore and Flores, 1987, fig. 1), which is in the impact layer. According to Luck and Turekian (1983), the claystone from Starkville South yielded a $^{187}Os/^{186}Os$ ratio of 1.29±0.04. This ratio is much different than that of modern crustal material, about 10. They (1983, p. 615) observed that the ratio of the Starkville South claystone (1.29) is clearly different than the ratio (1.65) of the marine boundary claystone (the so-called fish clay) from Stevns Klint, Denmark, so different that they represent asteroid material of different compositions or different events. They also suggested that the ratio is incompatible with a chondritic composition for the asteroid. It also should be emphasized that the impact layer in the Raton Basin may not be the exact time equivalent of the material from the boundary claystone (fish clay) analyzed by Luck and Turekian (1983) at Stevns Klint, Denmark (see figure 10).

Origin of Iridium Anomaly

The origin of the anomalous amounts of iridium in K-T boundary rocks was attributed by Alvarez and others (1980, p. 1102) to "* * * an abnormal influx of extraterrestrial material." They suggested that a 10-km-sized asteroid of chondritic composition struck Earth and created enormous amounts of dust-sized crustal and asteroidal material. Most of the asteroid was vaporized on contact with the Earth (Melosh, 1982, p. 124). This vapor and its condensate, relatively rich in platinum-group elements, was ejected into the atmosphere and globally dispersed.

McLean (1985), Officer and Drake (1983, 1985), and Officer and others (1987) suggested that the source of the iridium at the K-T boundary was terrestrial and not extraterrestrial in origin. According to them, the iridium was generated during an intense period of volcanic activity. If this volcanism was silicic, the geologic record does not support the idea that there was an intense period of silicic volcanism at the K-T boundary. Altered silicic ash beds (bentonites) are abundant in Cretaceous marine rocks of Western North America and represent several peaks of volcanic activity (Izett, 1981, p. 10202-10203), but they are rare in continental K-T boundary interval rocks. Silicic volcanic ash beds are also rare in marine K-T boundary interval rocks elsewhere in the world, including Europe, New Zealand, Alaska, and Asia. More importantly, silicic volcanic rocks are thought to have extremely low iridium concentrations.

The evolutionary igneous process that begins with the generation of silicic magma and that ends with the ejection into the atmosphere of enormous amounts of pyroclastic material has been an often repeated process throughout geologic time (Gilluly, 1965, p. 1). The record of silicic, pyroclastic volcanism in the Precambrian is known chiefly by the occurrence of clearly discernable metamorphosed ignimbrites and tuffs in Precambrian rock sequences. Precambrian pyroclastic rocks have been described in Arizona by Anderson and Creasey (1958); in Colorado by Olson (1976) and Wrucke and Dings (1979); in New Mexico by Barker and Friedman (1974); in the Midcontinent region by Hanson and Al-Shaieb (1980) and Sides and others (1981); and in Virginia by Rankin and others, (1969, p. 741-744). It seems reasonable to assume that a considerable number of silicic ash beds must have resulted from these Precambrian pyroclastic volcanic eruptions.

Gilluly (1965, p. 1) noted that volcanism in the Phanerozoic was "* * * episodic but on the whole quasicontinuous." He further stated (1973, p. 504) that "Except for two intervals in the Triassic, there have been no epochs as long as 5 m.y. without some record of siliceous magmatism in the North American Cordillera since the end of the Permian."

Numerous bentonites that record silicic, pyroclastic volcanism occur in Paleozoic marine rocks in the central and eastern United States. They have been reported in Ordovician rocks by Nelson (1921), Allen (1932), Bonnie and Honess (1929), Maddox (1930), Kay (1931), Rosenkrans (1936), Wilson (1949), and Ross and others (1981); in Silurian rocks by Honess (1923, p. 107-109); in Devonian rocks by Honess (1923, p. 123), Hass (1948), Fettke (1952), Flowers (1952), and Roen (1980); and in Mississippian rocks by Miser (1920), Honess, (1921, p. 70-71), and Niem (1977).

An extraordinarily large number of bentonite beds occur in Mesozoic rocks of the Western Interior of North America. A biotite-bearing bentonite bed in marine Lower Triassic rocks (Meekoceras zone) was found by G.D. Frazer, of the U.S. Geological Survey (written commun., 1978), in the Pequop Mountains of eastern Nevada. Thin beds of bentonite have been reported from Upper Jurassic marine rocks in Colorado (Pipiringos and others, 1969, p. N-22), Wyoming (Pipiringos and Imlay, 1979, p. 23), and Utah (Schultz and Wright, 1963). Bentonite beds in the Jurassic Morrison Formation of the Colorado Plateau have been collected for K-Ar dating (J.D. Obradovich, oral commun., 1985).

It is clear that an exceptionally large number of bentonite beds occur in marine Cretaceous shale units of the Western Interior of North America (Rubey, 1929; Knechtel and Patterson, 1956; Slaughter and Earley, 1965; Hattin, 1975). For example, Kauffman (1985, p. 8) recorded about 700 discrete bentonite beds in marine rocks of the Western United States that range in age from Albian to late Maastrichtian. Pyroclastic materials in Cretaceous rocks in Arkansas and adjoining areas were reported by Ross and others (1928). Peaks of volcanic activity within the Cretaceous were reported by Heathman (1939, p. 16-17) who counted 71 bentonite beds in a roadcut exposure of 73 m of the Lower Cretaceous Mowry Shale near Casper, Wyo. Mapel and Pillmore (1963) recorded 16 bentonite beds in the lower 23 m of the Mitten Black Shale Member of the Pierre Shale near Newcastle, Wyo.

Thin tonsteins (altered volcanic ash) are common in coal-bearing continental Cretaceous and lower Tertiary rocks of Colorado, New Mexico, Utah, and Wyoming (Bohor, 1977; Ryer and others, 1980), but they are not nearly as common as bentonite beds in marine Cretaceous rocks. They can be found in Paleocene and Eocene coal beds in Colorado, New Mexico, Utah, and Wyoming (Bohor, 1977). Altered volcanic ash beds in the Eocene Green River Formation in Colorado, Utah, and Wyoming were studied by Bradley (1964) and Desborough and others (1973). Altered volcanic ash beds are common in Eocene and Oligocene sedimentary rocks in the Gulf Coast states according to Ross and others (1928, p. 193).

Lenticular layers of glassy volcanic ash are ubiquitous in upper Tertiary sedimentary rocks of the Western United States. Volcanic ash beds are interbedded in sedimentary rocks of the Great Plains including the White River (Oligocene), Arikaree (Oligocene and Miocene), and Ogallala (Miocene) Formations and their equivalents in the Rocky Mountains and Great Basin (Izett and Barclay, 1973; Gausha and Blick, 1971). These ash beds attest to repeated episodes of pyroclastic volcanism in the Great Basin (Evernden and others, 1964; Izett, 1968). About 70 silicic ash beds were reported by Izett (1981) in Pliocene and Pleistocene sedimentary rocks of the Western United States.

From the foregoing discussion it is clear that silicic pyroclastic volcanism has been an often repeated process throughout Precambrian and Phanerozoic time. Moreover, observational evidence clearly shows that altered silicic volcanic ash beds (bentonites) are rare near the K-T boundary in North America. Not only are silicic volcanic ash beds rare in the K-T boundary interval rocks of North America, but also in marine K-T boundary interval rocks elsewhere in the world, including Europe, New Zealand, Alaska, and Asia.

A second possibility is that the intense period of K-T boundary volcanism proposed by McLean (1985), Officer and Drake (1983, 1985), and Officer and others (1987) was basaltic, and, indeed, such rocks contain more iridium than silicic ones (Palme, 1982). Moreover, relatively large amounts of iridium were detected in airborne particulate material ejected from Kilauea volcano on the island of Hawaii (Zoller and others, 1983; Olmez and others, 1986). Furthermore, it is true that an enormous volume of basalt, the Deccan traps, was extruded in India during the Late Cretaceous and early Cenozoic, as McLean (1985) pointed out, but Courtillot and others (1986, p. 362) stated that the ages of the basalt flows, range from 85 to 30 Ma, and, thus, the Deccan trap volcanism probably would not produce an extremely sharp iridium anomaly at 66 Ma.

There are other objections to the idea that iridium at the K-T boundary was of basaltic volcanic origin. As is well known, basaltic volcanism is not explosive, and the products of such volcanic activity are not transported far downwind from their eruption sites. Izett (1981) studied regionally distributed upper Cenozoic volcanic ash beds of the Western United States and nearly all are of silicic composition. For example, no basaltic ash beds have been found in upper Cenozoic rocks of the Western United States that are far downwind from the middle Miocene Columbia River basalt field. In brief, particulate airborne products of basaltic volcanism are found only near their vents.

A second objection to the idea that the K-T iridium anomaly is volcanic in origin is that relatively large amounts of shocked quartz grains have been found globally at the K-T boundary. These shocked quartz grains cannot be a product of basaltic volcanism, and, as will be argued elsewhere in this report, the particular types of shocked quartz in the K-T boundary impact layer cannot be of volcanic origin. Even if the shocked quartz were of volcanic origin, there is no known volcanic process that could distribute the large grains (about 100 micrometers) globally. Thus, the hypothesis of Alvarez and others (1980) that the iridium at the K-T boundary is of extraterrestrial origin is reasonable and in accord with Occam's razor in light of the established impact record of the Earth (Shoemaker, 1979).

Origin of the K-T Boundary Impact Layer

Evaluation of stratigraphic information that bears on the origin of the K-T boundary impact layer might lead to the conclusion that the claystone is formed entirely of air-fall, impact-derived material. The chief reason for this conclusion is that the impact layer forms a persistent unit about 5 mm thick that can be identified at 20 scattered localities in the Raton Basin. About 750 km north of the Raton Basin, the impact layer is well developed at the Teapot Dome, Wyo., locality. Farther north of the Raton Basin (1,000 and 1,500 km) at Brownie Butte, Mont., and along Red Deer Valley north of Calgary, Canada, the impact layer was first identified by me in samples collected by the author, B.F. Bohor, and D.J. Nichols. Thus, the widespread and thin nature of the layer seems to suggest that it is composed of air-fall material.

In contrast, observational data suggest that the impact layer has a complex origin and is composed of no more than 2.0 percent by volume of impact materials, principally shock-metamorphosed minerals. My observation that the impact layer is composed of finely laminated claystone implies that it did not form by simple diagenetic alteration of coarse-grained impact glass but rather by slow accumulation of clay particles. The texture of much of the impact-layer claystone, as determined by SEM analysis, is similar to smectite clays and is not pervasively microspherulitic like the boundary claystone. This fact also suggests that the claystone was not derived from volcanic or impact glass, if the contention of Pollastro and Pillmore (1987) is correct that the alteration of glass results in a microspherulitic texture. The occurrence of macerated plant detritus and twigs along bedding laminae implies, but does not prove, that the claystone accumulated slowly by fluvial or paludal (not air-fall) depositional processes. The presence in the impact layer of pellets of claystone that are texturally similar to the underlying boundary claystone suggests that it is composed of reworked material.

The trace-element content (platinum-group metals excluded) of the impact layer indicates that it does not contain a large component of asteroidal material. The low content of nickel and cobalt, in particular, are relevant to this conclusion.

In summary, observational evidence seems to suggest that the claystone of the impact layer formed principally from locally derived sediment and was not formed from altered volcanic or impact glass. The claystone only contains a minor fraction of impact-derived material, principally shock-metamorphosed minerals. A depositional model that fits the observational evidence is as follows. Vaporized and melted asteroidal material and shock-metamorphosed target rocks were produced during the K-T boundary impact event. Fine-grained shocked mineral grains and an iridium-rich aerosol were transported away from the impact point by upper stratosphere winds. Only a sprinkling of shock-metamorphosed minerals, some of which were relatively large (0.5-0.6 mm), and the iridium-rich aerosol fell throughout much of North America, but the materials were only preserved in exceptionally low-energy, colluvial-fluvial environments in coal basins that existed in the Raton Basin of Colorado and New Mexico and elsewhere in the Western Interior of North America. The materials fell on a paleosurface that existed at the top of the K-T boundary claystone. Locally, plants grew on this surface, and it may intermittently have been covered with water. The hydrologic regime in the Raton Basin may have been slightly upset by climate change as a result of the impact event, which triggered a new but amazingly gentle erosional-depositional cycle. During this cycle, shock-metamorphosed minerals were reworked gently into a slowly accumulating deposit composed of clay and decaying plant material. Because there is only one spike of shock-metamorphosed minerals found to date in the K-T boundary interval, there is no physical evidence to support the idea that there were multiple impacts during this time (see Hut and others, 1987). The model outlined above suggests that the K-T boundary impact event was considerably smaller than that proposed by Alvarez and others (1980).

ROCKS DIRECTLY OVERLYING THE K-T BOUNDARY IMPACT LAYER

A low-sulfur bituminous coal, informally called the "K-T boundary coal" bed (fig. 12), directly overlies the impact layer at most K-T boundary localities in the Raton Basin and at other localities in Western North America (Pillmore and others, 1984; Izett and Bohor, 1986a). The coal bed is generally 4-16 cm thick in the Raton Basin. The association of a coal bed with the K-T boundary claystone provides an important tool for prospecting for new K-T boundary sections. The coal beds in the Raton Basin are mostly bituminous A to B in rank (Pillmore and others, 1984). Fracture surfaces in the coal are commonly coated with jarosite.

The contact between the impact layer and the overlying boundary coal bed occurs within a 1-mm-thick interval and separates laminated claystone from coal (fig. 1). At the Raton, N. Mex., locality (fig. 2), the impact layer and boundary claystone rest on slightly carbonaceous claystone and are separated from the boundary coal bed by an intervening greenish-gray claystone 12-23 cm thick. This locality is the only place in the Raton Basin where the boundary claystone and impact layer are not directly overlain by coal. However, a unit of intervening claystone 4 to 7 cm thick occurs between the impact layer and the K-T boundary coal (lignite) bed at the Dogie Creek, Wyo., site (Bohor and Triplehorn, 1987).

The trace-element chemical content of the coal was determined by the ICP method, and the results are listed in table 4. Inspection of the data shows that the coal is impoverished in the analyzed elements as compared to other boundary interval rocks, except for zinc.

The iridium content of the boundary coal can nearly equal that of the underlying impact layer at some boundary sites and can exceed that in the impact layer at other sites. For example, the iridium content of the impact layer is only 1.2 ppb at the Raton, N. Mex., site. In contrast, the boundary coal contains about 10.9 ppb iridium at the Madrid, Colo., site (fig. 8). One explanation for the iridium anomaly in the coal is as follows. During diagenesis of impact-layer sediments, iridium was dissolved and moved into the adjacent coal where it was concentrated. An argument could be made that impact materials were moved upward from the impact layer into the coal by mechanical sedimentary processes, but there is no evidence for such a reworking event. In particular, the coal is generally free of quartz and other clastic mineral grains.

CLASTIC MINERAL GRAINS OF THE K-T BOUNDARY IMPACT LAYER

Non-clay, clastic, mineral-grain concentrates of the K-T boundary impact layer from Western North America K-T boundary sites were prepared to determine their mineralogy. I was particularly interested in the abundance and mineralogy of shock-metamorphosed minerals in K-T boundary rocks at these sites relative to the abundance and mineralogy at sites outside North America. I reasoned that if reliable estimates could be determined, these data might be used to narrow the search for the location of the K-T boundary impact. The method used was as follows. The exact stratigraphic location of the shock-metamorphosed materials in the K-T boundary interval in the Raton Basin was found by studying polished specimens of boundary interval rocks (Izett and Pillmore, 1985a). Grains were plucked from the polished surfaces of the claystone and some of the grains were found to contain multiple intersecting sets of closely spaced planar lamellae indicative of a shock-metamorphic origin. Subsequently, shock-metamorphosed grains were identified in thin sections, and gram-sized samples of shock-metamorphosed material were obtained for study by dissolving kilogram-sized samples of the impact layer.

Preparation of a clay-free concentrate of clastic mineral grains from the impact layer was accomplished by (1) collecting pieces of the boundary interval composed of both the boundary claystone and the impact layer, (2) scraping off the impact layer from the top of the boundary claystone with a razor blade, (3) pulverizing the material in water in a cocktail blender, and (4) dissolving this slurry in a hydrofluoric acid solution (about 5 percent) accompanied by ultrasonic cleaning or scrubbing. This slurry was again ultrasonically cleaned, but this time in an aqueous solution of Clorox (about 30 percent). The residuum was further reduced in volume by dissolution of the remaining clay minerals in a dilute (about 5 percent) hydrofluoric acid solution.

Study of the assemblage of mineral grains separated from the impact-layer claystone by the above process was accomplished by using both a stereomicroscope and a petrographic microscope. Some shock-metamorphosed quartz grains can be recognized with a hand lens or a stereomicroscope because they appear milky and exhibit closely spaced planar lamellae similar in some respects to polysynthetic twin lamellae in plagioclase. In some instances, the hydrofluoric acid selectively etches the planar lamellae, making them much more obvious. Grains suspected to be of shock-metamorphic origin because of their appearance under 100X magnification of a stereomicroscope were mounted on the sharpened tip of a steel wire spindle. The mounted grains were then inserted into the immersion cell of a Wilcox spindle stage (Wilcox, 1959). This device is ideal for the search and study of shock-metamorphic features in minerals because planar lamellae can be rotated vertical and their orientation in relation to optic directions can be measured and plotted on a stereonet. Using the spindle stage, the precision of measuring planar features and optical directions is about $\pm 2^{\circ}$ (Wilcox, 1959).

Microscopic study of the residuum of mineral grains obtained in this way showed that it consisted of (1) single grains of quartz, oligoclase, and potassium feldspar; and (2) composite grains (lithic fragments) composed of quartzite, metaquartzite, chert, and granite-like aggregates of quartz, oligoclase, and potassium feldspar. Extremely rare grains of zircon, garnet, epidote, and blue-green hornblende were seen in some concentrates. In addition, some grains of authigenic gypsum, barite, and jarosite survived the original acid digestion. These minerals were removed by further acid digestion (HCl and H_2SO_4).

Shock-Metamorphosed Minerals

About 40 percent of the clastic mineral grains in the impact layer are quartz and about 60 percent of the grains contain multiple sets of planar features variously termed "shock lamellae" by Chao (1967), "planar features" by Carter (1965), "planar elements" by Stöffler (1972), and "microfractures" by Chao and Goresy (1973, p. 291). Transmission electron microscopy has shown that many planar features in minerals that have experienced only low shock are open microfractures (Chao and Goresy, 1973, fig. 3). However, the planar features in quartz that have undergone moderate to high levels of shock are composed of glass termed "thetomorphic" by Chao (1967). Unshocked grains appear structureless and show normal sharp extinction, but a few contain Böhm lamellae and exhibit undulatory extinction. As seen under a petrographic microscope using plane-polarized light, the shocked quartz grains are colorless to pale light brown, and some contain numerous opaque inclusions whose origin is unknown (fig. 20C). These inclusions have an irregular shape, and some are arranged along poorly defined planes. In addition to the opaque inclusions, zircon and biotite were seen in a few quartz grains. No quartz

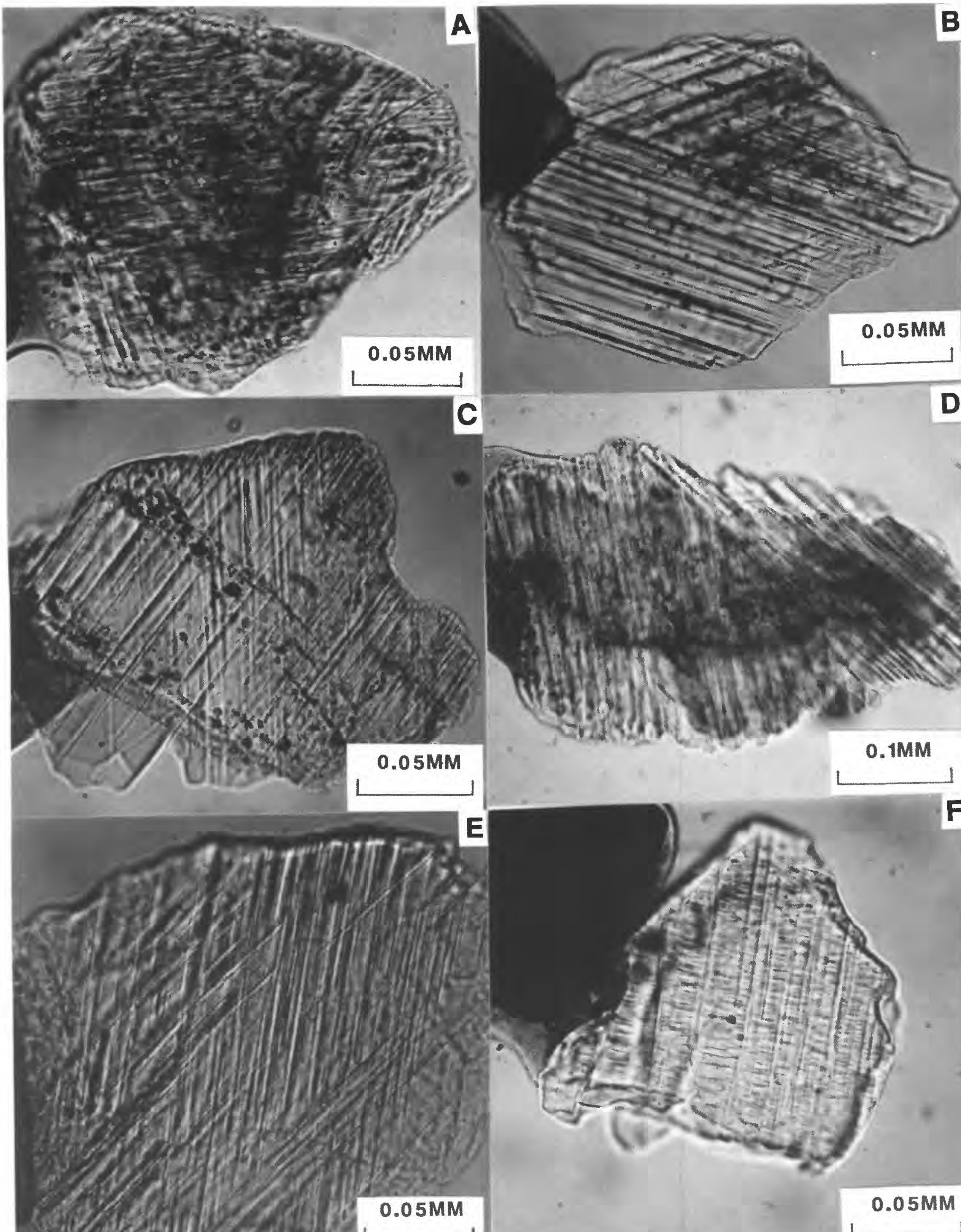


Figure 20.--Photomicrographs of shock-metamorphosed quartz grains from the K-T boundary impact layer. A, Grain from the Clear Creek North, Colo., locality, showing four sets of shock lamellae; B, Grain from the Brownie Butte, Mont., locality, showing two sets of shock lamellae; C, Grain from the Starkville South, Colo., locality, showing two sets of shock lamellae and numerous opaque inclusions; D, Grain 0.42 mm long from the Clear Creek North, Colo., locality, showing two sets of shock lamellae; E, Grain from the Starkville South, Colo., locality, showing two sets of shock lamellae; F, Grain from the Clear Creek North, Colo., locality, showing two closely spaced sets of shock lamellae. Photographed with plane-polarized light.

grains separated from the impact layer were seen to contain glass inclusions that are commonly found in silicate phenocrysts of pyroclastic volcanic rocks (Anderson, 1976).

