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$^{40}Ar/^{39}Ar$ Age of the Manson Iowa Impact Structure and Coeval Impact Ejecta in the Crow Creek Member of the Pierre Shale, South Dakota and Nebraska

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Abstract

A new set of 34 laser total-fusion $^{40}Ar/^{39}Ar$ analyses of sanidine from a melt layer in the Manson impact structure (MIS) in Iowa has a weighted-mean age of 74.1 ± 0.1 Ma. Our age for the sanidine is about 9.0 Ma older than $^{40}Ar/^{39}Ar$ ages of shocked microcline from the MIS reported previously by others. The 74.1 Ma age for the sanidine, which is a melt product of Precambrian microcline clasts, indicates that the MIS played no part in the K-T mass-extinction at 64.5 Ma. In addition, incremental-heating $^{40}Ar/^{39}Ar$ ages of the sanidine show that it is essentially free of excess ^{40}Ar and has not been affected by post-crystallization heating or alteration. An age spectrum of the matrix of the melt layer shows affects of ^{39}Ar recoil, including older ages in the low-temperature increments and younger ages in the high-temperature increments. At 17 places in South Dakota and Nebraska, shocked quartz and feldspar grains are concentrated in the lower part of the Crow Creek Member of the Pierre Shale (Upper Cretaceous). These grains are largest (3.2 mm) in southeastern South Dakota and decrease in size (0.45 mm) to the northwest, consistent with the idea that MIS was their source. The ubiquitous presence of shocked grains in a thin calcarenite at the base of the Crow Creek Member suggests it is an event bed recording an instant of geologic time. Ammonites below and above the Crow Creek limit its age to the zone of *Didymoceras nebrascense* of earliest late Campanian age. Plagioclase from a bentonite bed in this zone in Colorado has a $^{40}Ar/^{39}Ar$ age of 74.1 ± 0.1 Ma commensurate with our sanidine age of 74.1 Ma for the MIS. $^{40}Ar/^{39}Ar$ ages of bentonite beds below and above the Crow Creek are consistent with our 74.1 ± 0.1 Ma age for the MIS and limit its age to the interval 74.5 ± 0.1 to 73.8 ± 0.1 Ma. Recently, two origins for the Crow Creek have been proposed--eastward transgression of the Late Cretaceous sea and a MIS impact-triggered tsunami. Both hypotheses have drawbacks; nevertheless, we conclude that most data are in accord with an impact origin for the Crow Creek Member and are at odds with the marine transgression hypothesis.

Introduction

A buried impact structure (~35 km diameter) near Manson, Iowa, has attracted the attention of geologists for more than 80 years. Renewed interest in this structure followed the proposal of Alvarez and others (1980) that the Cretaceous-Tertiary (K-T) boundary mass-extinction event was caused by the impact of a large asteroid, calculated to be about 10 km in diameter. Chemical and physical evidence that supports their hypothesis has been found globally at the exact K-T boundary. The evidence includes 1) an iridium anomaly (Alvarez and others, 1980); 2) shocked quartz, feldspar, quartzite, and zircon (Bohor and others, 1984, 1992; Izett and Pillmore, 1985; Izett, 1987; Izett, 1990); 3) relic tektites (Izett and others, 1990; Izett, 1991a; Sigurdsson and others, 1991); and 4) Ni-rich spinel (Bohor and others, 1986; Robin and others, 1992). The Alvarez hypothesis triggered a search for the location of the impact, and the most promising impact structures turned out to be the Chicxulub on the Yucatan Peninsula (Penfield and Camargo, 1981; Hildebrand and others, 1991) and the Manson in Iowa (Izett, 1990; Anderson and others, 1996). Chicxulub has emerged as the leading candidate, and Manson has been excluded because it has been $^{40}Ar/^{39}Ar$ dated at 73.8 Ma (Izett and others, 1993b). This age for the Manson impact structure (MIS) is 9.3 million years older than the 64.5 Ma age of tektites (Izett and others, 1991a) in a bed that precisely marks the K-T boundary on Haiti.