Shock-metamorphosed quartz grains are invariably cloudy and exhibit a pearly luster, and these features provide a clue to their shock-metamorphic origin if no planar lamellae are visible for a particular orientation of the grain. Microstructures, such as Böhm lamellae, in quartz of tectonites might be confused with shock-induced microstructures. However, Böhm lamellae are far less regular and generally occur in single sets, rarely in two sets, and never in three or more sets. Bunch (1968, p. 419) stated that "The occurrence of two or more sets of planar features in quartz grains appears to be unique to impact sites." It is important, therefore, that shock-metamorphosed quartz grains in K-T boundary rocks typically average about three sets and some grains have as many as nine sets. Figure 20 shows six typical shock-metamorphosed quartz grains separated from the impact layer in the Raton Basin and at Brownie Butte, Mont. Such quartz grains are typical and common in the impact layer, and they provide a large amount of shock-metamorphosed material for study.

Quartzite and metaquartzite comprise about 30 percent of the grains in the impact layer in the Raton Basin. The distinction between quartzite and metaquartzite is somewhat arbitrary. Grains classified as quartzite are distinctly granular, have regular grain borders, and are fine grained (0.05 to 0.1 mm). Some of these grains or lithic fragments exhibit shock-metamorphic features. Individual quartz optical units that comprise a lithic fragment contain multiple intersecting sets of shock lamellae. Typical grains of shock-metamorphosed quartzite and metaquartzite from the impact layer are shown in figures 21 and 22. Some grains classified as metaquartzite are very fine grained (0.01 mm) and have optical boundaries that vary from curved to highly sutured. Some grains are slightly micaceous, foliated, and similar to sericitic quartz schist. Shock lamellae in each optical unit parallel crystallographic planes of the quartz lattice and terminate at optical grain boundaries. These observations are in accord with data from quartz at meteor impact sites (Carter and Friedman, 1965). Such grains also have been seen in K-T boundary mineral concentrates from Western North American sites and at the Caravaca, Spain, and Stevns Klint, Denmark, sites in Europe (fig. 23).

About 10 percent of the quartzite and metaquartzite grains appear to contain shock lamellae, but this is only a rough estimate because of our inability to rotate grains while point counting thin sections. The remainder of the quartzite and metaquartzite grains do not appear to contain shock lamellae and thus might be interpreted as non-impact clastic mineral grains incorporated in the impact layer during mixing and reworking with local sediment. However, the abrupt appearance of relatively large amounts of these materials (fig. 24) in the impact layer in contrast to their paucity in directly underlying rocks argues for an unusual event. This abrupt appearance of quartzite and metaquartzite grains at each K-T boundary site in the same microstratigraphic unit in the Raton Basin suggests that the provenance of these grains could not be from the normal erosion of older rocks (fig. 25).

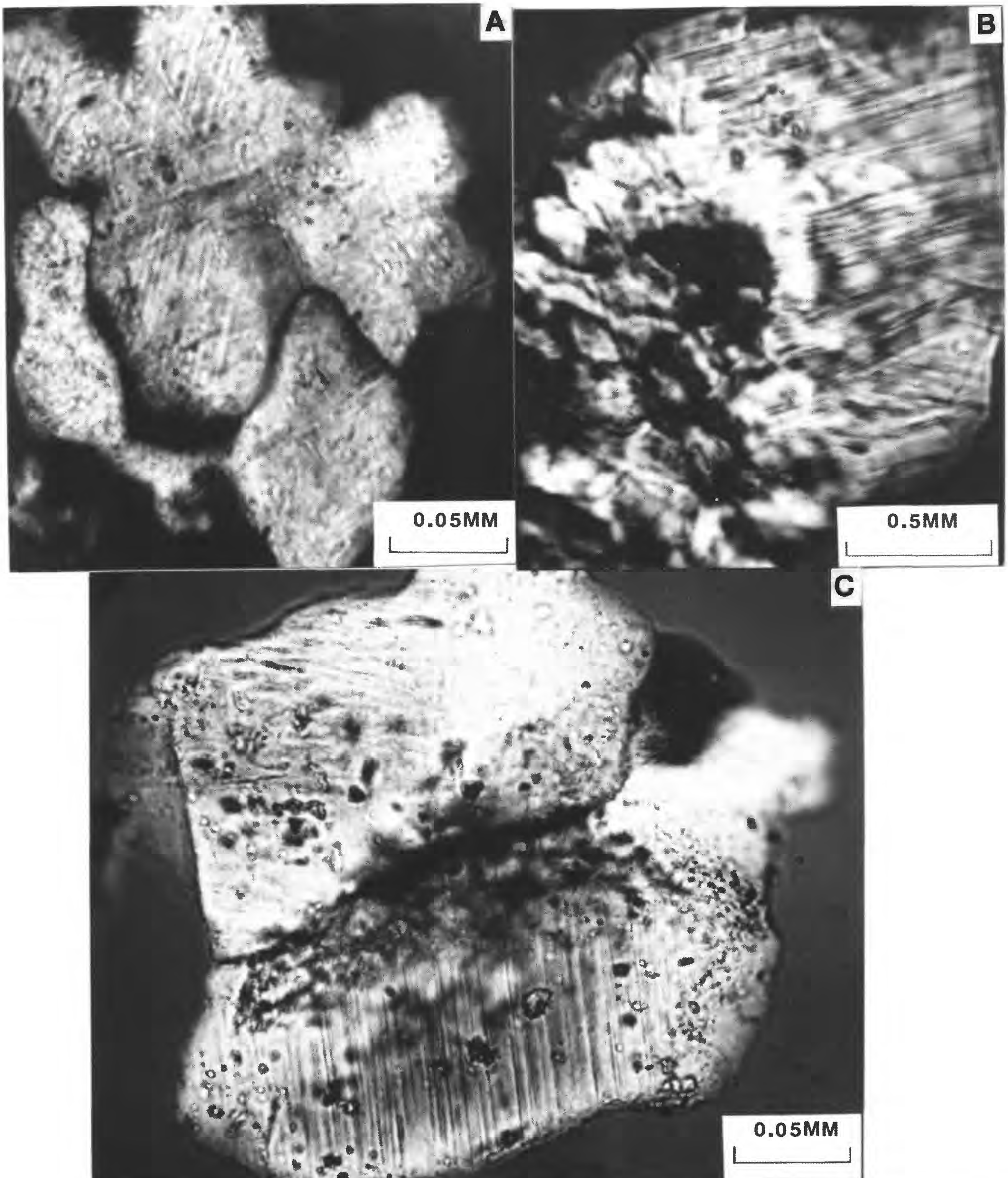


Figure 21.--Photomicrographs of shocked quartzite or metaquartzite? grains from the K-T boundary impact layer. A, Quartzite or metaquartzite? grain from the Berwind Canyon, Colo., locality. B, Quartzite grain from the Brownie Butte, Mont., locality composed of one large quartz optical unit and numerous smaller optical units. C, Quartzite grain from the Clear Creek North, Colo., locality composed of several optical units. Photographed with cross-polarized light.

A)

B)

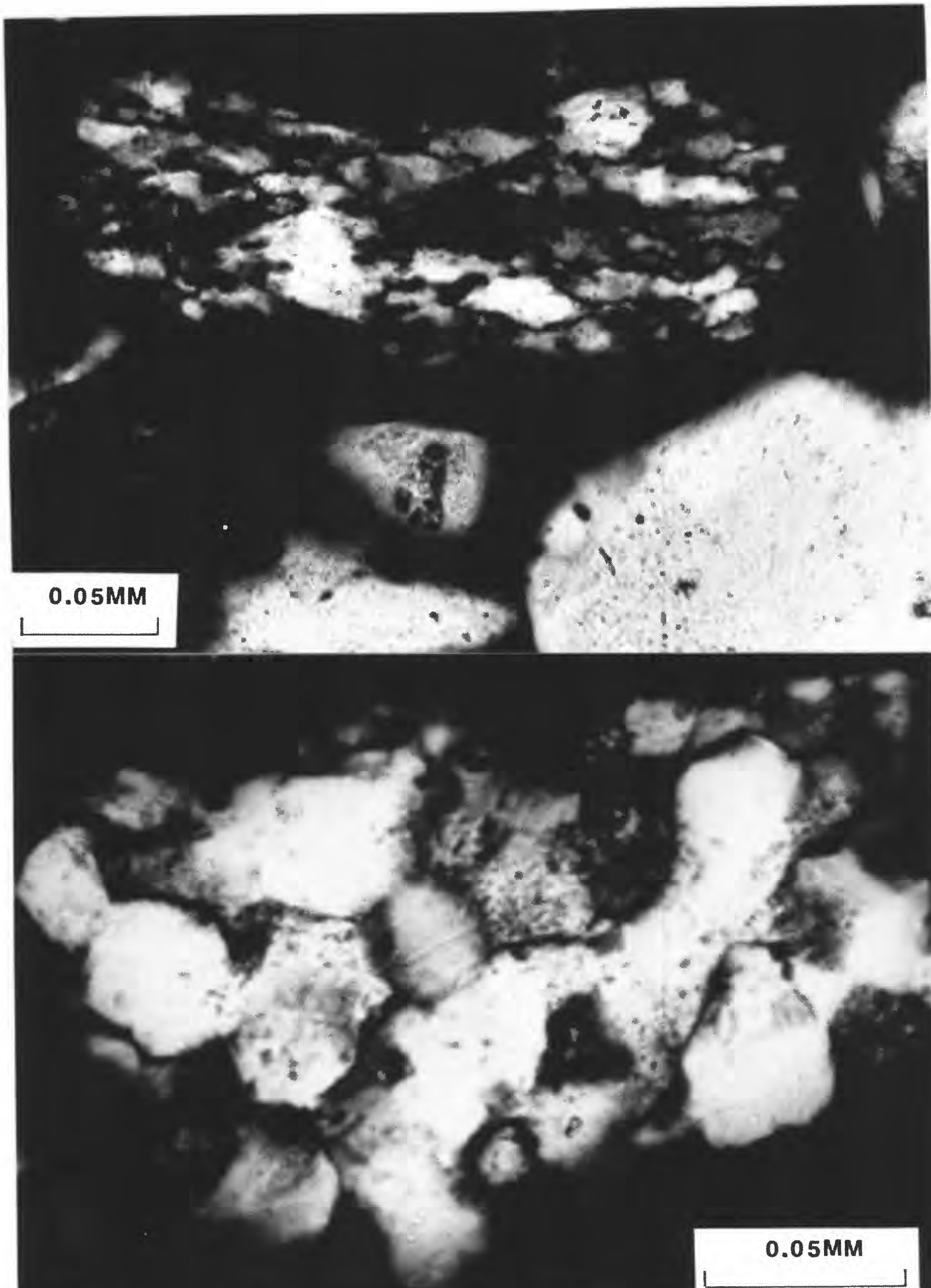
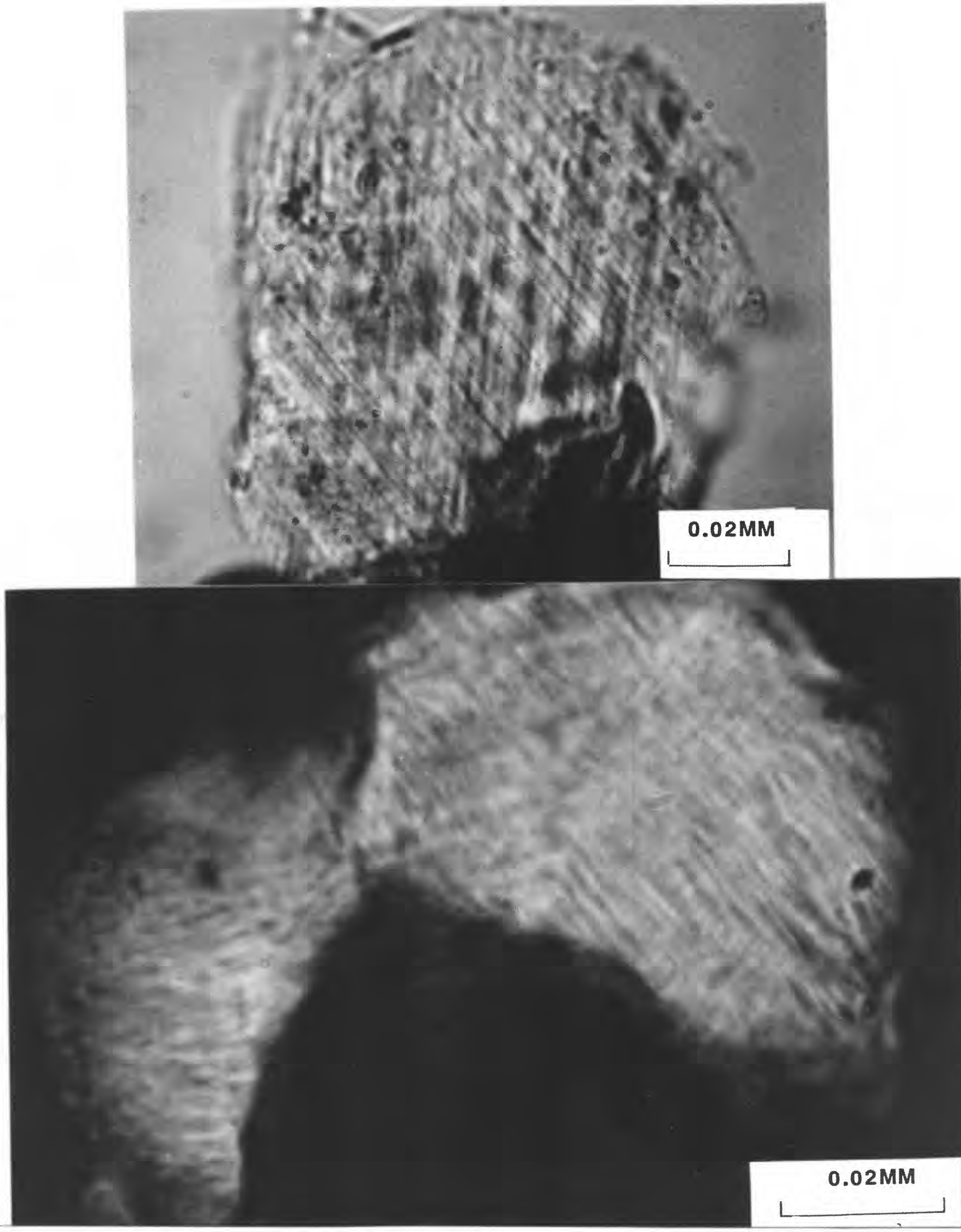


Figure 22.--Photomicrographs of shock-metamorphosed metaquartzite and quartzite? grains. A, Metaquartzite grain from the Berwind Canyon, Colo., locality showing foliated nature of grain. B, Metaquartzite or quartzite? grain. Under high magnification, multiple sets of shock lamellae can be seen in optical units of each grain. Photographed with cross-polarized light.



A)

B)

Figure 23.--Shock-metamorphosed quartz and quartzite grains from the K-T boundary impact layer at Stevns Klint, Denmark. A, Quartz grain showing two sets of shock lamellae. B, Quartzite grain containing three optical units and each of these contains multiple sets of shock lamellae. Photographed with cross-polarized light.

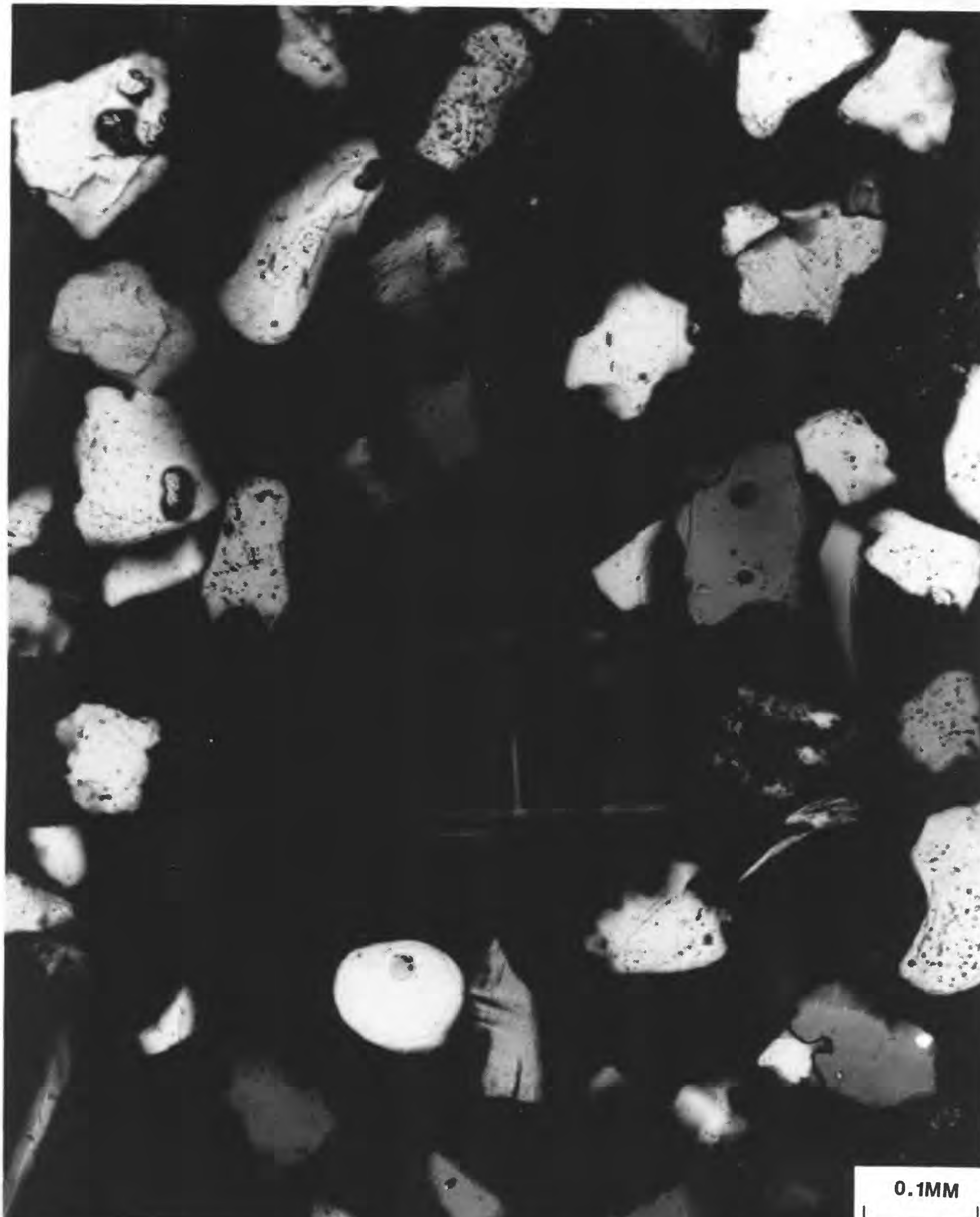


Figure 24.--Photomicrograph of a doubly polished thin section of clastic mineral grains from below the K-T boundary claystone. Grains from carbonaceous shale a few millimeters thick that directly underlies the boundary claystone at the Berwind Canyon, Colo., locality. Assemblage includes mostly quartz and a small number of quartzite and chert grains, none of which contain shock-metamorphic features. Scale indicated by large microcline grain near center of field which is 0.27 mm in diameter. Photographed with cross-polarized light.

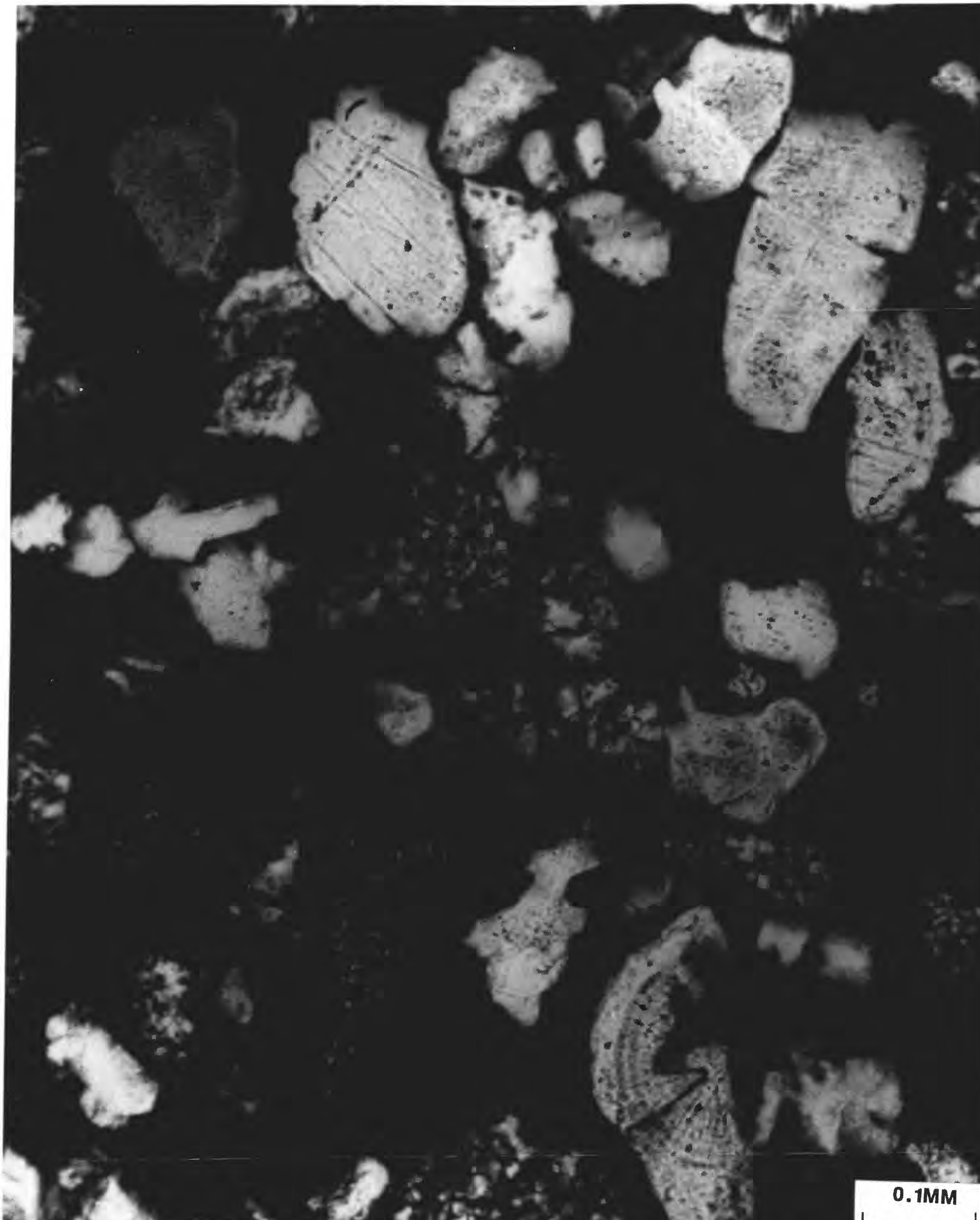


Figure 25. Photomicrograph of a doubly polished thin section of clastic mineral grains from the K-T impact layer. Grains from the same locality (Berwind Canyon, Colo.) as those in figure 24. Several shock-metamorphosed quartz grains are in upper part of photomicrograph. Assemblage of grains is markedly different than those shown in figure 24. Notice the high percentage of quartzite, metaquartzite, and chert grains. Unusual grain showing circular-radial structure in lower right corner of photomicrograph. Photographed with cross-polarized light.

Chert composes about 30 percent of the grains in the impact layer, and it occurs in two forms. One form is a texturally uniform microcrystalline aggregate of interlocking granules about 0.01 mm in diameter. A second form is composed of generally spherical grains of material here classified as chalcedony. When these grains are viewed under a petrographic microscope using cross-polarized light, they display a dark, pseudo-uniaxial interference figure that rotates as the microscope stage is turned. Shock lamellae do not occur in any of the chert grains in the impact layer. The abundance of chert in the impact layer in the Raton Basin in relation to that in underlying rocks argues for an unusual event to explain its abrupt appearance (figs. 24 and 25). This abrupt appearance of appreciable amounts of chert at each site and in the same microstratigraphic horizon in the Raton Basin and at other sites in Western North America suggests that the provenance of the chert could not be from the normal erosion of older rocks.

An exceedingly rare type of shock-metamorphosed lithic fragment composed of an aggregate of sodic plagioclase, potassium-rich feldspar, and quartz occurs in the impact layer in the Raton Basin (fig. 26). This composite-type of shock-metamorphosed grain suggests that the K-T impact target rock included a small amount of granitic rock.

Plagioclase and a potassium-rich feldspar compose less than a few percent of the grains in the impact layer (fig. 27). Optical properties of the shocked plagioclase grains, including their indices of refraction, optic angles, cleavage planes, extinction angles, and shock lamellae orientations, were determined in the immersion cell of a spindle stage designed by Wilcox (1959), and they were plotted on a stereonet. The optical data indicate that the plagioclase grains are mainly oligoclase. The potassium feldspar grains are low in sodium as judged by their refractive indices and by electron microprobe analyses. Some potassium-rich feldspar grains show typical albite-pericline grid twinning, which indicates that they are triclinic microcline. A few of the potassium feldspar grains studied contain several sets of shock lamellae and others show shock mosaicism. Another indication that the potassium feldspar is shock-metamorphosed is that its refractive indices and optic angle are slightly lowered below their lowest normal values. The optic angle ($2V_x$) of one potassium feldspar was about 30^o. The composition of the feldspars suggests they were derived from silicic, rather than basic, rocks at the impact site.

In summary, several different mineralogical types of mineral grains in the K-T boundary impact layer show shock-metamorphic features and are thought to reflect the composition of target rocks at the K-T impact site. Judging from the large amount of shock-metamorphosed quartzite, metaquartzite, and quartz in the impact layer, the target rocks were probably continental sedimentary and metasedimentary rocks. A minuscule amount of shock-metamorphosed oligoclase, microcline, and granite-like mixtures of feldspar and quartz suggest that the target rock was, in part, granitic. In contrast, many grains in the K-T boundary impact layer do not show shock-metamorphic effects. The source rocks for these grains is clearly different than the source rocks for subjacent sedimentary rock units. For example, lithic grains of quartzite, metaquartzite, and chert constitute only a few percent of the grains in the carbonaceous shale, mudstone, and sandstone that lie below the K-T boundary claystone (fig. 24). Yet, in the impact layer, these grains are a major component (fig. 25). This remarkable change takes place at the same stratigraphic position at all K-T boundary sites in the Raton Basin and at boundary sites in Montana and Wyoming and suggests that these unshocked minerals are also impact derived.

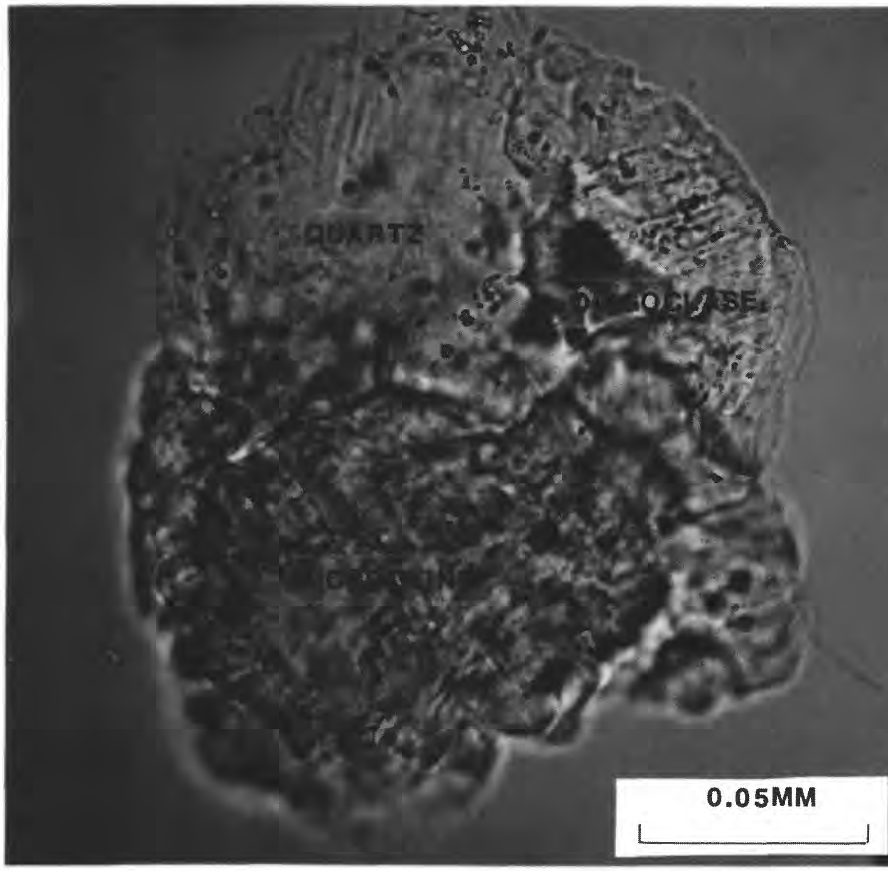


Figure 26.--Photomicrograph of a shock-metamorphosed granite-like grain. Such grains composed of quartz, oligoclase, and microcline are rare and indicate that the rocks at the K-T boundary impact site were in part granitic in composition. Faint shock lamellae in quartz and oligoclase can be seen in the grain oriented in this position. Such lamellae are better seen at high magnification when the grain is rotated so lamellae are vertical. Grain separated from the K-T boundary impact layer at the Clear Creek North, Colo., locality. Photographed with plane-polarized light.

A)

B)

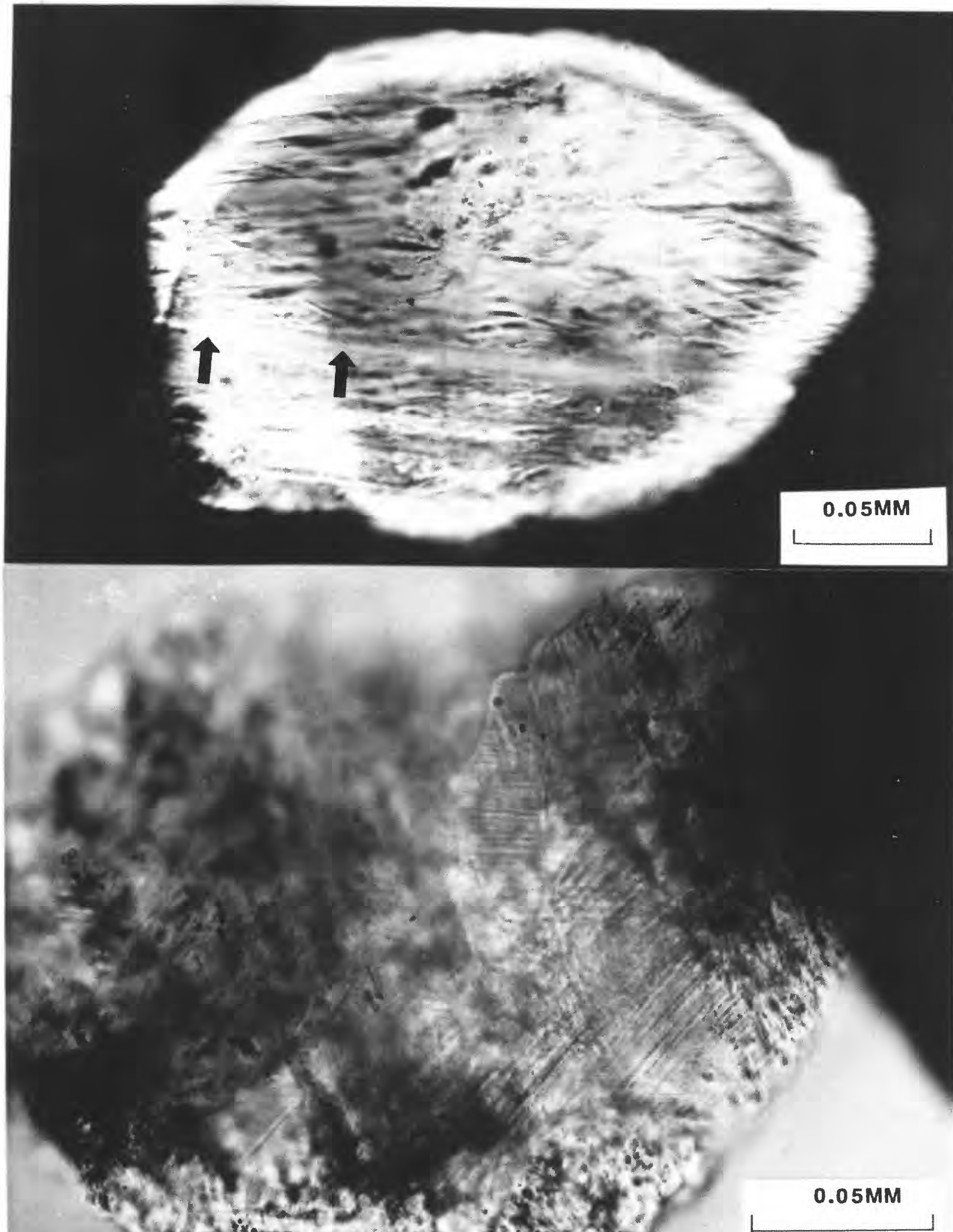


Figure 27.--Photomicrograph of shock-metamorphosed feldspar grains from the K-T boundary impact layer. A, Oligoclase grain from the Brownie Butte, Mont., locality showing two prominent sets of shock lamellae. Albite twin lamellae shown at arrow. Photographed with cross-polarized light. B, Microcline grain from the Brownie Butte, Mont., locality showing two prominent sets of shock lamellae. Photographed with plane-polarized light.

SHOCK-METAMORPHIC FEATURES IN K-T BOUNDARY MINERALS

Shock-metamorphic effects in minerals of K-T boundary rocks were first recognized by Bohor and others (1984). This discovery provided important physical evidence confirming that a large asteroid struck Earth at the close of the Cretaceous Period as suggested by Alvarez and others (1980). The evidence of shock-metamorphism reported by Bohor and others (1984) consisted of multiple intersecting sets of planar features in quartz grains in K-T boundary rocks. The planar features can be seen readily using a petrographic microscope and plane-polarized light, and they give the quartz a remarkable appearance (fig. 20) that is far different from the unshocked quartz of tectonites. Petrographers unfamiliar with the appearance of shock-metamorphosed quartz might be hesitant to sight-identify the mineral as quartz without determining its optical and X-ray properties, because of its highly unusual appearance. For those who have studied shock-metamorphosed rocks, the conclusion that the planar features in K-T boundary minerals are of shock-metamorphic origin is unambiguous. The microscopic properties of shock lamellae are best seen when they are oriented nearly parallel to the axis of a microscope. In this orientation, the lamellae appear as dark lines bordered by parallel bright lines.

In some grains, the lamellae are partly defined by closely spaced, micrometer-sized inclusions that can be seen at 1,250 magnification using an oil-immersion objective of a petrographic microscope (fig. 28). Some inclusions are thought to be filled with a fluid because they contain a minute bubble that moves with Brownian motion. Such fluid inclusions oriented like a string of beads along the planar features were termed "decorated lamellae" by Carter and Friedman (1965) and Robertson and others (1968). In grains having individual shock lamellae that are widely spaced, the fluid inclusions are relatively large. In grains having closely spaced, fine lamellae, the fluid inclusions are relatively small and barely can be discerned. In many shock-metamorphosed grains, the presence or absence of fluid inclusions along the lamellae cannot be confirmed because their size may be below the resolving power of the microscope. In short, a complete gradation probably exists in fluid inclusion size from those that hardly can be resolved microscopically to those that may be present but are invisible at 1,250X.

The origin of the fluid inclusions in shock-metamorphosed minerals at meteor impact sites was attributed by Stöffler (1972) to post-shock annealing in a high-temperature environment. This interpretation for the origin of the fluid inclusions in shock-metamorphosed K-T boundary minerals is inconsistent with the fact that K-T boundary minerals probably did not undergo high-temperature annealing at the impact site. Instead, shock-metamorphosed minerals were lofted instantaneously into the atmosphere, quenched, and transported to their ultimate depositional sites without undergoing high-temperature annealing.

Optical measurement of shock lamellae orientations in quartz grains shows that they parallel rational crystallographic planes, and the most common orientation is about 23 degrees from the basal plane and parallel to Omega, the form $(10\bar{1}3)$ (Chao, 1967). Figure 29 is a frequency diagram showing the orientation relative to the basal plane of 200 sets of shock lamellae in quartz grains from the Clear Creek North, Colo., locality (fig. 2). About 50 percent of the total number of measured sets in quartz are oriented parallel to the form Omega $(10\bar{1}3)$. In quartz grains that have only two or three lamellae sets, the planes are more widely spaced than in grains having a large number of differently oriented sets. Although most lamellae are planar, some

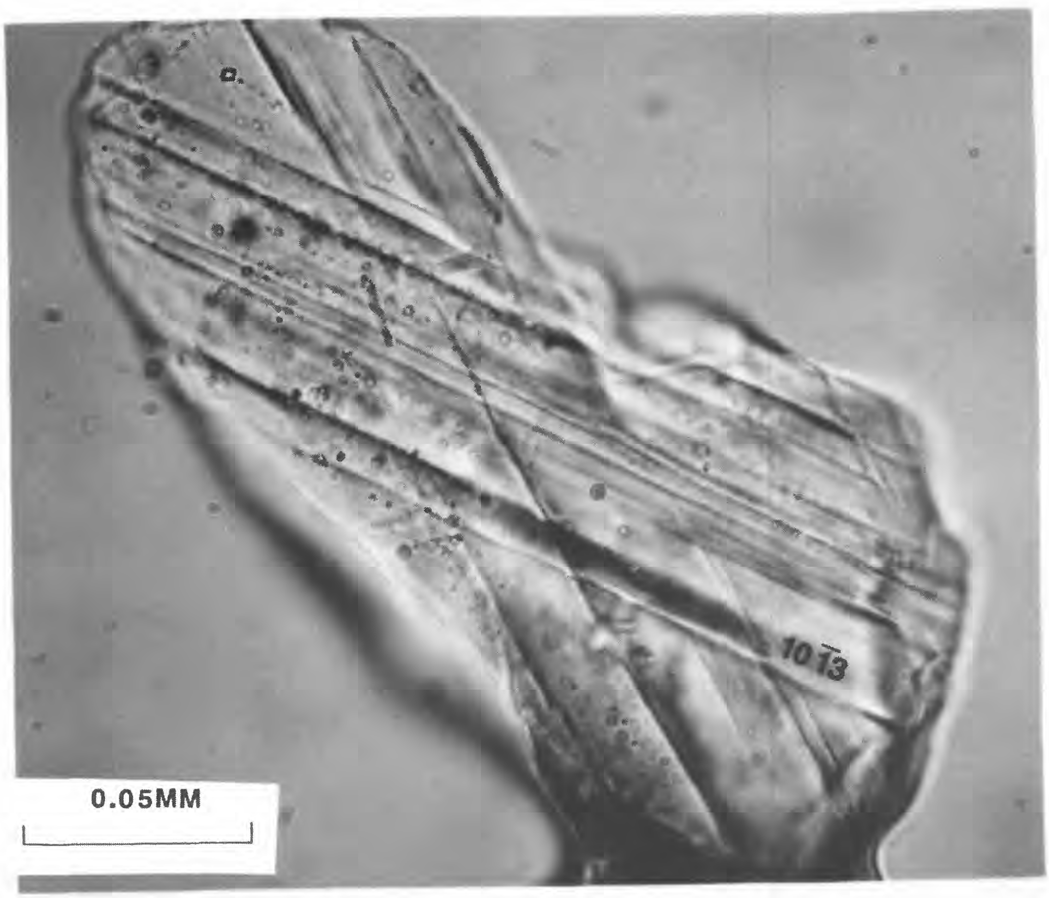


Figure 28.--Photomicrograph of a shock-metamorphosed quartz grain. Two prominent sets of shock lamellae are partly defined by fluid inclusions. Such sets have been termed "decorated lamellae" by Carter and Friedman (1965) and Robertson and others (1968). Grain from the K-T boundary impact layer at the Clear Creek North, Colo., locality. Photographed with plane-polarized light.