Two different methods have been used to correlate impact structures with their distal ejecta. An example of the first method, which was used in the present study, involved two steps. First, the isotopic age of the Manson structure was established and second a prediction was made that distal impact ejecta would be found in sediments of the Cretaceous Western Interior seaway at an appropriate biostratigraphic level. The fact that this prediction was fulfilled strengthens the idea that impact ejecta found in the Crow Creek Member had its source at the Manson impact structure (MIS). In a second method, traces of material of impact origin (iridium) were found serendipitously in K-T boundary sedimentary rocks that record a major extinction event (Alvarez et al., 1980). Eleven years after this discovery, the source impact structure for the iridium was identified on the Yucatan Peninsula (Hildebrand et al., 1991). Ironically, this site (Chicxulub) had been discovered previously by Penfield and Camargo (1981), but ignored by crater specialists. Similarly, in a search for dateable volcanic material in middle Proterozoic marine rocks of the Adelaide Geosyncline, Gostin et al. (1986) recognized a layer of impact ejecta at Bunyeroo Gorge. They then linked this impact-ejecta layer with its source crater, the Lake Acraman impact structure of the Gawler Ranges, South Australia (Williams, 1986).

Since Izett and others (1993b) published their preliminary age of 73.8 Ma for the MIS we have acquired a new set of $^{40}Ar/^{39}Ar$ analyses of sanidine from core of recent drilling of the MIS. We herein report the ages and associated analytical data that establish a reliable age for the structure. In addition, we describe impact ejecta (shocked mineral grains) in the Crow Creek Member of the Pierre Shale (Upper Cretaceous) in South Dakota and Nebraska that arguably is fallout from the MIS event. Ammonites from below and above the Crow Creek indicate the member was deposited in the range zone of *Didymoceras nebrascense* of earliest late Campanian age. We also report $^{40}Ar/^{39}Ar$ ages for bentonite beds in the Pierre Shale just below and above the Crow Creek Member that are consistent with our age for the MIS and provide an independent check on its age. Two origins for the member are briefly evaluated.

Manson Impact Structure

About 40 years ago, drilling and geophysical work showed that a complex circular structure lies buried beneath a thin (~30 m) blanket of Pleistocene glacial drift in northcentral Iowa. Data then available indicated that the structure consisted of a relatively large central zone composed of Proterozoic granite and gneiss surrounded by an annular zone of faulted Phanerozoic sedimentary rock. The Proterozoic crystalline rocks had been uplifted at least 5 km above their normal basement position. Hoppin and Dryden (1958) thought that the structure formed by cryptovolcanic processes; however, Short (1966) showed that the crystalline rocks of its core were shock-metamorphosed, the result of the impact of a large asteroid or comet. To explore the structure, a 146 m deep core hole (Manson 2A) was drilled in 1953 in the Proterozoic crystalline rocks near the center of the structure (Hoppin and Dryden, 1958).

This buried structure became an early candidate for the K-T boundary impact site when French (1984) suggested that the impact occurred either at the Sierra Madre, Texas, or Manson, Iowa, structures. He considered Manson to be the more likely because of the large size (0.1 mm) of shocked minerals found in K-T boundary sedimentary rocks in Montana by Bohor and others (1984). Izett (1987, 1990) also suggested that the MIS might be the impact site after recovering substantially larger mineral grains (0.64 mm) from K-T boundary sedimentary rocks in Colorado, New Mexico, and Montana.

Shoemaker and Izett (1992) speculated that the K-T boundary mass extinction was caused by multiple impacts rather than a single event. In fact, two lithologically different claystone layers separated by a sharp contact constitute the K-T boundary sequence at most Western Interior sites (Izett, 1990). On the basis of this evidence, Izett (1991b) and Shoemaker and Izett (1992) proposed that altered microtektites in the lower kaolinte claystone layer originated from the Chicxulub impact structure and that shocked mineral grains in the upper smectite-kaolinite claystone layer originated from the MIS. This idea, although appealing, is incorrect, and the origin of the sharp contact separating the two lithologically different claystone layers remains a mystery. Several explanations for the origin of the two claystone layers (for example Alvarez and others, 1995) have been offered, but they fail to account for the strikingly different texture and composition of the two claystone layers.

Previous Isotopic Age Studies

The first attempt to isotopically date the MIS failed to produce a precise result. C.W. Naeser and G.A. Izett obtained a fission-track apatite age of 60.5 Ma for Precambrian gneiss from the Manson 2A core hole (Izett, 1990). Although the analytical uncertainty was large (± 18 Ma), nonetheless, the measured age is consistent with the post-Late Cretaceous and pre-Pleistocene stratigraphic age (Anderson and Witzke, 1994) of the Manson structure.