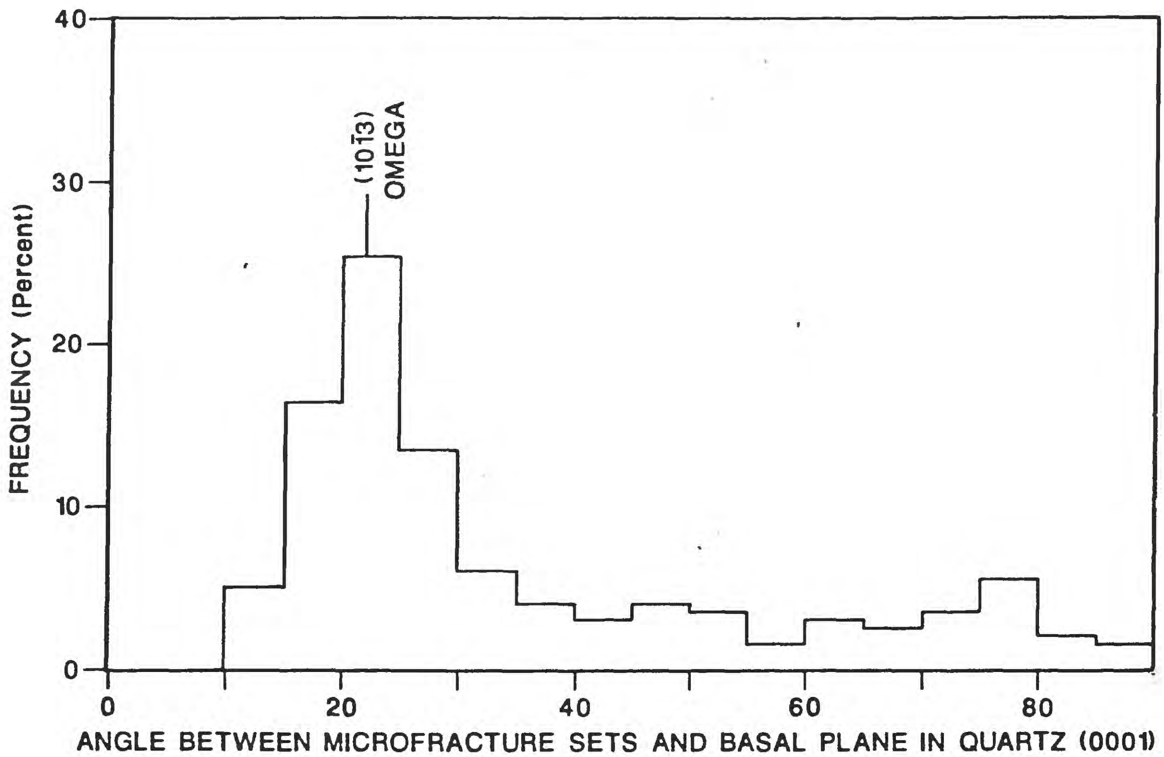


Figure 29.--Frequency diagram showing the orientation of planar shock lamellae in quartz grains. Orientation of 200 planar sets measured in grains from the K-T boundary impact layer from the Clear Creek North, Colo., locality. About 50 percent of the shock lamellae are oriented about 23 degrees from the basal plane and parallel to the form Omega (10$\bar{1}$3).

are gently curved (fig. 30). Lamellae are more likely to be curved in grains having widely spaced, decorated sets.

Another property of silicate minerals that have experienced hypervelocity shock is the lowering of their indices of refraction and birefringence. Chao (1967) reported that in quartz of tectonites the refractive indices are never lowered below their normal values (n_o=1.544; n_e=1.553). Xie and Chao (1985) showed that the mean index of refraction of most shocked quartz in rocks at the Ries impact structure in Germany had been lowered from n_o=1.5485 to n_o=1.46. Bohor and others (1984) reported that the refractive indices of shocked quartz in K-T boundary rocks in Montana are lowered below their normal values. Similarly, in the Raton Basin, the refractive indices of shocked quartz in the impact layer are lowered, but not drastically. Figure 31 shows that only about 30 percent of shock-metamorphosed quartz grains in the impact layer in the Raton Basin have had their refractive indices lowered, but even in these grains the indices of refraction have not been lowered below n_o=1.535. This result is curious in view of the fact that the indices of refraction of shock-metamorphosed quartz in rocks at large impact structures such as the Ries impact structure (25 km in diameter) in Germany have been drastically lowered, as low as quartz glass (n=1.460). The K-T boundary asteroid is thought by some (Alvarez and others, 1980) to be much larger than the one that formed the Ries structure, and, accordingly, one might expect to find marked lowering of the refractive indices and birefringence of quartz in the K-T boundary impact layer. However, the size of the asteroid is not the only factor that controls the degree of shock-metamorphism such as the lowering of the refractive indices of quartz in target rocks. The nature of the target rocks is also critical. For example, the K-T impact may have occurred in an area such as Manson, Iowa, underlain by a thick sequence of sedimentary rocks that may have responded differently to the shock wave than the crystalline granitic rocks that were near the surface in the Ries area.

Yet another optical property of silicate minerals that have undergone hypervelocity shock at meteor impact and nuclear explosion sites has been termed shock "mosaicism" by Stöffler (1972, p. 87). The optical effects include a highly irregular patchy extinction pattern probably involving the formation of disoriented domains in the crystal structure (fig. 32). Shock mosaicism commonly occurs in quartz and feldspar of K-T boundary rocks, not only in grains containing shock lamellae but also in some grains without such features. Again, the K-T impact may have occurred in an area underlain by a thick sequence of sedimentary rocks that may have responded differently to the shock wave than the crystalline granitic rocks that were near the surface in the Ries area.

A few faint X-ray diffraction lines for the mineral stishovite, a high-pressure polymorph of SiO_2, were reportedly found in quartz grains containing multiple sets of shock lamellae by Bohor and others (1984). The existence of stishovite in the quartz grains would provide supporting evidence for the origin of the shock lamellae. E.E. Foord of the U.S. Geological Survey, who identified the faint lines as belonging to stishovite, kindly made the X-ray diffraction films available to me. B.F. Leonard of the U.S. Geological Survey and I examined and remeasured the lines on the films. Contrary to the observations of Bohor and others (1984), we did not find any lines that would suggest the occurrence of stishovite or coesite in the quartz grains.

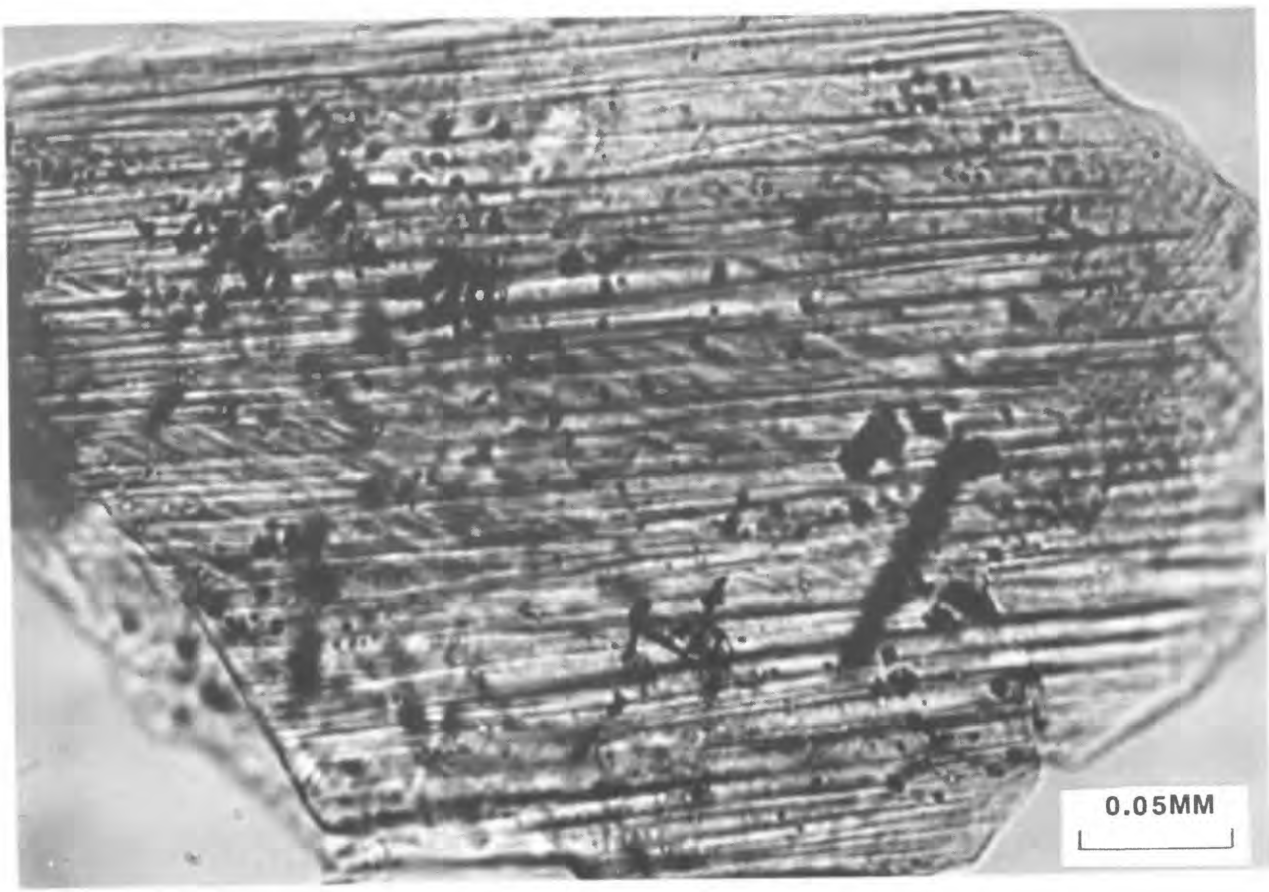


Figure 30.--Photomicrograph of a shock-metamorphosed quartz grain. Three prominent sets of slightly curved shock lamellae can be seen in this orientation of the grain. Notice large opaque inclusions of unknown composition in grain. Grain from the K-T boundary impact layer at the Clear Creek North, Colo., locality. Photographed with plane-polarized light.

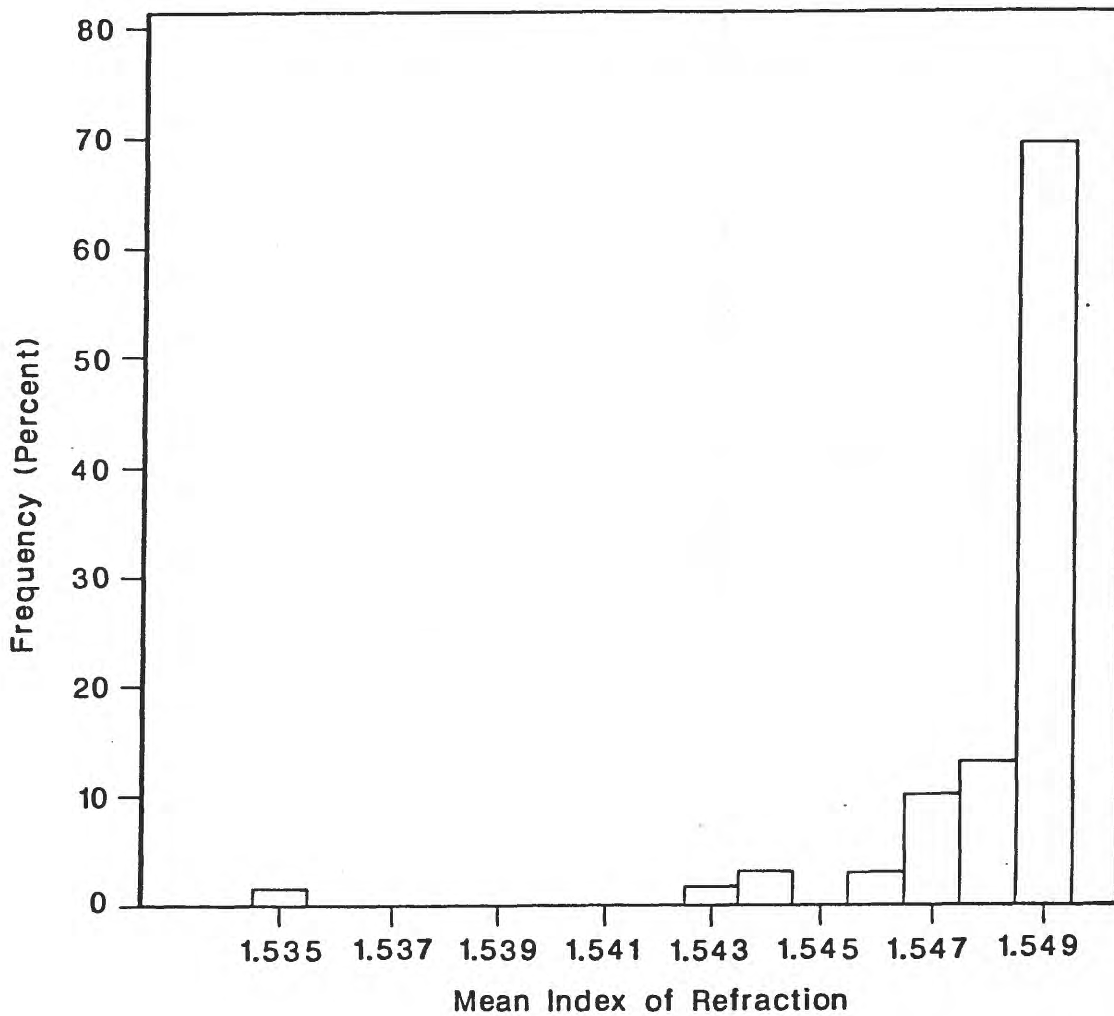


Figure 31.--Frequency diagram showing the average index of refraction of shock-metamorphosed quartz grains. Indices of refraction of 63 quartz grains from the K-T boundary impact layer at the Clear Creek North, Colo., locality were measured. Indices of refraction measured using the focal masking technique described by Wilcox (1983).

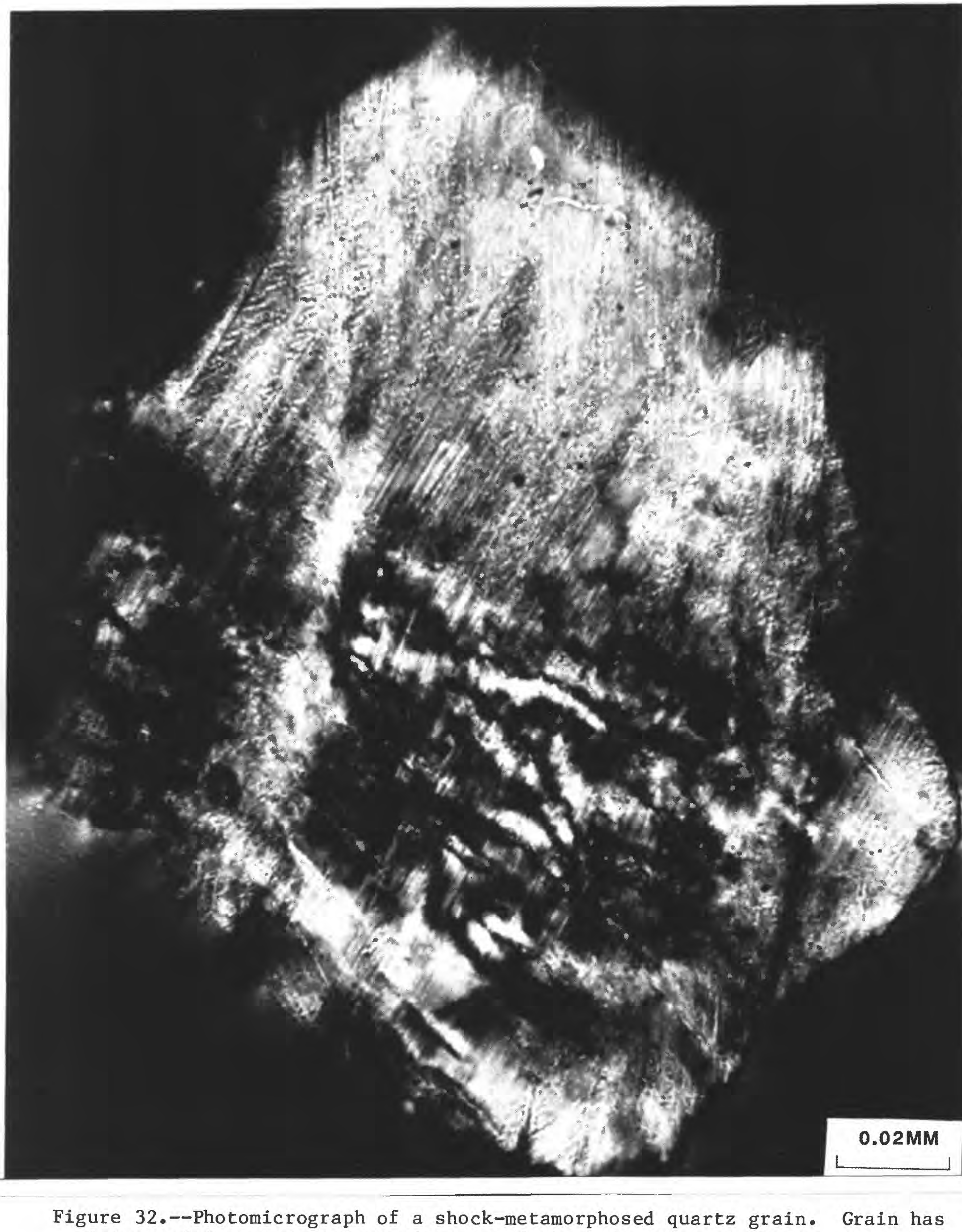


Figure 32.--Photomicrograph of a shock-metamorphosed quartz grain. Grain has a minimum of four and perhaps six sets of shock lamellae in this orientation. An area of anomalous extinction termed shock "mosaicism" occurs in the lower part of the grain. Grain from the K-T boundary impact layer at the Clear Creek North, Colo., locality. Photographed with cross-polarized light.

Finally, shock-metamorphosed quartz in rocks at large meteor impact sites typically contains vitreous material along shock lamellae termed "thetomorphic glass." It forms by shock-induced, solid-state conversion of quartz to glass (Chao, 1967). Thus, thetomorphic quartz glass would be expected to occur along planar features in some shock-metamorphosed quartz grains in K-T boundary rocks. The K-T boundary asteroid is thought by some to be much larger than the one that formed the Ries structure, and one would expect to find thetomorphic glass along shock lamellae, but none was seen during optical studies. Again, the K-T impact may have occurred in an area such as Manson, Iowa, underlain by a thick sequence of sedimentary rocks that may have responded differently to the shock wave than the crystalline granitic rocks that were near the surface in the Ries target area.

ORIGIN OF SHOCK-METAMORPHIC FEATURES IN K-T BOUNDARY MINERALS

Bohor and others (1984) concluded that the multiple sets of planar lamellae in quartz in K-T boundary rocks formed by shock-metamorphic processes during the impact of an asteroid that struck Earth at the close of the Cretaceous. They arrived at this conclusion on the basis of two areas of research: (1) the known impact record of the Earth (Shoemaker, 1979; Short and Bunch, 1968; Grieve, 1982) and (2) experimental and empirical mineralogy of impact materials (Wackerle, 1962; Carter and others, 1964; Carter, 1965, 1971; Chao, 1967, 1968; McIntyre, 1968, Ahrens and Rosenberg, 1968; Bunch, 1968; Carter, 1971; Robertson and others, 1968; Short, 1968; Englehardt and others, 1968; Englehardt and Bertsch, 1969; Stöffler, 1971, 1972). These studies have shown that the most important evidence of shock-metamorphism in silicate minerals is the occurrence of multiple intersecting sets of planar microstructures. Such features have been seen only in silicate minerals that have undergone hypervelocity impact at meteor impact sites and nuclear and high yield explosion sites (Chao, 1967). Carter (1968, p. 463) stated that "These planar features are very different from all other microstructures observed in quartz from rocks deformed either naturally or experimentally under static or dynamic loading conditions." Such features apparently can only form at pressures greater than 60 kb (Short, 1968, p. 203).

Other features indicative of shock metamorphism in silicate minerals are the lowering of their refractive indices (Xie and Chao, 1985), the solid-state conversion of crystalline material to thetomorphic glass (Chao, 1967, p. 195), and the formation of high-pressure polymorphs of quartz such as stishovite and coesite (Chao, 1967). In addition, the formation of mosaic extinction (Stöffler, 1972) in quartz and feldspar is suggestive of shock-metamorphism, but this feature might be confused with anomalous extinction in minerals of tectonites. In summary, there is unequivocal evidence that minerals in the K-T boundary interval were shocked because they contain multiple intersecting sets of planar lamellae. Nevertheless, some have raised the question of whether the planar lamellae could be produced during a large scale pyroclastic volcanic eruption.

Recently, Carter and others (1986) described microstructures in quartz, plagioclase, and biotite in the Toba Tuff of Sumatra and suggested that their formation was a result of volcanic-induced shock metamorphism. They thus prompted a debate (Izett and Bohor, 1986b; Alexopoulos and others, 1987) as to whether shock-metamorphic features can form in phenocrysts in a silicic magma or in silicate minerals in wallrock of magma chambers during explosive pyroclastic eruptions. If the proposal of Carter and his colleagues were correct, it would provide an alternative mechanism for the formation of shock-metamorphic features found in minerals in K-T boundary rocks. Such features occur globally in minerals of K-T boundary rocks (Bohor and others, 1984; Izett and Pillmore, 1985a, 1985b; Bohor and Izett, 1986; Preisinger and others, 1986; Badjukov and others, 1986; Alexopoulos and others, 1987). Two sources of information, one from study of K-T boundary minerals and another from study of pyroclastic volcanic rocks, cast doubt on the idea that shock-metamorphic features can form during violent silicic pyroclastic eruptions.

Information Gained from Study of K-T Boundary Rocks

Information gained during this study of K-T boundary rocks in the Raton Basin suggests that the shock-metamorphic features in minerals originated during hypervelocity impact of an asteroid and not during large-scale silicic pyroclastic eruptions. A seemingly important observation is that euhedral silicic volcanic minerals are absent in the K-T boundary impact layer. It seems highly probable that some of these would have survived diagenesis of the alleged volcanic ash precursor. A second observation is that no shock-metamorphosed quartz or feldspar grains in the impact layer have been seen to contain glass inclusions that are typically found in silicate phenocrysts of pyroclastic rocks (Anderson, 1976; Watson, 1976; Sommer, 1977). About 30 percent of the shock-metamorphosed grains in the impact layer are composite and consist of two or more optical units separated by sutured boundaries. These composite grains or lithic fragments consist chiefly of quartzite and metaquartzite and cannot be primary volcanic phenocrysts.

Information Gained from Study of Pyroclastic Rocks

Carter and others (1986) maintained that shock-metamorphosed silicate minerals can be found in silicic pyroclastic rocks such as the Toba Tuff of Sumatra and, by analogy, other large-volume silicic tuffs such as the Bishop Tuff of eastern California. The most important criterion for the recognition of shock-metamorphism in silicate minerals is the presence of multiple intersecting sets of planar lamellae (Chao, 1967; Bunch, 1968, p. 419). I stress the fact that Carter and his colleagues (1986) did not find multiple sets of shock lamellae in the Toba Tuff. However, they did describe anomalous extinction in quartz similar to the shock mosaicism mentioned by Stöffler (1972). They also described diffuse single lamellae in feldspar that they attributed to shock-metamorphic processes.

Carter and others (1986) proposed that silicate minerals in the Toba Tuff do not contain multiple sets of shock lamellae because the lamellae were annealed out during their residence in the high-temperature regime of a cooling ash-flow tuff. If this reasoning were correct, it would account for the lack of such features in quartz and feldspar of silicic pyroclastic rocks. However, shock lamellae are readily found in shock-metamorphosed quartz and feldspar in the high-temperature suevite breccia of the Ries structure in Germany (Chao, 1967, fig. 1). Grieve (1987, p. 253) stated that impact-generated melt sheets "* * * are heavily charged with shocked and unshocked lithic and mineral clasts (>50%) at the top and bottom." Thus, observational evidence indicates that shock lamellae seem to survive high temperature annealing. A possible explanation for this is that some shock lamellae are actually open microfractures (Chao and Goresy, 1973, figs. 3 and 5; Chao, 1976, fig. 7) as determined by transmission electron microscopy, and it may take complete melting of a grain to erase this type of shock lamellae. Taking the suggestion of Carter and others (1986) at face value that shock lamellae are annealed out of volcanic rocks, the fact that shock lamellae are common in K-T boundary shocked minerals should confirm the non-volcanic origin for such grains.

Carter and others (1986) alternatively proposed that the reason that silicate minerals of the Toba Tuff do not contain multiple sets of shock lamellae is that such lamellae do not form in the high-temperature regime of silicic magma chambers. This question can be tested by study of phenocrysts in pumice lapilli in Plinian air-fall deposits. Pumice lapilli are quenched magma samples, and in air-fall deposits their phenocrysts have not undergone a period of thermal residence in hot ash flows. To address this question, I studied polished thin sections of mineral grain concentrates obtained from air-fall pumice lapilli of the Bishop Tuff of eastern California, the Huckleberry Ridge and Lava Creek Tuffs of Yellowstone National Park, the Bandelier Tuff and Cerro Toledo Rhyolite of the Jemez Mountains of New Mexico, and the Toba Tuff. No anomalous optical features of any kind were seen in the mineral concentrates of the clean pumice lapilli--no undulatory extinction, strain shadows, Böhm lamellae, mosaicism, or multiple intersecting sets of shock lamellae. In brief, the grains are monotonously uniform and structureless, go to extinction sharply, and only contain small glass inclusions typical of silicate phenocrysts of pyroclastic rocks (Anderson, 1976). This evidence confirms the idea of Carter and others (1986) that shock-metamorphic features do not form in phenocrysts in a magma.

Carter and others (1986) assumed that the silicate minerals that contain the alleged shock-metamorphic features are phenocrysts and not minerals accidentally incorporated into the Toba magma or ash flows as they were emplaced. Another test of the idea that the alleged shock-metamorphosed silicate minerals in the Toba Tuff are phenocrysts and not xenocrysts would be to perform electron microprobe analyses on such grains. Hildreth (1977, 1979) showed that phenocrysts in pumice lapilli of the Bishop Tuff, for example, have narrow compositional ranges. Accordingly, analyses of the alleged shock-metamorphosed feldspar grains of the Toba Tuff could be made to determine if their composition matches those of primary plagioclase phenocrysts in pumice samples which are quenched samples of the Toba magma.

Refractive indices of quartz and feldspar grains in the Toba Tuff, alleged to contain shock-metamorphic features, could be measured to determine if they are lower than normal values for these minerals. Lowering of the indices of refraction of quartz is commonly found in shock-metamorphosed quartz (Xie and Chao, 1985), but has never been reported in quartz of tectonites.

Two obvious choices are available to address the question of origin of the xenoliths and xenocrysts in ash-flow tuffs that potentially could contain shock-metamorphic features. The first choice is that the xenoliths and xenocrysts are picked up from the land surface by an ash flow during its turbulent emplacement. The fact that ash-flow tuffs contain xenocrysts is well known. Ross and Smith (1961, p. 35) stated that "A great variety of materials derived from rocks through which the tuff forming magmas made their way to the surface, or that were picked up on the surface by the overriding ash flow have been observed by the authors in the course of studies of many hundreds of thin sections of welded tuffs from different regions of the world." They also stated that "Many tuffs show a wide range in the proportion of anorthite in the plagioclase crystals, and the more calcic ones are alien materials." They (1961, figs. 21, 25, 43, 80, 81, 91) illustrated many examples of such xenoliths and xenocrysts.

Geochronologists who determine the age of mineral concentrates in ash-flow tuffs such as the Toba and Bishop Tuffs also are keenly aware that they generally contain xenocrysts. Naeser and others (1981) summarized the problem of contamination in ash-flow tuffs and distal volcanic ash beds and the problem contamination presents to obtaining reliable ages. For example, feldspar concentrates from ash-flow tuffs of the Bishop Tuff yielded ages of 0.9 to 1.2 Ma (Evernden and Curtis, 1965). Dalrymple and others (1965) became aware that the Bishop Tuff contains older Sierran-age feldspar picked up by the ash flows as they were emplaced. They (Dalrymple and others, 1965), therefore, collected only pumice lapilli, prepared pure mineral concentrates, and refined the isotopic age of the Bishop Tuff at 0.71 Ma (average of six ages). More recently, Izett (1982) made 17 new K-Ar age determinations of minerals from pumice lapilli of the Bishop and redetermined its age at 0.738 ±0.003 Ma. No features attributable to shock-metamorphism were seen in the quartz or feldspar crystals used in this study of the Bishop Tuff.

I also have in my collection numerous high-quality mineral concentrates of upper Cenozoic Plinian pumice deposits of the Western United States, and these are available to those interested in searching for shock-metamorphic features in pyroclastic rocks. In particular, I have many mineral concentrates of the Bishop Tuff, including one consisting of about 500 g of sanidine and smaller amounts of oligoclase, allanite, and zircon separated from a 1-m-diameter pumice boulder of the Bishop Tuff (Izett, 1982).

The second choice for the origin of the xenoliths and xenocrysts in ash-flow tuffs is that they are derived from the walls and conduits of silicic magma chambers. If this idea is correct, the xenoliths and xenocrysts either should be coated with glass of the magma or concentrated in glassy pumice lapilli in ash-flow tuffs. Evidence gathered during study of volcanic ash beds derived from silicic magmas bears on this question. During the last 25 years, I have studied mineral concentrates of hundreds of volcanic ash beds, some of which are not contaminated with locally derived sediment, and air-fall pumice lapilli deposits of late Cenozoic age in the Western United States (Izett, 1981, 1982; Izett and Wilcox, 1982). My study shows that uncontaminated, distal volcanic ash beds do not contain xenocrysts or xenoliths, but only contain glass-mantled, primary phenocrysts that do not exhibit anomalous extinction or shock lamellae. Pristine pumice lapilli also contain only glass-mantled, primary phenocrysts.

Although the idea that shock features might be found in the wallrock of magma chambers and conduits should be further investigated, it is difficult to understand how the great pressures (>60 kb according to Short, 1968, p. 198) and high strain rates (10^6/s according to Chao and Goresy, 1973, p. 291) required to produce multiple sets of shock lamellae could be produced in high-level silicic magma chambers. Nevertheless, thin sections of wallrocks and lithic grains commonly found in near-source Plinian pumice deposits should be studied to determine if they contain shock-induced features.

Finally, it should be pointed out that multiple sets of shock lamellae have not been found in quartz or other silicate minerals from the comprehensively-studied, gigantic Pleistocene ash-flow tuffs (and related Plinian pumice deposits) of the United States, such as the Yellowstone Tuff (Christiansen and Blank, 1972), the Bishop Tuff (Hildreth, 1979), and the Bandelier Tuff (Smith and Bailey, 1966). Furthermore, shock lamellae have not been reported from Oligocene to Pliocene ash-flow tuffs of the United States (Ekren and others, 1971; Byers and others, 1976; Steven and Lipman, 1976), to cite only a few studies. It seems unlikely that the many excellent petrologists who have studied pyroclastic rocks have missed seeing such obvious, startling, and unique features as multiple intersecting sets of shock lamellae in quartz or feldspar.

AMOUNT OF SHOCK-METAMORPHOSED MINERALS IN THE IMPACT LAYER

The amount of clastic mineral grains in the 5-mm-thick impact layer was determined by point counting ultrathin, doubly polished thin sections. The counting showed that the total percentage of quartzite, metaquartzite, quartz, and chert grains is variable, but at some localities the total was as much as 2 percent by volume. The proportion of these grains that contain shock-metamorphic features is difficult to ascertain in thin sections or oil-immersion mounts because some grains may not be oriented so that their shock lamellae can be seen and counted. Despite this difficulty, an effort was made to determine the minimum percentages of shock-metamorphosed minerals by point counting concentrates of the impact layer in oil-immersion grain mounts. On the basis of counting about 3,000 grains, it is estimated that about 30 percent of the mineral grains in the K-T boundary impact layer show definite effects of shock-metamorphism; that is, they contain multiple intersecting sets of planar lamellae. Although many grains do not contain shock lamellae, some exhibit shock mosaicism, and, if these were added to the number that show shock lamellae, the total may be as much as 40 percent. By using the average thickness (5 mm) and concentration of impact-derived mineral grains in the impact layer in the Raton Basin and by assuming that the impact layer evenly covered a minimum area of about 12.6×10^6 km^2 in North America (2,000 km radius), I calculated that the total amount shock-metamorphosed mineral grains would be about 1.2 km^3 in this area alone.

The amount of quartz and other clastic mineral grains in the K-T boundary impact layer was determined by a second method. Weighed samples from five different K-T boundary localities in the Raton Basin and one from the Teapot Dome locality in Wyoming were dissolved in dilute hydrofluoric acid (about 5 percent) using the technique described elsewhere in this report. Sample weight (corrected water free) and the weight percent of the residue containing shocked minerals are given in table 8. Inspection of the data shows some

Table 8.--Clastic mineral grain content of the K-T boundary impact layer from Colorado, New Mexico, and Wyoming.
[Sample weight listed corrected for 15 percent water (see table 1); g, gram]

Location	Sample wt. (g)	Wt. clastic grains (g)	Wt. percent clastic grains
Clear Creek North, Colo.	1,700	9.22	0.54
Clear Creek North, Colo.	76	0.02	.03
Clear Creek North, Colo.	850	2.77	.33
Clear Creek North, Colo.	438	.63	.14
Starkville South, Colo.	536	1.02	.19
Berwind Canyon, Colo.	468	2.78	.59
Madrid, Colo.	263	0.38	.14
Madrid East, Colo.	259	1.82	.70
Canadian River, N. Mex.	56	.01	.02
Teapot Dome, Wyo.	701	.32	.05

variability in weight-percent residue in samples only a few meters apart, such as at the Clear Creek North, Colo., locality. Moreover, the amount of clastic mineral grains, which varies from about 0.02 to 0.70 weight percent, is considerably less than the amount determined by point counting thin sections of the impact layer (as much as 2 percent by volume). Four factors may account for the discrepancy between the point count and acid digestion results: (1) some grains may have been decanted with clay during sample preparation, (2) other grains may have been dissolved during acid treatment, (3) non-impact layer material may have been inadvertently included in the sample during collection, and (4) it may not be possible to determine precisely the amount of a trace constituent by the point-count technique. Whatever the cause of the discrepancy, the apparent amount of clastic mineral grains obtained by acid digestion is probably less than the actual amount.

Although the amount of shocked mineral grains in the impact layer in the Raton Basin may be as much as 0.8 percent, the amount of shock-metamorphosed mineral grains at K-T boundary sites outside of North America is considerably less (Bohor and Izett, 1986; Izett, 1987a). At some K-T boundary sites outside of North America, shock-metamorphosed minerals are exceedingly rare. For example, I did not find a single shock-metamorphosed mineral grain in any sample I collected from three New Zealand sites (Woodside Creek, Chancet Rock, and Flaxbourne River). These sites have recently been discussed by Strong and others (1987, fig. 1). Bohor and others (1985) reported that they did not find a single definitely shocked grain in a search of 500 grains recovered from samples from Caravaca, Spain, Woodside Creek, New Zealand, and two sites in Denmark, Nye Kløv and Stevns Klint. Subsequently a few shocked grains were found at each of these sites (Bohor and Izett, 1986).

At European K-T boundary sites shock-metamorphosed mineral grains are much more abundant than at New Zealand K-T boundary sites (Bohor and Izett, 1986; Izett, 1987a). A sample of the 1.0 mm-thick impact layer from the Caravaca, Spain, locality was collected in 1987 by the author because it is representative of European K-T boundary sections and it is exceptionally complete and well exposed. I determined the amount of shocked mineral grains by point counting the acid insoluble residue of the impact layer in immersion oil. Only two clearly shocked quartz grains (two sets of shock lamellae) were counted in a group of 1,000 grains. Thus, the amount of shocked mineral grains at Western North American sites is probably several orders of magnitude greater than at European sites.

SHAPE OF SHOCK-METAMORPHOSED MINERAL GRAINS

The shape of shock-metamorphosed mineral grains was determined by optical and scanning electron microscopic methods. The grains are angular to subangular, and some grains have a tendency to form tabular-shaped grains similar to feldspar. A tabular shape for quartz grains is unusual because of its lack of cleavage. Many shocked quartz grains contain and break along shock-induced planar features, most of which are parallel to $(10\bar{1}3)$, the rhombohedral planes in quartz (Englehardt and Bertsch, 1969). Therefore, many shocked quartz grains are flattened parallel to a plane about 23 degrees from the basal plane (0001).

SIZE OF SHOCK-METAMORPHOSED MINERAL GRAINS

Bohor and others (1984) reported that shock-metamorphosed quartz grains in the K-T boundary claystone at Brownie Butte, Mont., were only 0.1 mm in diameter. The apparent size of the shocked minerals was smaller than grains measured by me (0.5-0.6 mm) in samples from the Raton Basin (Izett and Pillmore, 1985b). This apparent difference in size was of considerable interest because of the possibility that grain-size distribution could be used to narrow the search for the K-T boundary impact site. In addition, the size distribution might reveal dispersal mechanisms of material away from the impact site. Study of samples from Montana provided to me by C.L. Pillmore, R.H. Tschudy, and B.F. Bohor showed that shock-metamorphosed grains were as large as those in the Raton Basin. I found shock-metamorphosed minerals concentrated in a microlaminated claystone (impact layer) that overlies the boundary claystone. Bohor and his colleagues did not find large grains because they only studied the boundary claystone and had scraped off the impact layer that contains the concentration of shock-metamorphosed minerals.

I determined the size of clastic grains in the impact layer, some of which contain shock lamellae, by making a sieve analysis on about 9 g of clastic mineral grains obtained by dissolving the clay mineral fraction from a large sample of the impact layer collected at the Clear Creek North, Colo., locality. This analysis showed that about 30 percent of the grains are less than 0.1 mm, 45 percent are 0.1-0.2 mm, 20 percent are 0.2-0.3 mm, and only about 5 percent are 0.3-0.6 mm.

To obtain a more accurate picture of the grain-size distribution of clastic grains in the impact layer, I measured the size of individual grains, using a petrographic microscope equipped with a calibrated micrometer ocular and the point-count technique. The long axis of 1,500 grains was measured, recorded, and used as the index of grain size. Measurements were made on grains exhibiting multiple sets of shock lamellae as well as grains that did not contain such features. Statistical analysis of the size data for shocked and unshocked grains showed that they have nearly identical means. For example, the mean size of 600 apparently unshocked grains from the Clear Creek North, Colo., locality is 0.17±0.06 mm, whereas the mean size of 400 shock-metamorphosed grains from the same locality is 0.20±0.06 mm.

The size distributions of shock-metamorphosed grains in the impact layer from two sites in the Raton Basin (Clear Creek North and Berwind Canyon), a newly discovered site near Teapot Dome, Wyo. (Wolfe and Izett, 1987), and a site at Brownie Butte, Mont. (Bohor and others, 1984) are shown on figure 33. In addition, the size distribution of 100 shock-metamorphosed quartz grains from the Red Deer Valley locality of Lerbekmo and St. Louis (1985) is shown on figure 34. The mean grain size for the 100 grains was determined to be 0.26±0.06 mm. The mean size of shocked grains from the two sites in the Raton Basin is 0.20±0.06 mm and 0.16±0.06 mm, whereas the mean size of grains from Wyoming and Montana are 0.14±.04 mm and 0.15±0.05 mm. The size distribution shows that grains less than about 50 micrometers were not encountered or measured in grain mounts; such small grains were probably lost during sample preparation. That is not to say that shock-metamorphosed grains as small as 10 micrometers do not occur in the impact layer for they have been seen in thin sections. Although the mean size of shocked quartz grains in the impact layer ranges from 0.14 to 0.26 mm, many grains are 0.2-0.3 mm long. Rare grains, as long as 0.50 to 0.64 mm, are found at all sites in Western North America from New Mexico to Canada (fig. 35). In contrast, the largest shock-metamorphosed mineral grains are only 0.19 mm long at sites outside of North America (fig. 35).