In a second attempt Kunk and others (1987) used the $^{40}Ar/^{39}Ar$ incremental-heating technique to analyze shocked microcline from Proterozoic basement crystalline rocks of the Manson 2A core (70-m-depth). Their interpretation of the resulting age spectra was that the MIS formed at 65.7 ± 1.0 (± 2σ) Ma, and this age is roughly compatible with the Late Cretaceous to Pleistocene stratigraphic age for the structure (Hoppin and Dryden, 1958). This preliminary age supported the working hypothesis of Izett (1990) that the MIS was the source of shocked mineral grains in K-T boundary rocks at North American Western Interior sites. Subsequently Kunk and others (1989) obtained a second set of ages for shocked microcline from the 131-m-depth of the Manson 2A core. Their interpretation of these new $^{40}Ar/^{39}Ar$ age spectra again was that the microcline underwent a shock-induced thermal event and significant argon loss at 65-66 Ma.

Despite these age studies other emerging evidence indicated that the MIS is not of K-T boundary age. Steiner and Shoemaker (1993) determined that certain Manson crater-fill rocks have normal remanent magnetization. Taken at face value, this data excludes the possibility that the Manson impact event occurred exactly at the K-T boundary because the rocks that contain the boundary are reversely magnetized (Shoemaker and others, 1987). In addition, U-Pb ages of shocked zircons from the upper of two K-T boundary claystone layers in the Raton Basin of Colorado are not compatible with the age of Precambrian crystalline basement rocks in the Manson, Iowa, area, or sedimentary rocks derived from the erosion of such rocks (Bohor, and others, 1992; Krogh and others, 1992; Premo and Izett, 1993). Although their data base was small, Blum and others (1993) determined that the isotopic composition (Rb/Sr, Nd/Sm, $\delta^{18}O$) of K-T boundary tektites from Haiti and MIS melt rocks are significantly different.

Because the ages of Kunk and others (1989) were suspect for reasons noted above, it became important to obtain samples of a theoretically possible melt layer to establish a reliable age for the MIS. Twelve holes were drilled in the MIS crater fill rocks in 1991-92 to search for a possible melt layer, to find fossiliferous post-impact crater fill sediments, and to explore mineralogical-geochemical variations. One of the 12 drill holes (M-1, 216 m total depth) was drilled 6 km northeast of the center of the MIS and on the flank of its Proterozoic crystalline rock central peak (see Anderson and others, 1996, Figure 3, for location of drill holes). Core from the M-1 hole proved to be the most interesting and promising target for isotopic age studies.

Description of M-1 Core

Several names have been applied to the polymict breccias of the M-1 core. Anderson and others (1994) called the rocks igneous clast breccia with sand matrix, igneous clast breccia with glass matrix, and sedimentary clast breccia. More recently, Anderson and others (1996) referred to the same rocks as suevite breccia, impact melt breccia, Keweenawan shale-clast breccia, and

Phanerozoic clast breccia. Koeberl and others (1996) called the rocks impact melt breccia, suevite, and fragmental breccia. Bell and others (1996) referred to them as melt-poor suevite with coarse matrix, impact melt breccia with crystalline matrix, impact melt breccia with fine-grained to glassy matrix, and sedimentary clast breccia. These authors used the term "suevite" for some units of the M-1 core, but their use of this term is odd because M-1 rocks are devoid of aerodynamically shaped glass bombs, a necessary component of suevite (Bates and Jackson, 1980). Moreover, no glass has been identified or recovered in any samples of Manson core studied by us; nevertheless, Anderson and others (1994), Bell and others (1996), and Koeberl and others (1996) maintained that the M-1 core contains glassy matrix.

The M-1 core contains a distinctive 40-m-thick interval (107-147 m depth) composed of a dense, fine-grained rock containing small clasts. On the basis of its petrographic characteristics, including haloes of clinopyroxene crystals around shock-metamorphosed quartz grains, Izett and others (1993a) suggested that the rock is melted and partially melted polymict breccia. Clinopyroxene haloes around shocked quartz grains are a common feature of melt rocks at impact structures (Floran and others, 1978, Fig. 3a). Another key feature of the M-1 core is that large potassium feldspar clasts (Fig. 1) from the bottom (216 m) of the hole upward to the 150-m-depth are pink and fresh-appearing microcline, although microscopically the microcline has shock-metamorphic features. In the 40-m-thick impact melt rock, microcline and other feldspar clasts are chalky white as a result of being partially melted.