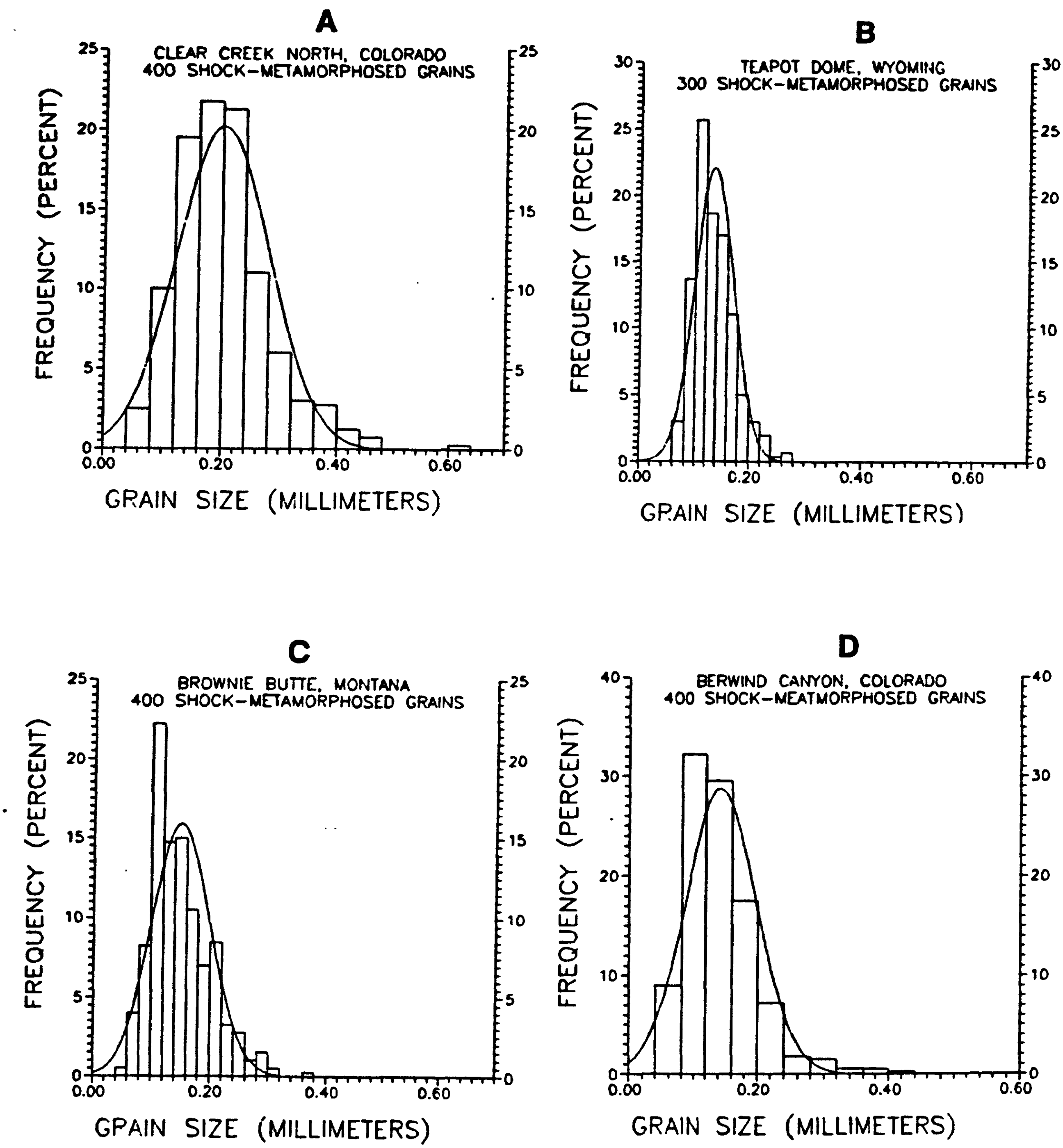


Figure 33. Frequency diagrams showing size of shock-metamorphosed mineral grains in the K-T boundary impact layer. Superposed chi-square distribution curves are also shown. A, 400 shock-metamorphosed grains from the Clear Creek North, Colo., locality. B, 300 shock-metamorphosed grains from the Teapot Dome, Wyo., locality of Wolfe and Izett (1987). C, 400 shock-metamorphosed grains from the Brownie Butte, Mont., locality. D, 400 shock-metamorphosed grains from the Berwind Canyon, Colo., locality.

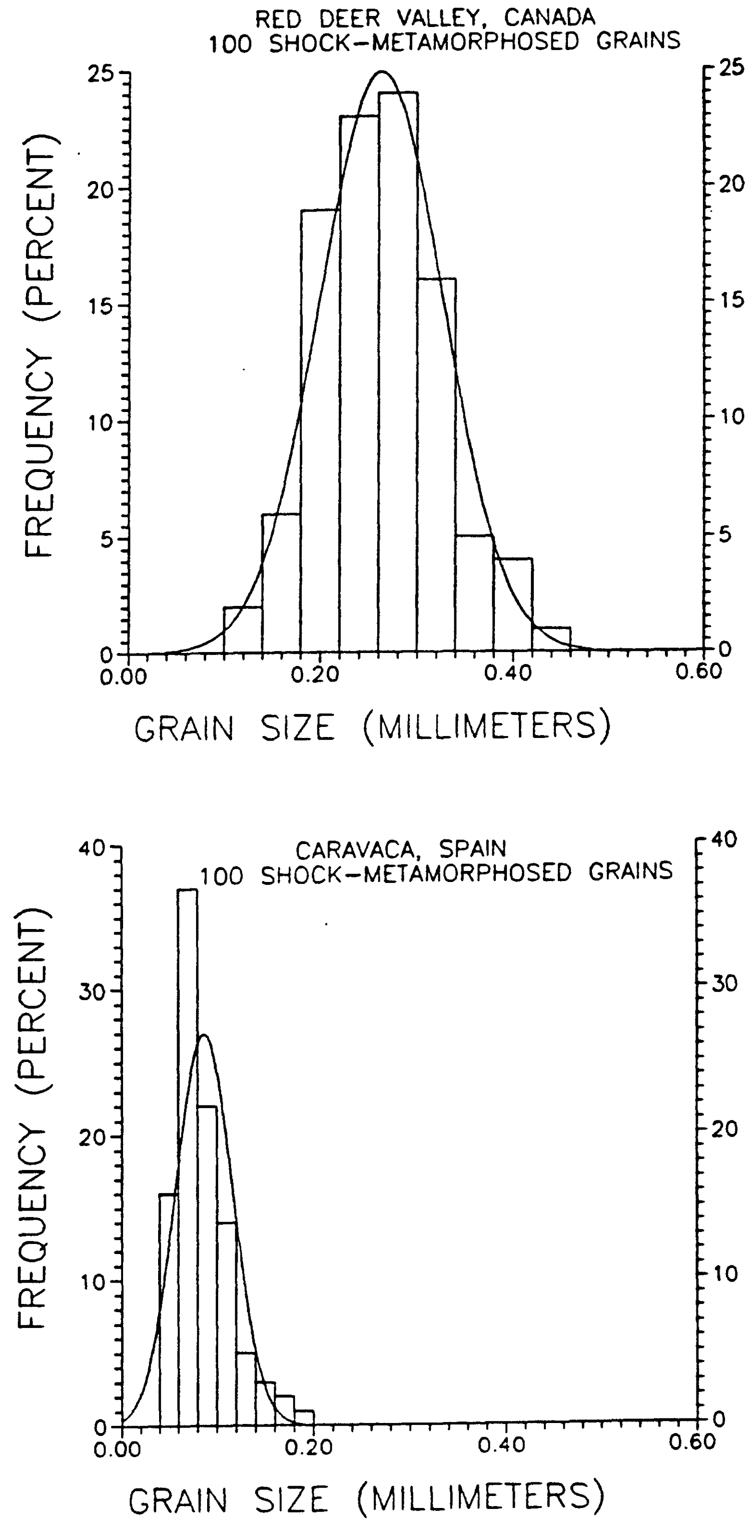


Figure 34. Frequency diagrams showing size of shock-metamorphosed mineral grains in the K-T boundary impact layer at the Red Deer Valley, Canada, and Caravaca, Spain, localities. Superposed chi-square distribution curves are also shown.

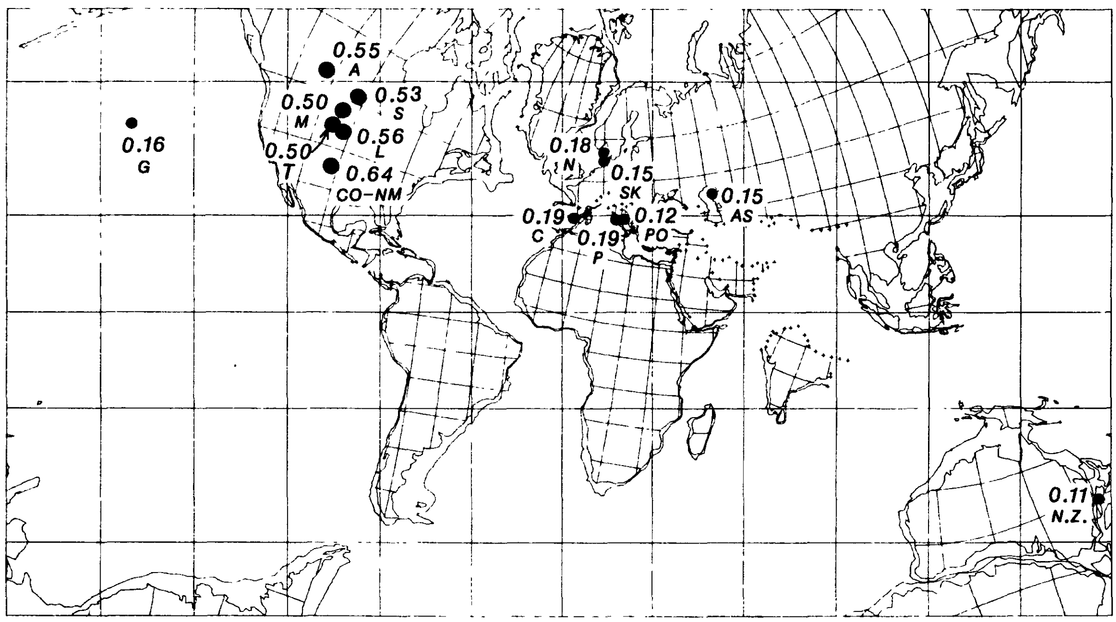


Figure 35.--Paleocontinental map showing the maximum size of shock-metamorphosed quartz grains in K-T boundary rocks. Grain size given in millimeters. Size of locality dots indicates graphically that shock-metamorphosed minerals are several orders of magnitude more abundant at North American boundary sites than elsewhere in the world. Information relative to sites follows: A, Alberta, Canada (Lerbekmo and St. Louis, 1985); S, Saskatchewan, Canada (Nichols and others, 1986); AS, Turkmenia (Badjukov and others, 1986); CO-NM, Raton Basin, Colo., and N. Mex. (fig. 2, this paper); L, Dogie Creek area, Wyo. (Bohor and others, 1987b); T, Teapot Dome, Wyo. (Wolfe and Izett, 1987). Other sites shown by Bohor and Izett (1986) and Izett (1987a) are: M, Brownie Butte, Montana; C, Caravaca, Spain; GPC-3, Giant piston core, North Pacific Ocean; N, Nye Kløv, Denmark; SK, Stevns Klint, Denmark; N.Z., Woodside Creek, New Zealand; P, Pontedazzo, Italy; PO, Petriccio, Italy. Base map (early Paleocene time) from Smith and Briden (1977).

To determine the size distribution of shock-metamorphosed quartz grains at K-T boundary sites outside of North America, a large sample of the impact layer (see fig. 10) from the Caravaca, Spain, locality (Izett, 1987a) was collected in 1987. A size analysis of shocked mineral grains from this particular European locality was made because the 1-mm-thick K-T boundary impact layer readily can be shaved from the enclosing sedimentary rocks. Moreover, a reasonably large number of quartz grains can be recovered from the shavings by dissolving them in a dilute hydrofluoric acid solution. The size of shock-metamorphosed grains in the quartz grain concentrate was determined by the point-count technique as described above. Previous studies aimed at charting the size of shocked quartz grains on a global scale had reported only the size of the largest shocked grains recovered at a particular site (Izett, 1987a). The largest shocked grain (0.19 mm) measured outside of North America was found at the Pontedazzo, Italy, site (fig. 35). The absence of large grains could be attributed to either real differences in the original global size distribution or to the general paucity of large shocked grains. Large grains are even rare at North American sites (fig. 33).

The size distribution of shocked quartz grains in the K-T boundary impact layer at the Caravaca, Spain, K-T boundary locality is shown on figure 34. One hundred grains were measured; the mean grain size is 0.09±0.03 mm, and the range is 0.04-0.19 mm. Thus, the mean grain size of shock-metamorphosed quartz grains from the Caravaca, Spain, locality (0.09 mm) is considerably less than the mean size of shock-metamorphosed quartz grains at North American K-T boundary sites (0.14-0.26 mm). The largest grain measured from the Caravaca, Spain, site (0.19 mm) is about the same size as the mean grain size (0.20 mm) from sites in the Raton Basin (see fig. 33).

The discovery by Thein (1987) of shocked quartz grains at the base of a 12-cm-thick layer of coarse-grained glassy impact material of late Eocene age (Glass, 1987) in DSDP 612 near the Baltimore Canyon off the New Jersey Coast adds information relative to the size-distribution of shocked mineral grains away from an impact point. The source of this layer of impact material is possibly a newly discovered early Tertiary impact crater (Jansa and Pe-Piper, 1987) in the Atlantic Ocean off Nova Scotia about 800 km north of DSDP site 612. I measured the size of 40 shocked quartz grains in the sediments of DSDP 612. The grains range in size from 0.10 to 0.5 mm and average 0.26±0.1 mm (fig. 36). The size distribution of shocked quartz grains in DSDP 612 is similar to the size distribution in K-T boundary sediments in Western North America (figs. 33 and 34). This similarity and the possibility that the source of the shocked quartz grains in DSDP 612 is 800 km away suggests that the K-T boundary impact was a similar distance away from Western North America K-T boundary sites.

The process by which shock-metamorphosed mineral grains, which range from 0.1 to 0.19 mm, were transported globally is unknown. It seems unlikely that such large grains could be transported by upper stratospheric winds for more than several thousand kilometers from the K-T boundary impact site. By analogy, the size of glass shards in volcanic ash beds generated during the largest pyroclastic eruptions decreases systematically downwind from the sources of the ash beds as shown in table 9. The largest glass shards in the Lava Creek B volcanic ash bed are only about 0.45 mm in diameter 2,000 km downwind from the source of the ash in Yellowstone National Park, Wyo. A mechanism for transporting small-size impact material outside of the atmosphere was proposed by Jones and Kodis (1982). They concluded that particles as large as 10 micrometers could be transported globally above the atmosphere within a few hours after the K-T impact.

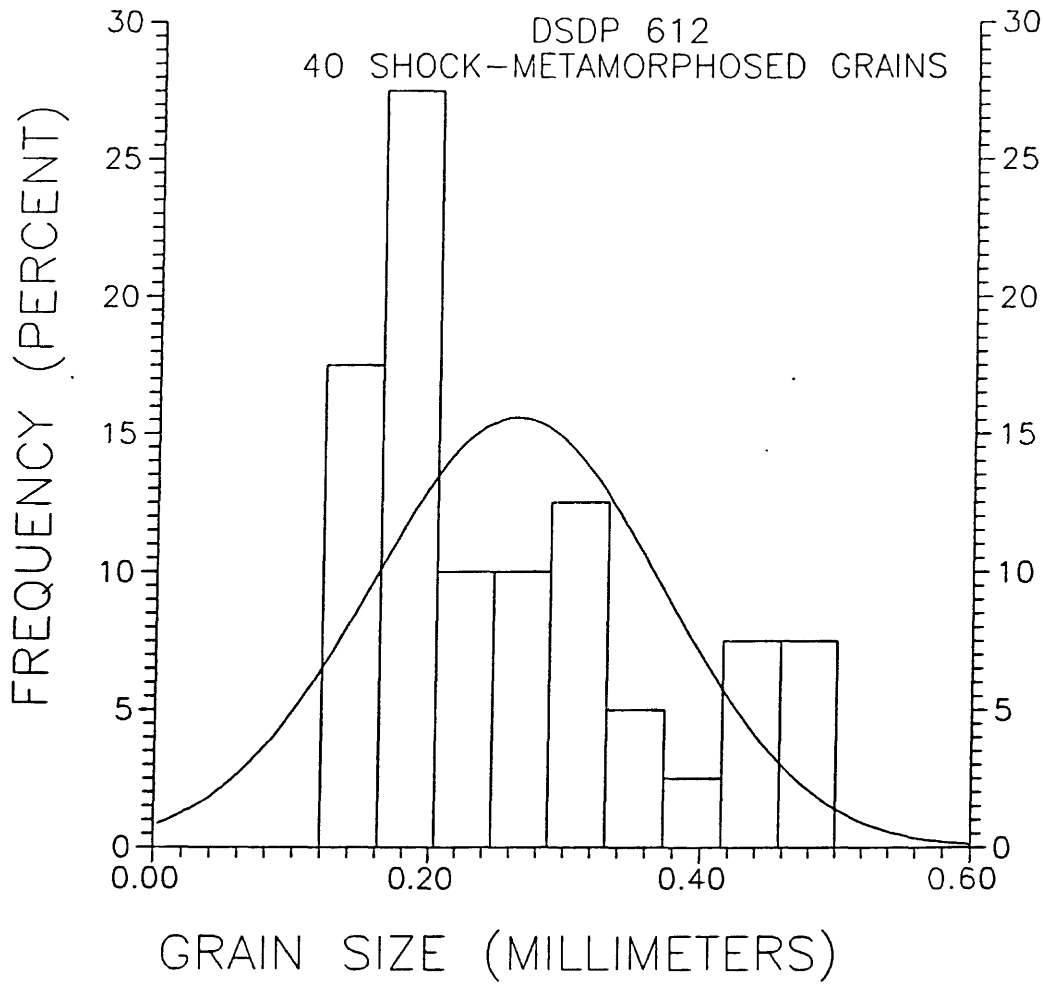


Figure 36.--Frequency diagram showing size of shock-metamorphosed quartz grains from the base of a 12-cm-thick layer of upper Eocene tektite-rich sediment in DSDP site 612 drilled on the continental slope off New Jersey. Superposed chi-square distribution curve is also shown.

Table 9.--Maximum size of glass shards and crystals in the Lava Creek B volcanic ash bed. Size of crystals (in millimeters) in the ash tabulated as a function of distance from its source in Yellowstone National Park (YNP). See Izett and Wilcox (1982) for descriptions of the 0.62 Ma Lava Creek B volcanic ash bed.
[No data (---); Cpx, clinopyroxene; Chev, chevkinite]

Location	Distance from YNP (km)	Cpx	Zircon	Chev	Glass shards
Sublette, Co., Wyo.	170	1.66	0.80	0.82	1.14
Fremont, Co., Wyo.	200	2.45	.82	.71	1.13
Routt, Co., Colo.	550	---	.75	.60	0.95
Garfield, Co., Colo.	650	0.90	.88	.61	1.05
Jefferson, Co., Colo.	700	1.00	.68	.55	.86
Jefferson, Co., Colo.	700	.82	.58	.68	.82
Sheridan Co., Kans.	1,020	.32	.27	.32	.76
Meade, Co., Kans.	1,200	.79	.64	.52	.84
Meade, Co., Kans.	1,200	.65	.54	.50	.71
Meade, Co., Kans.	1,200	.41	.57	.56	.79
Meade, Co., Kans.	1,200	.48	.43	.50	.60
Meade, Co., Kans.	1,200	.57	.55	.36	.77
Meade, Co., Kans.	1,200	.43	.64	.50	.82
Pottawattamie, Co., Ia.	1,200	---	.53	.37	.75
Dickens, Co., Tex.	1,460	.29	.30	.35	.68
Eddy, Co., N. Mex.	1,470	.31	.30	.35	.57
Johnson, Co., Kans.	1,480	.32	.34	.34	.59
Presidio, Co., Tex.	1,700	.29	.32	.25	.45
Bastrop, Co., Tex.	2,000	.15	.16	.14	.45

LOCATION OF THE K-T BOUNDARY IMPACT

The location of the K-T boundary impact crater is controversial and not known with assurance. Some have suggested that the impact occurred in an ocean basin (Emiliani and others, 1981; DePaolo and others, 1983). If this were true, then the Eastern Pacific Ocean or the continental slope of Western North America might be the logical place to look for the crater because of the relatively large amounts and large size of shock-metamorphosed minerals in Western North American K-T boundary rocks. But the rarity and small size of shocked minerals in sediments of a giant piston core (GPC-3) in the North Pacific Ocean (Bohor and Izett, 1986; Izett, 1987a), in relation to those in Western North American boundary sites, argue against the idea that the impact occurred off the west coast of North America. Hartnady (1986) suggested that the K-T boundary impact occurred in the 300-km-diameter Amirante Basin in the Western Indian Ocean. This idea is seemingly incompatible with mineralogic evidence found in K-T boundary rocks in Western North America, including relatively large amounts and large-sized grains of shock-metamorphosed quartz (Izett, 1987a, fig. 10; this report, figs. 33, 34, and 35).

On the basis of the large grain size of shock-metamorphic minerals in the K-T boundary rocks, Izett and Pillmore (1985b) proposed that the impact occurred within a few thousand kilometers of the Raton basin of Colorado and New Mexico. This proposal was made by assuming that the decrease in size of impact material away from the K-T impact point was analogous, <u>in part</u>, to the decrease in size of glass shards and microphenocrysts away from the sites of gigantic pyroclastic volcanic eruptions (Izett, 1987b; Rose and Chesner, 1987). Table 9 shows the size of euhedral, glass-mantled microphenocrysts and glass shards in the Lava Creek B volcanic ash bed (0.62 Ma). This ash bed was generated in Yellowstone National Park, Wyo., during one of the largest volume pyroclastic eruptions known in the geologic record (Christiansen, 1979; Izett, 1981; Izett and Wilcox, 1982). The size of microphenocrysts in the ash was tabulated as a function of distance from its source in Yellowstone National Park, Wyo. The largest glass-mantled microphenocrysts seen in oil-immersion mounts were measured using a calibrated micrometer ocular. Inspection of the data shows that microphenocrysts systematically decrease in size downwind from Yellowstone National Park in Wyoming. Clinopyroxene microphenocrysts are as large as 2.45 mm in Lava Creek B ash near its source, but they are only about 0.3 mm in length 2,000 km downwind near Austin, Texas. Comparison of the data of table 9 and figures 35 and 36 suggests to me that the K-T boundary impact occurred in North America. The major portion of the shock-metamorphosed mineral grains produced by the impact were transported by atmospheric winds away from the impact point to their depositional sites in Western North America. A small amount of small shock-metamorphosed mineral grains were transported globally by an unknown mechanism.

French (1984), Izett and Pillmore (1985b), and Hartung and others (1986) suggested that the K-T boundary impact occurred in northwestern Iowa near the town of Manson at a structurally complex area buried by 30 m of Pleistocene till. Although the Manson structure was originally thought to be a cryptovolcanic structure (Hale, 1955), Short and Bunch (1968, fig. 6) pointed out that the Manson structure was formed by impact processes. Drilling in the region had established that the normal Paleozoic and Mesozoic sedimentary rock sequence is highly disturbed within an elliptically shaped area whose major axis is 40 km and minor axis is 29 km (Hale, 1955, fig. 26). The exact geometry of the Manson impact structure is poorly known because of the Pleistocene till cover, and it may be slightly larger than the limits defined by drilling records.

Three geophysical studies have been undertaken to better define the geometry of the Manson structure. Holtzman (1970) conducted a gravity survey, and he concluded that the structure is defined by a large negative gravity anomaly, but its boundary cannot be precisely located. A seismic refraction study of the Manson structure was made by Smith (1971). He suggested that the structure is circular and has a central uplift (700 m above the basement rocks) composed of crystalline rock surrounded by deformed Cretaceous shale. An aeromagnetic map of the area was prepared by Henderson and Vargo (1965).

Paleozoic sedimentary rocks in northwestern Iowa include Cambrian sandstone, limestone, and shale; Ordovician dolomite and sandstone (St. Peter Sandstone); Devonian and Mississippian dolomite, limestone, and shale; Pennsylvanian black shale; and Permian shale, sandstone, and gypsum. Mesozoic rocks include the Cretaceous Dakota Sandstone and Upper Cretaceous marine shale. The total thickness of these sedimentary rocks is about 760 m.

Igneous and metamorphic crystalline rocks near the center of the structure are uplifted about 760 m above their normal position (Hoppin and Dryden, 1958). They are mainly composed of granitic and gneissic rocks that apparently form the central peak of an impact structure (Dryden, 1955a, 1955b). A geologic sketch map of the Manson, Iowa, impact structure showing the Precambrian subcrop below Pleistocene till is shown on figure 37. This map is a modified version of a map compiled by Sims (1985) from data provided to him by R.R. Anderson of the Iowa Geological Survey.

A thick sequence of Keweenawan age Proterozoic quartzose sandstone and conglomerate associated with the Midcontinent Rift probably occurs in the subsurface in the region of the Manson structure. The maps of Sims (1985) and Anderson (1982) of the Midcontinent region show that quartzose clastic rocks of Proterozoic age probably extend along the Midcontinent Rift from the Lake Superior region through Iowa into Kansas. In addition, thin equivalents of the Baraboo, Barron, Flambeau, McCaslin, Waterloo, Sioux, and Bessemer Quartzites that outcrop in Minnesota, Wisconsin, and South Dakota (Morey and Ojakangas, 1982; Ojakangas and Morey, 1982; Sims, 1985) might extend into northwestern Iowa.

A borehole (Manson 2A) was drilled to a depth of 146 m in 1953 in the crystalline rock of the Manson structure, and a 3.8-cm-diameter core was obtained by the U.S. Geological Survey and Iowa Geological Survey. The borehole is north of the town of Manson in the SW1/4SW1/4SW1/4 sec. 29, T. 90 N., R. 31 W., Pocahontas County. Records and core samples provided to me by R.R. Anderson (written commun., 1986) indicate that the depth of the borehole was 146 m. The drilling records show that about 30 m of till overlies about 58 m of highly brecciated granitic and gneissic rocks including rare, altered, mafic-rock fragments (Dryden, 1955b). About 58 m of non-brecciated or less brecciated granitic and gneissic rock lies below the highly brecciated interval. A second borehole (Manson 1) was drilled in the disturbed shales that surround the granitic rock core of the structure (Hoppin and Dryden, 1958, p. 695).

The granitic rock in the Manson 2A core was described by Dryden (1955a, 1955b), and the results of my study of the core are, in general, in agreement with his. The granitic rock is a pink to white medium-grained rock that has a allotriomorphic texture. X-ray diffraction analysis of the granitic rock from the core shows that it is composed of quartz, potassium feldspar, and plagioclase. A small amount of kaolinite is also present. Modal analysis (1,000 points counted) of a thin section of the granite from the 144-m depth showed it chiefly consists of the following (volume percent): quartz, 34; potassium feldspar, 44; plagioclase, 16; biotite, 4; and myrmekite, 2.

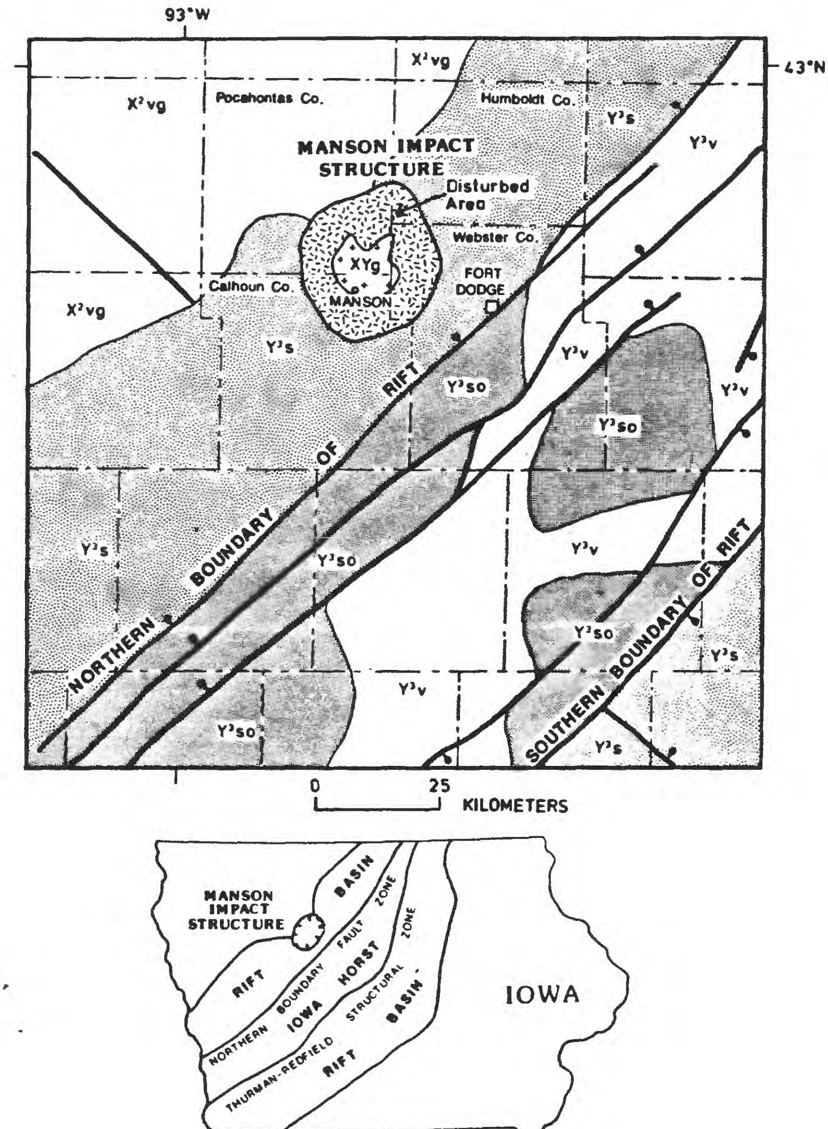


Figure 37.--Geologic sketch maps of the Manson, Iowa, impact structure. Small map shows the location of the Manson impact structure in relation to the Midcontinent Rift (modified from Anderson, 1982). Large map shows Precambrian basement rocks in the Manson, Iowa, region (modified from Sims, 1985). Lower Proterozoic rocks: X^2vg, granitoid and volcanic rocks undivided. Lower and middle Proterozoic rocks: XYg, granite and gneiss. Middle Proterozoic rocks: Y^3v, basalt, rhyolite, and volcanic and sedimentary rocks; Y^3so, red lithic sandstone, siltstone, and mudstone; Y^3s, red to buff quartzose

Accessory apatite, zircon, and magnetite also occur in the rock. Thin sections of the rock show that it contains healed fractures and has granulated grain boundaries.

The quartz is anhedral and typically occurs in grains about 0.3 mm in diameter. It has a pervasive yellowish-brown color in thin sections. The potassium feldspar is triclinic as shown by X-ray diffraction analysis and the local presence of albite-pericline twins, typical of microcline. The microcline occurs as anhedral grains as large as 4.0 mm. Its optic angle, which was estimated by the curvature of optic axis isogyres, is large ($2V_x$=80). Dryden (1955b) mentioned that the optic angle of some potassium feldspar in the granitic rock is locally small. As previously noted, the optic angle of some shock-metamorphosed triclinic potassium feldspar grains in the K-T boundary impact layer was lowered ($2V_x$=30) by the shock metamorphism. The refractive indices indicate that the composition of the feldspar is rich in potassium and very low in sodium (N_x=1.516, N_y=1.520, and N_z=1.525). The refractive indices were measured in the immersion cell of a spindle stage using the focal masking technique (Wilcox, 1959, 1983). The plagioclase is calcic oligoclase as determined by measuring extinction angles [15^o; X makes an angle with (010), maximum] in albite twin lamellae using a universal stage. The refractive indices of the plagioclase are N_x=1.534, N_y=1.539, and N_z=1.544. The optic angle as measured on the spindle stage is large, near 90^o. The biotite is generally altered to chlorite, and it occurs in smeared out masses.

The quartz and potassium feldspar (fig. 38) in the granitic rock pervasively contain microstructures consisting of multiple intersecting sets of shock lamellae and mosaic extinction typical of shock-metamorphosed minerals (Short and Bunch, 1968, fig. 6). The quartz contains opaque inclusions similar to those in quartz in the K-T boundary impact layer (compare figs. 30 and 38A). The oligoclase, less commonly than the potassium feldspar, also contains shock lamellae. The potassium feldspar consistently shows mosaic extinction and poorly developed shock lamellae (fig. 38A). Albite twin lamellae in the plagioclase are locally bent and dislocated. The oligoclase is partially altered to kaolinite and contains flecks of sericite. No thetomorphic glass was seen in either the shocked quartz or feldspars during study of five thin sections of the granitic rock of the Manson core.

The gneiss in the Manson 2A core was described by Dryden (1955a; 1955b), and the results of my study of the core are in general agreement with his descriptions. I found the gneiss to consist mainly of quartz, plagioclase, and biotite. Locally it contains hornblende, pink garnet, sphene, epidote, magnetite, and pyrite. The quartz occurs in interlocking grains about 0.5 mm in diameter. Multiple sets of shock lamellae occur throughout the quartz. The plagioclase is generally altered to clay and its composition is difficult to ascertain; however, unaltered plagioclase was seen in two thin sections. Its composition is calcic oligoclase as indicated by measurement of extinction angles [15^o; X makes an angle with (010), maximum] in albite twin lamellae. The biotite is dark yellowish brown and relatively unaltered. An unusual mineralogic feature of the gneiss from the 138-m depth is the occurrence of magnetite crystals as large as 4.0 mm in quartz-feldspar augen.

The stratigraphic age of the Manson structure is bracketed between the Late Cretaceous and Pleistocene. The Late Cretaceous age is based on the presence of the fossil clam Inoceramus in the deformed shales of the Manson 1 borehole (Hoppin and Dryden, 1958).

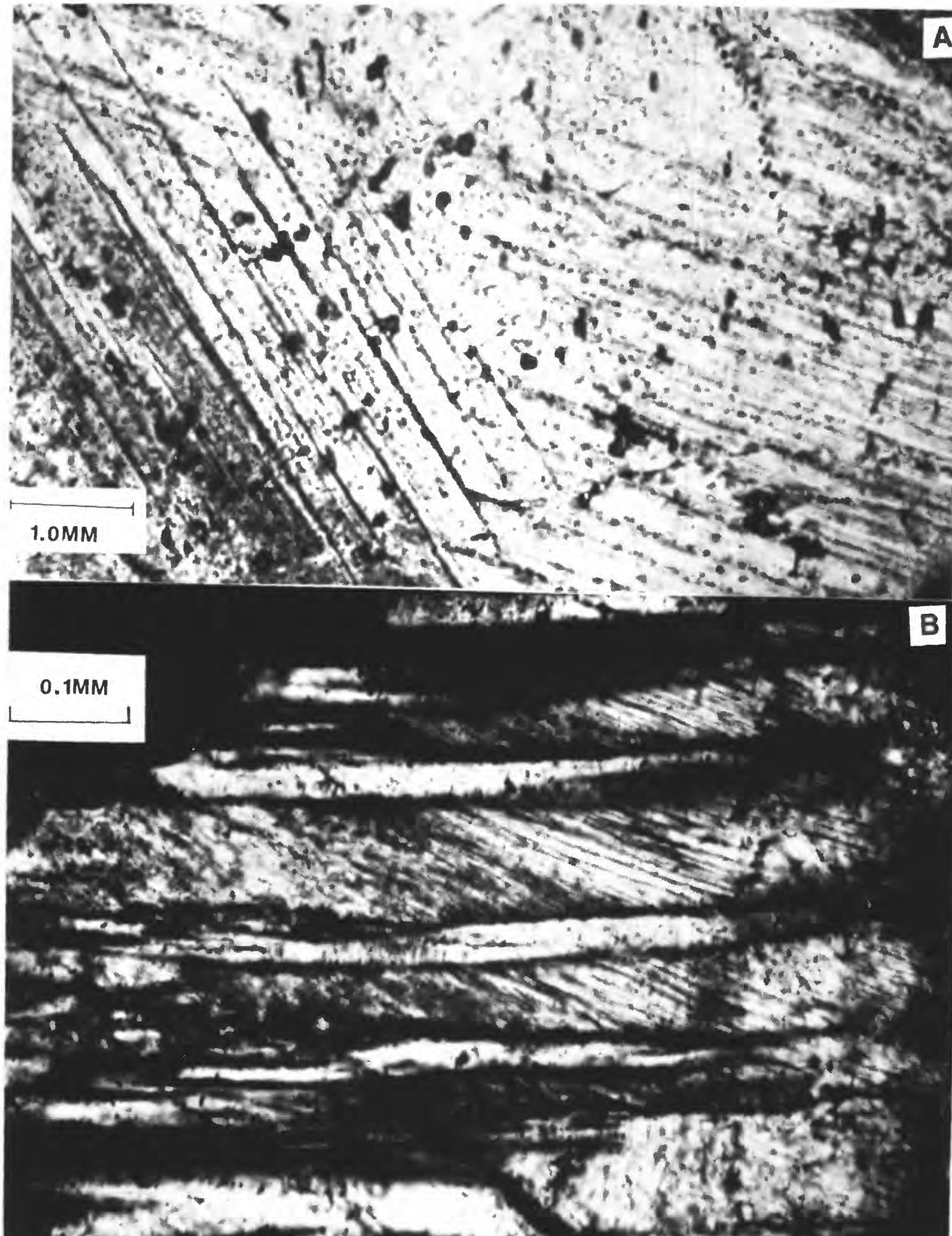


Figure 38. Photomicrograph showing shock lamellae in quartz and feldspar in granitic rock from the Manson 2A core hole. A, Two sets of shock lamellae in quartz. Note opaque inclusions similar to those in shocked quartz in the K-T boundary impact layer (fig. 30). Most quartz grains in thin sections of the granite show multiple sets of shock lamellae regardless of their orientation. B, Shock lamellae in albite twin lamellae in oligoclase. Photographed with plane-polarized light.

Z.E. Peterman of the U.S. Geological Survey determined the age of the granite by the rubidium-strontium, whole-rock isochron method. He analyzed six samples of granite from the Manson 2A core hole, and provided me with a plot of the data (fig. 39). The data are aligned along a 886 Ma isochron that, according to Peterman, does not reflect the emplacement-age of the granite. Rubidium and strontium were mobilized during a metamorphic event, and the isochron was rotated to its present slope of 886 Ma (written commun., 1987). Several model ages for the granite were calculated by Peterman, using the method of Peterman and others (1985), and they average about 1,410 Ma. The average age of 1,410 Ma probably reflects the original emplacement age of the granite and is similar to the age of a granite (1,433±4 Ma) from a drill core in northwest Iowa dated by Van Schmus and others (1987).

Two biotite rubidium-strontium ages of 1,070 and 1,130 Ma (old decay constants) were reported for biotite gneiss from the Manson core by Lidiak and others (1966). More recently, the age of the metamorphic rock from the Manson core was determined by Z.E. Peterman. He measured the ratios of rubidium and strontium in four samples of shock-metamorphosed gneiss from the Manson 2A core hole (fig. 39). Three of the four data points for the gneiss are aligned along a 1,620 Ma isochron. The gneiss, according to Peterman (written commun., 1987), is probably Early Proterozoic in age, near 1,800 Ma, and was overprinted during a pervasive 1,620 Ma thermal event throughout the region.

The fission-track age of a fragment of gneiss from the Manson 2A core was determined by C.W. Naeser, of the U.S. Geological Survey. I pulverized a piece of the rock from the 138-m depth and obtained an apatite mineral concentrate by heavy-liquid separation techniques. My purpose was to determine the age of the apatite by the fission-track method and, possibly, the shock-metamorphic age of the Manson structure. As expected, the apatite was highly fractured, but no shock lamellae were seen, although the host rock is thoroughly shock-metamorphosed. The fission-track age of the apatite as determined by C.W. Naeser is 60.5±18 Ma (±2 sigma level) (table 10).

Table 10.--Fission-track age of apatite crystals from gneiss from the Manson 2A core hole, Manson, Iowa.
[Analyst, C.W. Naeser; precision of age determination is ± 2 sigma; P_s, spontaneous track density; P_i, induced track density]

Lab. No.	Number of grains	P_s X10^6 t/cm^2	Fossil tracks counted	P_i X10^6 t/cm^2	Induced tracks counted	Neutron dose X10^{15} n/cm^2	Age (Ma)
5102	6	0.128	58	0.824	187	6.55	60.5±18

Decay constant for spontaneous decay of ^{238}U (λ_f)=7.03X$10^{-17}yr^{-1}$

Unfortunately, the analytical uncertainty of the age is large owing to the highly fractured state of the apatite. As noted previously, the apatite occurs in a Proterozoic age gneiss, and the most reasonable interpretation for the 60 Ma fission-track age of the apatite is that it experienced a heating event probably associated with the shock metamorphism of the gneiss during the Late Cretaceous or early Tertiary.

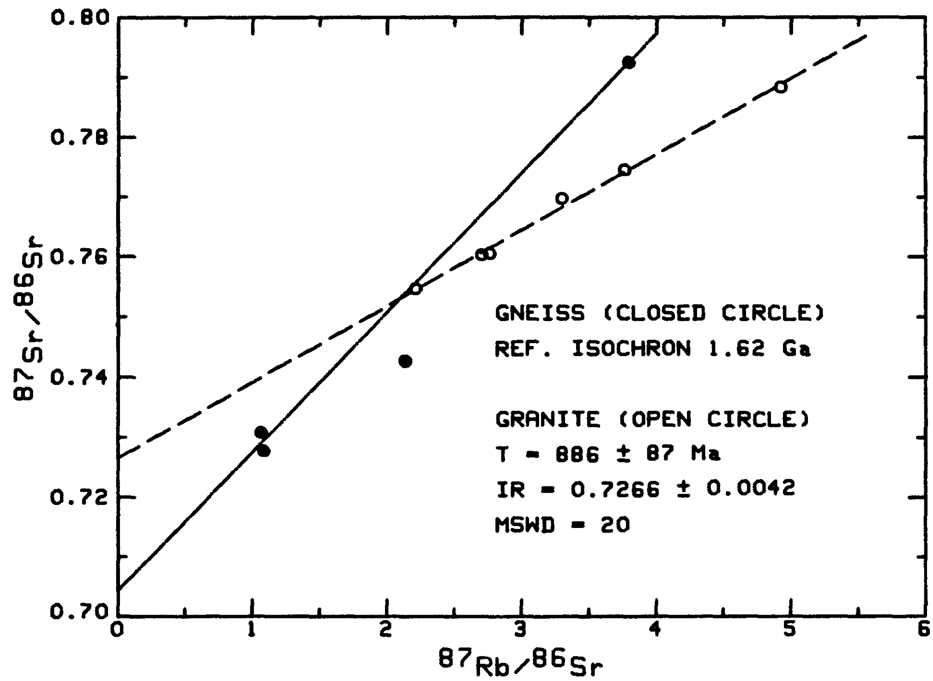


Figure 39.--Rubidium-strontium isochron diagram for granitic and gneissic rocks from the Manson 2A core hole. Data from Z.E. Peterman (U.S. Geological Survey, written commun., 1987).

To establish a more precise shock-metamorphic age for the crystalline rocks of the Manson structure, I sent M.J. Kunk and J.F. Sutter, of the U.S. Geological Survey, a sample from the 70-m depth of granitic rock from the Manson 2A corehole. Using heavy-liquid separation techniques, they separated potassium feldspar from the sample and determined its age using the $^{40}Ar/^{39}Ar$ step-heating method. The age spectrum of the feldspar was reported by Hartung and others (1986). They suggested that the Proterozoic granitic rock containing the potassium feldspar underwent a heating event "* * * less than, but not much less than, 70 Ma ago." More recently, Kunk and others (1987) made a new $^{40}Ar/^{39}Ar$ age determination of potassium feldspar from granitic rock at the 131-m depth in the Manson 2A core hole. They stated that the feldspar underwent "* * * severe argon loss about 66 million years ago." About 32 percent of the gas was released during heating steps 4 to 7, and the calculated ages of these steps are concordant at 66 Ma. According to M.J. Kunk (written commun., 1987), the ages for steps 1 to 3 are slightly older because the feldspar possibly contains excess argon.