The matrix of the melt rock is a greenish-black flinty material, and scanning electron microscopic (SEM) examination shows that it consists mainly of microcrystalline feldspar, quartz, and clinopyroxene. Small (5-50 μm) crystals of titanomagnetite dispersed in the matrix and in quartz haloes around shocked quartz grains account for the magnetic properties of the rock (Izett and others, 1993a). Temperatures (760° C) calculated for coexisting titanomagnetite and ilmenite (Fig. 2) crystals in quartz haloes surrounding shocked quartz grains were apparently sufficient to melt or partially melt some small breccia clasts. Some ilmenite crystals are skeletal (Fig. 3).

Although we did not find glass suitable for isotopic dating, a more promising material, sanidine, which is an excellent K-Ar geochronometer, was identified in thin sections of the matrix of the melt lens in the M-1 core (Fig. 4) . The presence of sanidine in the melt matrix spurred a search for large microcline clasts that had been converted to sanidine. We identified two chalky-white, microcrystalline feldspar clasts several centimeters in diameter, one at a depth of 114 m and a second at the 143 m in the M-1 core. The chalky white feldspar clasts (Figs. 1B and 4) consist of spherulitic, radially bladed masses of sanidine (11.4 ± 0.7 wt.% K_2O; $Or_{68}Ab_{31}An_1$) and trace amounts of quartz (Izett and others, 1993b). Their spherulitic texture is similar to that of melt rock in the Manicouagan impact structure described by Floran and others (1978, Fig. 4C). The texture and composition of the sanidine in the M-1 melt rock imply that it crystallized from a liquid derived from melted Precambrian microcline clasts.

$^{40}Ar/^{39}Ar$ Ages of M-1 Core

Kunk and others (1993) analyzed shock-metamorphosed microcline and melt rock of the M-1 core to obtain a third set of incremental-heating $^{40}Ar/^{39}Ar$ ages. Their interpretation of the new age spectra was again that the rocks have an overall apparent age of 65.4 ± 0.4 Ma. In a companion study, Zeitler and Kunk (1993) noted that their age spectra of shocked microcline were saddle shaped suggesting minimum ages in the range 67-72 Ma. Izett and others (1993b) used the laser total-fusion $^{40}Ar/^{39}Ar$ method to obtain a weighted-mean age of 73.8 ± 0.3 Ma for a sanidine (melted Precambrian microcline) clast from the 114-m-depth of the M-1 core. This age is 9.0 million years older than the ages of Kunk and others (1989, 1993), of Zeitler and Kunk (1993), and the 64.4 Ma age of the K-T boundary (Dalrymple and others, 1993). Moreover, incremental-heating $^{40}Ar/^{39}Ar$ ages (Dalrymple and Izett, 1994) of the sanidine and the matrix of the M-1 core melt rock imply that the Manson impact structure is ~74 Ma. Zeitler (1996) recently obtained new incremental-heating ages for shocked-metamorphosed Precambrian microcline from the M-1 core and commented on previous attempts at dating the shocked microcline. His combined data gave an age of 73.3 ± 0.3 (1σ) Ma, significantly less than our new age for the Manson structure, if the difference in ages of the fluence monitors used is taken into account.

From 1993 to 1996, we obtained five new sets of $^{40}Ar/^{39}Ar$ analyses using methods and equipment similar to those used by Izett and others (1993b), except for two differences. Small chips of degassed zero-age basalt were added to each aliquot of sanidine dated in the subsequent five sets to enhance melting of sanidine. The ages of Izett and others (1993b) were measured using sanidine from a depth of 114 m, whereas in the subsequent analyses, 14 analyses were measured using sanidine from the 143-m-depth. The analytical methods used in this study are described in Appendix A, and analytical data are in Appendices B-1 to B-5.

One purpose for making additional analyses was to obtain a more precise age for the M-1 sanidine, but just as important we wanted to explore the range of ages from new independent irradiations of the same material. We also wanted the Manson M-1 sanidine to be precisely dated so it could be used as a secondary standard for dating bentonite beds below and above Manson impact ejecta in marine Campanian-age sedimentary rocks (Pierre Shale) in South Dakota. The six sets of new data (Table 1) can be identified by their irradiation numbers as follows: GLN9, JDO13, GLN12, GLN13, JDO21, and JDO22. The means and standard deviations (1 σ) for the six sets of ages listed sequentially are: 74.17 ± 0.50, 74.11 ± 0.49, 74.17 ± 0.57, 74.18 ± 0.31, 74.15 ± 0.47, and 74.00 ± 0.27 Ma. Statistically there is no difference between the mean ages of the six different groups at the one sigma level, and we conclude that not much precision is gained by making an overly large number of analyses. We infer that the methods of determining J and correcting for undesirable reactor-produced Ar isotopes does not contribute any significant error to the analyses.

The weighted-mean age of the 34 new analyses of the Manson M-1 sanidine is 74.1 ± 0.1 Ma (Table 1, Appendix B-1). The standard error of the mean (95% C.L. level) for the 34 analyses is ± 0.14 Ma. Five analyses of the Manson M-1 sanidine were not included in the final summary (Table 1, Appendix B-1) because the ages were either two sigma below or above the weighted-mean age. Nevertheless, had these five analyses been included, the resulting weighted-mean age was nearly identical to that for the 34 selected analyses of 74.1 Ma.

To determine if the Manson M-1 sanidine might contain excess ^{40}Ar trapped in its lattice during crystallization, we plotted the analytical data for the 34 ages on an inverse-correlation diagram (Dalrymple and others, 1988). On this diagram (Fig. 5), $^{36}Ar/^{40}Ar$ is plotted on the ordinate and $^{39}Ar/^{40}Ar$ (in this study multiplied by 1/J to normalize for different J values) is plotted on the abscissa. The computed age using the inverse of the intercept on the abscissa (23.8132) is 74.2 ± 0.1 Ma (95% C.L.). This age is statistically the same as the weighted-mean age for the 34 selected total-fusion sanidine ages of 74.1 ± 0.1 Ma. The inverse of the intercept on the ordinate indicates that the trapped argon component has a ^{40}Ar to ^{36}Ar ratio of 290.8 ± 8.5 (Fig. 5), and this ratio is statistically the same as the ratio of these isotopes (295.5) in atmospheric argon. Thus, the data indicate that the sanidine does not contain a significant amount of excess ^{40}Ar, and the calculated ages provide an accurate crystallization age.

We also obtained three incremental-heating ages (19 and 22 heating steps) for the M-1 sanidine (Fig. 6), two from the 143-m-depth and one from the 114-m-depth. The analyzed samples were small (~1 mg) to optimize sample homogeneity relative to size. The two samples from 143-m-depth have simple age spectra with prominent intermediate-temperature plateaus, although there is a slight increase in the increment ages in the high-temperature steps. Both plateaus have the same apparent age of 73.5 ± 0.4 Ma. The correlation diagrams (not illustrated) for these two experiments have intercepts on the ordinate indicating ^{36}Ar to ^{40}Ar ratios of 294.4 ± 3.2 and 294.6 ± 2.0, respectively, and these are statistically indistinguishable from atmospheric argon (^{40}Ar /^{36}Ar = 295.5). Accordingly, we infer that the sanidine does not contain extraneous ^{40}Ar and that the plateau ages are the cooling and crystallization age of the sanidine.

Individual increment ages of the sanidine from 114-m-depth appear to increase slightly but systematically with increasing ^{39}Ar release (Fig. 6). The age spectrum has a near plateau (9 of 22 steps) from 725 °C to 1,050 °C, and these 9 steps contain 61% of the ^{39}Ar released. The age for the 9 steps is 74.0 ± 0.4 Ma. The correlation diagram (not illustrated) for this sanidine has a ^{40}Ar/^{36}Ar intercept of 257.7 ± 23.1, which is statistically indistinguishable from atmospheric-argon composition. Accordingly, we infer that the sanidine does not contain extraneous ^{40}Ar.