In summary, several lines of evidence are compatible with the idea that the Manson, Iowa, impact structure is the place where the K-T boundary asteroid struck the Earth. The evidence includes (1) the proximity of the Manson, Iowa, impact structure and Western Interior North American K-T boundary sections that record a profound pollen break at the exact K-T boundary [apparently this pollen break was limited to the Western Interior of North America (Tschudy and Tschudy, 1986)]; (2) the compatibility of the composition of known and inferred subsurface rocks in the Manson area (including Proterozoic granite, gneiss, quartzite, and metaquartzite overlain by Phanerozoic sedimentary rock) and of shock-metamorphosed minerals in K-T boundary rocks; (3) the lack of thetomorphic glass in shock-metamorphosed quartz and feldspar in the impact layer and in granitic and gneissic rocks of the Manson 2A core; (4) the proximity of the Manson impact structure to Western North American K-T boundary sites that contain the most abundant and largest shock-metamorphosed mineral grains in the K-T boundary impact layer; (5) the large size of the impact structure (a minimum of about 35 km) (fig. 37); (6) the concordance of the shock-reset isotopic age of 66 Ma of potassium feldspar (Kunk and others, 1987) in granitic rocks of the central uplift of the Manson structure and the isotopic age of the K-T boundary (66 Ma).

K-T BOUNDARY IMPACT CRATER SIZE

Although considerable information, summarized in the preceding section of this report, suggests that the impact structure in the Manson, Iowa, area may be the location of the K-T boundary impact, its diameter (about 35 km) is smaller than the crater size (200 km) estimated for the K-T event by R.A.F. Grieve of the Canadian Geological Survey (in Alvarez and others, 1980, p. 1105). He estimated the size of the K-T crater on the basis of the mass and density of the K-T asteroid (10±4 km) calculated by Alvarez and others (1980). However, the exact relation between the diameter of an asteroid and its impact crater is imperfectly known (Roddy, 1987). So-called "scaling laws" have been developed that attempt to relate asteroid and crater diameters, but these scaling laws are open to serious limitations, as pointed out by Schmidt and Holsapple (1982). Among the variables that control the diameter of a crater formed by an asteroid striking the Earth are its size, velocity, shape, density, and entry angle. In addition, the physical properties of the target material are important. It is obvious that most of these variables are either unknown or imperfectly known for the K-T asteroid, and, thus, estimates that attempt to relate asteroid and impact crater diameters must be used with caution and restraint.

The size and composition of the K-T asteroid are two of the important parameters that must be established if the diameter of its associated impact crater is to be calculated. Alvarez and others (1980) used four different methods to calculate the size of the K-T asteroid. The first method was based on the amount of iridium in K-T boundary interval rocks in Italy and Denmark. They (1980, p. 1105) assumed that the asteroid was chondritic and contained a mass fractional abundance of iridium equal to $0.5X10^{-6}$ (see Ganapathy, 1980, table 2). They (1980, p. 1105) used the measured surface density of iridium (8 ng/cm^2) in the K-T boundary interval at Gubbio, Italy, and the surface area of the Earth ($5.1X10^{18}$ cm^2) and arrived at a mass of $7.4X10^{16}$ g for the asteroid. They divided the calculated mass of the asteroid by the so-called Krakatoa fraction (0.22), resulting in a mass of $3.4X10^{17}$ g for the asteroid. They calculated its diameter to be 6.6 km assuming that the asteroid had a density of 2.2 g/cm^3. A diameter of 14 km was calculated using the measured surface density of iridium (80 ng/cm^2) from the Stevns Klint, Denmark, locality.

Thus, the size of the asteroid was markedly increased by their use of the Krakatoa fraction. Alvarez and others (1980) drew attention to historical accounts of the 1883 eruption of the Krakatoa volcano. About 78 percent of the particulate material lofted into the atmosphere stayed in the source area; the remaining 22 percent was injected into the stratosphere and allegedly stayed aloft for 2 to 2.5 years. Supposedly this material was transported globally and caused remarkable sunsets for several years following the eruption. The Alvarez team used the Krakatoa fraction, for a lack of other information, as a method for partitioning proximal and distal impact ejecta. Toon and others (1982, p. 188) noted that particulate silicate dust from large volcanic eruptions settles out within a few months following an eruption, and therefore, the use of the Krakatoa analogy is inappropriate.

If the Krakatoa fraction is not used, the diameter of the K-T asteroid, using the iridium data of Alvarez and others (1980) for Gubbio, Italy, is 4.0 km. Using the iridium data of Alvarez and others (1980) for the boundary claystone (fish clay) at Stevns Klint, Denmark, the diameter of the K-T asteroid would be 8.8 km. If the average iridium surface density of 50 ng/cm^2 for the Raton Basin is used and the mass fractional abundance of iridium in the K-T asteroid were equal to $0.5X10^{-6}$, the diameter of the K-T asteroid would be 7.5 km. If the mass fractional abundance of iridium in the K-T asteroid were 1.0×10^{-6} (upper limit given by Palme, 1982, fig. 1), the diameter of the asteroid would be 5.8 km.

It is important that the above estimates of asteroid size are based on the premise that it was of chondritic composition and contained about about 500 ppb iridium. However, Ganapathy (1980, p. 922) stated that "* * * the abundance pattern of noble metals in the boundary clay does not exactly match that in Cl chondrites." Palme (1982, p. 230) observed that "The pattern of siderophiles [in the boundary claystone] does not match the siderophile element pattern of unfractionated meteorites (chondrites)." On the basis of ratios of osmium isotopes, Luck and Turekian (1983) concluded that the asteroid was composed of nickel-iron. Kyte and others (1985) showed that the abundance ratios of platinum-group metals for individual samples of boundary interval rocks at Caravaca, Spain, are not strictly chondritic. They noted, however, that, if the samples were treated as a whole they approach chondritic abundance ratios.

The second method used by Alvarez and others (1980) to estimate the asteroid size was based on data for earth-crossing asteroids and the impact-cratering record of the Earth. They predicted that 10-km-sized asteroids should intersect the Earth, on average, about every 100 million years. This information seems to be compatible with the impact record of the Earth and the frequency of major mass extinction events.

The third method that Alvarez and others (1980) used to estimate the size of the K-T asteroid was based on the assumption that the 1-cm-thick boundary claystone at Gubbio, Italy, and Stevns Klint, Denmark, is air-fall impact material and that it occurs globally. They assumed that (1) the surface density of the boundary claystone is 2.5 g/cm^2; (2) the asteroid had a density of 2.2 g/cm^3, and the fraction of the material in the claystone of asteroidal origin is 0.5; (3) the ratio of crustal mass to asteroid mass is about 60; and (4) the Krakatoa fraction is 0.22. Using the surface area of the Earth (5.1×10^{18} cm^2), they calculated that the asteroid was 7.5 km in diameter. However, their assumption that the K-T boundary claystone at Gubbio and Stevns Klint is air-fall impact material may not be justified (Rampino, 1982), and accordingly their estimate of asteroid size may be incorrect. For example, Izett (1987a) suggested that the fallout layer at Caravaca, Spain, and Stevns Klint, Denmark, is only about 1 mm thick. Moreover, the clay minerals that compose the boundary claystone may not be impact derived. Instead, they may be normal marine clay as suggested by Rampino and Reynolds (1983).

The fourth method that Alvarez and others (1980) used to estimate the size of the K-T asteroid was according to them "* * * not able to set close limits on the mass of the incoming asteroid." This method was based on a requirement of their impact-extinction theory that the volume of impact-generated dust (and opacity of the sky) be considerably greater than that produced during the Krakatoa eruption of 1883.

Several estimates of the transient K-T crater size have been made using the size (10 km) of the asteroid as calculated by Alvarez and others (1980) as follows: 60-100 km by O'Keefe and Ahrens (1982, p. 104), 50-70 km by Melosh (1982, p. 125), and 66 km by Schmidt and Holsapple (1982, p. 93). On the basis of computer simulations, Roddy and others (1987) concluded that a 10-km-sized spherical asteroid traveling at 20 km/s would form a transient crater about 80 km in diameter in continental crust and about 105 km in diameter in oceanic crust. They also indicated that at the oceanic site the upper part of the mantle may have been melted, but none of it was ejected from the crater. These transient craters would be enlarged by secondary processes, but according to Schmidt and Holsapple, 1982, p. 99) the "* * * degree of enlargement remains in debate." Roddy and others (1987, p. 533) calculated that "* * * no mantle rock was melted or ejected at the continental site." If the diameter of the K-T asteroid were only about 5 km and its velocity about 25 km/s, the transient crater would be about 40 km (Schmidt and Holsapple, 1982, fig. 6).

It is important to keep in mind that one of the most important parameters controlling the relation between asteroid and crater diameters is the angle of incidence of the asteroid. The above cited estimates of the K-T boundary crater size were based on vertical incidence. Schultz and Gault (1982, p. 173) pointed out that about 50 percent of all asteroids have angles of incidence of about 45° from the horizontal. Roddy and others (1987, p. 527) stated that "* * * vertical impact angles require slightly lower kinetic energies than do lower impact angles to form the same size crater."

The ratio of excavated crustal material relative to asteroidal material was estimated at 60 by Alvarez and others (1980, p. 1106), 10-100 by O'Keefe and Ahrens (1982, p. 118), 12 by Schmidt and Holsapple (1982), and 135 by Roddy and others (1987). Melosh (1982, p. 126) concluded that the mass of material injected into the atmosphere was equal to the mass of the asteroid. Schultz and Gault (1982, p. 173) stated that "* * * only a small amount of the available fine-size ejecta fraction produced by such an impact [K-T] may be distributed globally." Clearly, there is no consensus on the amount of target rock lofted into the atmosphere as a result of the K-T impact.

It could be argued that the Manson, Iowa, impact structure is not the K-T boundary impact site because it is too small. Many impact structures are as large or larger than the Manson structure (Grieve, 1987), but, unlike the Manson, the other large impact structures are not apparently associated with mass extinction events. Inherent in this reasoning are the assumptions that the mass of the K-T boundary asteroid is known and that it struck with near vertical incidence. Neither of these assumptions may be warranted as pointed out elsewhere in this report. If the angle of incidence of the asteroid were very low, a smaller crater would be formed as compared to a vertical strike. The atmospheric effects, including the generation of large volumes of nitrogen oxides (Prinn and others, 1987), of a low angle strike probably could be enormous. The killing mechanism for the K-T extinction event may not have been the lofting of large volumes of dust-sized target materials into the atmosphere and the resulting global darkness as predicted by Alvarez and others (1980). Instead the killing mechanism may have been the generation of large volumes of nitric acid produced in the atmosphere from nitrogen oxides. The resulting acid rain would have a disastrous effect on plant life, especially in the Western Interior and the biota of the photozone of the oceans. Recently, O'Keefe and Ahrens (1988) suggested that the killing mechanism may have been carbon dioxide produced by the K-T asteroid impacting carbonate rocks in a shallow marine environment. The thick section of Paleozoic carbonate rocks in the Manson, Iowa, area is more appealing for their scenario because the impact probably occurred in a continental rather than an oceanic setting as discussed previously in this report.

K-AR AGE OF THE K-T BOUNDARY

The 66 Ma $^{40}Ar/^{39}Ar$ age of shock-metamorphosed feldspar from the Manson, Iowa, impact structure (Kunk and others, 1987) is concordant with $^{40}Ar/^{39}Ar$ ages of sanidine obtained from bentonite beds at the K-T boundary (Obradovich, 1984; Obradovich and Sutter, 1984). These age determinations were made in the same laboratory using the same hornblende standard (MMhb-1) as a flux monitor.

I prefer to use 66 Ma for the numerical isotopic age of the K-T boundary, rather than the commonly cited 65 Ma. The 65 Ma age of the boundary is based (in part) on a K-Ar age determination on plagioclase (64.8 Ma) obtained from rounded dacite cobbles 11 m above the K-T boundary in the Denver Formation at South Table Mountain east of Golden, Colorado (Evernden and others, 1964, p. 187). The boundary lies above dinosaur remains and below a mammalian fauna (Puercan) of Paleocene age (Brown, 1943). In 1970, J.D. Obradovich and I collected samples from a layer of monolithologic dacite cobbles on South Table Mountain that occurs 21 m above the Puercan mammal locality described by Evernden and others (1964). The layer of cobbles is presumably 10 m stratigraphically higher than the collection made by Evernden and others (1964). A K-Ar age determination (Obradovich and Cobban, 1975) made on biotite from the dacite cobbles yielded an age of 64.3 Ma, concordant with the age of plagioclase-bearing (no biotite?) dacite cobbles dated by Evernden and others (1964). These ages recalculated using decay constants recommended by Steiger and Jäger (1977) are about 2.5 percent older than the previously calculated ages, that is about 65.8±1.4 Ma (Obradovich, 1984). Because these age determinations (about 65 Ma) were made on rounded dacite cobbles, they represent a maximum age for the K-T boundary, as pointed out by Evernden and others (1964). The 65 Ma may also be the age of the K-T boundary if the dacite was transported and deposited shortly after its eruption.

Shafiqullah and others (1964) K-Ar dated sanidine and biotite from bentonite beds just above the K-T boundary in Canada and Montana. The age of the K-T boundary based only on the sanidine ages and decay constants then in use was 63.4±1.0 Ma. R.S.J. Lambert (in Harland and Francis, 1971, p. 79) reevaluated the geochronologic data and placed their age at 64.0±1.0 Ma. The 64.0 Ma age for the boundary recalculated using decay constants recommended by Steiger and Jäger (1977) is about 65.6 Ma. Lerbekmo and others (1979) dated 11 sanidine concentrates from four bentonite layers in a 18.5-m-thick interval that overlies the Nevis coal bed in the Red Deer Valley area of Alberta, Canada. The Nevis coal bed directly overlies the K-T boundary claystone and impact layer. Shock-metamorphosed quartz grains have been found in the impact layer (Izett, 1987b) and an iridium abundance anomaly was detected in the claystone and overlying Nevis coal bed (Lerbekmo and St. Louis, 1985). They (1979) reported a mean age of 63.1±0.5 Ma (new decay constants) for the four bentonite layers. Baadsgaard and Lerbekmo (1983) dated mineral concentrates from a bentonite that occurs just above the K-T boundary in eastern Montana. They dated zircon by the U-Pb method and biotite and sanidine by the Rb-Sr method. The zircon yielded a 63.9 (+0.6-0.8) Ma age, and the biotite and sanidine gave an isochron age of 63.7±0.3 Ma.

Obradovich (1984) and Obradovich and Sutter (1984) K-Ar dated splits of three sanidine concentrates originally K-Ar dated by Lerbekmo and others (1979) at 63.1±0.5 Ma. Two of the sanidine concentrates were from a bentonite that lies 17 m above the palynological boundary and about 19 m above the K-T boundary claystone in the Red Deer Valley area of Canada, and the third sanidine concentrate was from a bentonite that is about 1 m above the K-T boundary in eastern Montana. Obradovich and Sutter used the $^{40}Ar/^{39}Ar$ step-heating method, instead of the conventional potassium-argon method used in previous studies (Lerbekmo and others, 1979), because it provides considerably more precision. Obradovich (1984) reported that the bentonite closest to the K-T boundary (1 m) is 66.0 ±0.54 Ma. The cause for the difference in K-Ar ages for the K-T boundary as measured in Reston, Virginia by Obradovich and Sutter (66 Ma) and those measured in Alberta, Canada, by Baadsgaard has yet to be determined but was discussed by Obradovich and Sutter (1984).

CONCLUSIONS

My study of the K-T boundary interval rocks in the Raton Basin of Colorado and New Mexico and elsewhere in the world has resulted in the following conclusions:

1. Two kaolinitic claystone microstratigraphic units, the K-T boundary claystone and overlying K-T boundary impact layer, mark the K-T boundary and form a couplet at 20 localities in the Raton Basin. Far to the north, these same units have been identified in Wyoming, Montana, and Canada.
2. The lower unit, the boundary claystone, is a 1- to 2-cm-thick kaolinitic tonstein that is found only at North American, continental K-T boundary localities and not at boundary localities elsewhere in the world. At most localities, the unit lacks clastic quartz and feldspar grains. It typically contains macerated plant material and paleoroots. The texture of the claystone is microspherulitic and probably did not form from an impact or volcanic glass precursor. Its trace-element chemical composition is different from coarse-grained kaolinitic tonstein layers, but similar to average North American shale, except for its above-background content of iridium. The iridium concentration is as much as 66 times less than that in the overlying impact layer and carbonaceous-rich rocks. The claystone contains only a few ppm of nickel and cobalt. Although the origin of the unit is uncertain and controversial, petrographic and chemical evidence suggests it is not altered impact material. The most important fact that bears on the origin of the K-T boundary claystone is that shock-metamorphosed minerals are restricted essentially to the K-T boundary impact layer and are not found in the underlying K-T boundary claystone. It is difficult to understand how the boundary claystone could originate from impact material that would be entirely free of clastic mineral grains, in particular, shock-metamorphosed minerals.
3. Solid kaolinite and solid and hollow goyazite spherules are found in the K-T boundary claystone and not in the overlying K-T boundary impact layer. The combined weight of stratigraphic relations and thin section study of the spherules indicates that they are not of impact origin.
4. The upper unit, the impact layer, is about 5 mm thick and consists mainly of microlaminated claystone that contains laminations of vitrinite and pellets of kaolinite. The contact between the boundary claystone and impact layer is generally sharp and records a significant change in depositional regime. The SEM texture of the impact layer claystone is similar to smectite clay different from the pervasive microspherulitic texture of the boundary claystone. The chemical composition of the impact layer is similar to the underlying boundary claystone except that the former contains significantly less lithium and slightly more iron, potassium, barium, chromium, copper, vanadium, and zinc than the latter. The claystone contains only a few ppm of nickel and cobalt. The iridium content is generally, but not always, the highest in boundary interval rocks at a particular site. Presumably the iridium is mobile during diagenesis and moved into adjacent carbonaceous-rich rocks. Shock-metamorphosed minerals and an unknown phase that carried iridium are the only impact-related components in the claystone. The impact-related components probably were transported atmospherically away from the impact site and fell as a light sprinkling throughout central North America soon after the K-T boundary impact.

5. Mineral grains composed of quartz, oligoclase, and microcline and lithic fragments of quartzite, metaquartzite, chert, and granite-like mixtures of quartz, oligoclase, and microcline constitute as much as 2 percent of the impact layer. A minimum of 30 percent of the grains (except chert) in the impact layer contain shock lamellae.
6. Shock-metamorphosed minerals are concentrated in the K-T boundary impact layer and not in the underlying boundary claystone. Only one such concentration of shock-metamorphosed minerals has been observed at or near the K-T boundary, supporting the idea that there was only one impact event. Rare shocked mineral grains in the boundary claystone and the underlying carbonaceous claystone were introduced along paleoroot systems.
7. Unshocked lithic grains of quartzite, metaquartzite, and chert are the main components of the impact layer, but they comprise only a few percent of the grains in rocks below the boundary claystone. This remarkable change takes place at the same stratigraphic position at each site in the Raton Basin and at sites in Wyoming, Montana, and Canada. For this reason, these unshocked mineral grains are probably unshocked target material.
8. The impact layer contains larger and a significantly greater number of shock-metamorphosed minerals in Western North America than outside North America. The compositional types of shock-metamorphosed materials in the impact layer are thought to reflect the composition of target rocks at the impact site. Because of the large amount of quartz, quartzite, metaquartzite, and chert, the target rocks were probably quartzose metasedimentary and sedimentary rocks. Small amounts of shocked feldspar and granite-like mixtures of feldspar and quartz suggest that the target rock was, in part, granitic.
9. Shock-metamorphosed minerals in K-T boundary rocks are of impact and not of volcanic origin. This conclusion was reached from two different sources of data, one from the study of shock-metamorphosed minerals in K-T boundary rocks and a second from study of silicic pyroclastic volcanic rocks. A large percentage of the minerals in the impact layer are quartzite and metaquartzite whose source could not have been silicic pyroclastic eruptions. Multiple intersecting sets of shock lamellae have not been reported by petrologists who have intensively studied minerals of silicic pyroclastic rocks.
10. The impact layer is considerably thinner than the 1-cm-thickness that was visualized for the global fallout layer by Alvarez and others in 1980. At European K-T boundary localities, the impact layer is only about 1 mm thick and at North American boundary localities about 5 mm thick, on average. Thus, the magnitude of the K-T boundary impact event is considerably smaller than that estimated by Alvarez and others in 1980.
11. The Manson, Iowa, impact structure is a likely place where the K-T boundary impact occurred because of (1) the compositional similarity of known and inferred Manson subsurface rocks and shocked K-T boundary minerals, (2) the large size (about 35 km diameter) of the impact structure, (3) the essentially identical $^{40}Ar/^{39}Ar$ ages (66 Ma) of shocked feldspar from Manson granitic rocks and sanidine from bentonites that overlie the K-T boundary coal bed in Canada and Montana, and (4) the proximity of the Manson impact structure to Western North American K-T boundary sites that contain relatively abundant and large grain-size minerals.

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R.R. Anderson, of the Iowa Geological Survey, provided me with core from the Manson 2A core hole and made available unpublished information pertaining to the Manson, Iowa, impact structure.

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