We also obtained an age spectrum for a tiny fragment (1.058 mg) of melt matrix adjacent to the dated sanidine clast from the 143-m-depth in the M-1 core (Fig. 6). The spectrum has characteristics of ^{39}Ar recoil, including older ages in the low-temperature increments (~0-20% ^{39}Ar released) and younger ages in the high-temperature increments (~75-95% ^{39}Ar released). Theoretical and experimental data show that this pattern results from the recoil (during irradiation) of ^{39}Ar out of high potassium and less retentive phases, such as fine-grained feldspar, mica, and glass, and implantation in low potassium refractory phases such as olivine and pyroxene (Turner and Cadogan, 1974; Huneke and Smith, 1976). The increments affected by recoil have no age significance, but intermediate temperature plateaus from such spectra typically yield crystallization (cooling) ages. In detail, the age spectrum for the M-1 melt matrix has four intermediate temperature increments that form a plateau having an age of 75.0 ± 0.4 Ma. This apparent age is slightly older than the total-fusion age (74.1 Ma) of the M-1 sanidine. We think that this apparent age for the melt matrix is due perhaps to the presence of small older clasts and do not think that it represents the age of the Manson impact event. Total-fusion ages for five fragments of the melt matrix range from 73.3-78.1 Ma (not shown in Table 1), strengthening the hypothesis that the ages of the melt matrix are adversely affected by recoil and older clasts.

On an inverse-correlation diagram using the data from the incremental-heating experiment (Fig. 7), the intercept on the abscissa indicates the age of the melt matrix is 76.5 ± 1.8 Ma (1 σ). This apparent age is slightly older than the total-fusion age (74.1 Ma) of the sanidine from the M-1 drill core, but actually it is statistically no different because of the large analytical uncertainty of 1.8 Ma. The large uncertainty resulted for the most part from using a tiny fragment of the melt matrix. The inverse of the intercept on the ordinate indicates that the trapped argon component has a ^{40}Ar to ^{36}Ar ratio of 284.7 ± 13.2, and this value is statistically indistinguishable from the atmospheric argon composition (^{40}Ar/^{36}Ar = 295.5).

In summary, the new 34 ^{40}Ar/^{39}Ar ages of sanidine, which is unquestionably an impact melt product of microcline, provide consistent results and indicate that the sanidine cooled and crystallized at 74.1 ± 0.1 Ma relative to an age of 27.92 Ma for Taylor Creek Rhyolite based on an age of 513.9 Ma for Mmhb-1 hornblende fluence monitor. Many geochronologists use an age of 520.4 Ma for Mmhb-1 hornblende, and this value results in an age of 75.2 Ma for the sanidine clast in the melt rock of the Manson impact structure. We consider the 34 new ages of this report to be more reliable than the 12 ages (73.8 ± 0.3 Ma) for the MIS sanidine reported by Izett and others (1993b) because in that study basalt glass was not used as a flux, and the sanidine was only partially melted. Results of $^{40}Ar/^{39}Ar$ incremental-heating resistance furnace analyses of sanidine are statistically the same as the laser total-fusion analyses. The data show that the sanidine is essentially free of excess ^{40}Ar and has not been affected significantly either by post-crystallization heating or alteration. Incremental-heating and total-fusion analyses of melt matrix from the 143 m depth show effects of ^{39}Ar recoil and small much older undegased clasts. We attribute no reliable age significance to our results for the melt matrix.

Crow Creek Member of the Pierre Shale

In eastern and central South Dakota and northeastern Nebraska, a silty and sandy laminated limestone and an overlying uniform marlstone constitute a conspicuous pair of beds in the lower part of the Pierre Shale (Upper Cretaceous, Campanian). Together they are generally about 3 m thick. Crandell (1950) named these two units the Crow Creek Member of the Pierre Shale for exposures at the mouth of Crow Creek in Buffalo County[1] and remarked that the formation had been the subject of considerable unpublished discussion and speculation. Recently the formation has aroused more interest after it was found to contain noteworthy amounts of shocked mineral grains arguably derived from the Manson, Iowa, impact structure (Izett and others, 1993b).

Although a thin stratigraphic unit making up only a few percent of the total thickness of the Pierre, the Crow Creek Member is widely distributed and extends about 300 km from northeast of Yankton to Fort Pierre, mainly along the Missouri River and its tributaries. Within its outcrop area, the Crow Creek is locally absent in the Big Bend region of the Missouri River. A search for the member in equivalent-age rocks around the Black Hills, southern North Dakota, and northwestern Nebraska, along the Chadron arch of northwest Nebraska, and in eastern Colorado was unsuccessful. In the subsurface, we identified the Crow Creek on geophysical logs of oil and gas test wells drilled in central and southeastern South Dakota. These logs reveal that the member extends 100 km west of Chamberlain and 100 km northwest of Pierre (see also Rice, 1977; Bretz, 1979). The Crow Creek underlies an area of at least 30,000 square miles. Place names of this report are all in South Dakota unless otherwise noted.

The Crow Creek Member has two facies; one is northwest and the second southeast of Wheeler on the Missouri River (Crandell, 1952, p. 1754). The northwestern facies consists of two units of unequal thickness. The lower unit is silty and sandy limestone typically 25-38 cm thick, although sites where it is 66 cm thick have been reported (Bretz, 1979). Optical and SEM examination of the rock indicates it is composed dominantly of fine-grained calcium carbonate matrix and disseminated detrital grains of calcite, many of which are of fine sand size. Acetic acid insoluble residues of the calcarenite show that it commonly has ~20% detrital grains of dolomite. Schultz (1965) reported that the calcarenite contains 15-35% dolomite on the basis of X-ray diffraction analyses. Also disseminated in the rock are detrital grains of quartz (~10%) and minor to trace amounts of feldspar (microcline and plagioclase), phosphate, mica, chert, and silicate heavy minerals. Many of the quartz, calcite, and dolomite grains are 75-200 μm (very fine to fine sand), as measured in thin sections, but most are of silt size. The rock also contains scattered

[1] Place names of this report are all in South Dakota unless otherwise noted.

calcareous microfossils in the carbonate matrix, generally concentrated in the lower part of the unit along coarse-grained laminae. At the Black Dog locality southwest of Chamberlain (Table 2, Fig. 8), an impression of a palm frond about 25 cm in diameter was found on a bedding plane surface. Polished blocks of the lowermost part of the unit commonly show abundant shreds of carbonaceous material (0.1-0.5 mm) aligned parallel to bedding.

Some authors have called the silty and sandy unit of the Crow Creek a sandstone, notwithstanding the facts that it contains 1) an HCl-insoluble residue generally less than 50% and 2) an assemblage of detrital quartz grains that on average are silt rather than sand size. Witzke and others (1996) pointed out that the rock has a wackestone to packstone fabric (Dunham, 1962). According to the classification of Folk (1959) the rock is a fossiliferous biomicrite. For brevity and to emphasize the fact that it contains significant amounts of detrital dolomite, calcite, and quartz, we herein refer to the rock as a calcarenite (Grabau, 1903).

Crandell (1952) proposed that the source of detrital quartz grains in the calcarenite was from the erosion of older units, such as the Codell Sandstone (Upper Cretaceous) of eastern South Dakota. However, our analysis of Codell samples showed that it generally lacks significant amounts of microcline, and thus the detrital microcline grains in the Crow Creek may have had a source different from the Codell, most likely the Manson impact structure.

The lowest part of the calcarenite of the northwestern facies is locally foreset-bedded, and displays cut and fill stratification (Crandell, 1952). The upper part is commonly horizontally laminated (1-10 mm), and the laminations seem to be grain-size controlled as seen in thin sections and plastic impregnated polished blocks. At the Black Dog site (Fig. 8, Table 2) southwest of Chamberlain, the lower 3 cm of the calcarenite contains a sequence of alternating fine and coarse grained laminae. The coarse-grained laminae are more porous than adjacent fine-grained laminae.

Slopes below the Crow Creek calcarenite are littered with tabular blocks having small scale cross-laminations (<25 cm) and other bed forms similar to hummocky cross stratification. Although hummocky cross stratification (Harms and others, 1975; Bourgeois, 1980; Dott and Bourgeois, 1982; Dott, 1983; Duke, 1985) has not been produced in laboratories (Duke, 1985), these authors stated that it typically occurs in shelf deposits that record storm surges associated with hurricanes, intense winter storms, and possibly tsunamis. Klein and Marsaglia (1987) warned that equating hummocky cross stratification with hurricanes to the exclusion of extratropical storms or tsunamis is premature.

Well exposed outcrops of the Crow Creek are not common, but in bulldozer excavations of the calcarenite at Oacoma (Table 2), some large blocks have well developed hummocky cross stratification. A few of these blocks have unusual large, sinuous, asymmetric ripple marks that have maximum wave lengths of 50 cm and amplitudes of 25 cm. Typically, the thickness of the layering systematically decreases from about 1 cm to less than a millimeter as a ripple crest is approached. At the Chamberlain site (Table 2, Fig. 8) some large slabs of the calcarenite display well developed parting lineation on bedding plane surfaces. The observations that the Crow Creek calcarenite underlies a large area of central South Dakota, is only about 30 cm thick, and has hummocky cross stratification and locally large ripple marks raise the possibility that the calcarenite records a high-energy event such as a megastorm or tsunami.

The upper unit of the Crow Creek in the northwestern facies consists of 2-3 m of massive marlstone much like the upper part of the Crow Creek in the southeastern facies described in the following paragraph. Thin sections show that it is a uniform sparse biomicrite containing scattered skeletal calcareous fossils and trace amounts of detrital quartz grains, some of which are shock metamorphosed.

The Crow Creek in the southeastern facies constitutes a single lithologic unit 2-3 m thick composed of marlstone lacking the distinctive basal laminated calcarenite of the northwestern facies. The equivalent of the basal calcarenite in the southeastern facies consists of 10-15 cm of marlstone that is a sparse biomicrite. Point counting of thin sections of the rock shows that it typically contains ~15% light-gray marlstone and shale chips (1-2 mm) that are oriented parallel to bedding. The basal Crow Creek not only contains the small chips but others as large as 11 cm (Witzke and others, 1996) composed of similar lithologies. Schultz (1965, p. B12) noted that the abundance of these large shale fragments is much greater in the southeastern facies.

The lower part of the marlstone also contains disseminated grains of quartz (5-10%), calcite and dolomite (10%), feldspar (<2%), and trace amounts of other mineral grains, including mica, glauconite, and, phosphate, composite quartz and feldspar, and silicate heavy minerals. The quartz, carbonate, and feldspar grains range in size from 30 μm to as large as 4.0 mm. The basal 5 cm of the Crow Creek Member at the Niobrara State Park site (Fig. 8, Table 2) contains several types of bivalve shell fragments (up to 1.0 cm), foraminifera, ostracoda, sponge spicules, fish scales and teeth, and wood fragments.

Near Yankton the lowest few centimeters of the Crow Creek contains gray phosphatic nodules (up to 6 cm) having highly polished surfaces as noted by Simpson (1960). In this area, the Crow Creek overlies the Sharon Springs Member, and it is probably the source of the nodules. Curiously, a few nodules have shapes identical to those of spherical- and dumbbell-shaped tektites, however, most have irregular shapes not typical of tektites.

The Crow Creek Member is bounded by unconformities. The lower contact separates noncalcareous shale of the Gregory Member from overlying calcarenite of the basal Crow Creek at most places in South Dakota. East of Yankton and in western Iowa the Crow Creek is assumed to overlie the Niobrara Formation of Late Cretaceous age. All who have studied this contact have concluded that it is a disconformity. In southeasternmost South Dakota and northeastern Nebraska, the Crow Creek overlies the Sharon Springs Member and the Gregory Member was either not deposited or removed by erosion. Crandell (1952) noted that channeling of the Gregory below the disconformable surface has not been observed, although Schultz (1965, p. B15) subsequently described two areas near Wheeler where no more than 5 m of shale was removed allegedly by submarine erosion. Witzke and others (1996) reported that the Crow Creek rests unconformably on the Niobrara Formation south of Mount Vernon (Table 2, Fig. 8); however, in this area we found Crow Creek outcrops resting on the Sharon Springs Member. The upper contact is knife sharp and separates marlstone of the Crow Creek from overlying noncalcareous gray shale of the DeGrey Member. Bretz (1979) observed burrows in the upper 10 cm of the marlstone at 8 of 15 localities studied and suggested that the contact is a diastem. We agree that the upper surface of the Crow Creek is a disconformity, although occupying only a geologically brief time because of biostratigraphic constraints and $^{40}Ar/^{39}Ar$ ages of bentonites below and above the Crow Creek Member.

We recently studied the Gregory and DeGrey contact at two places in southeastern North Dakota. At a highway I-90 roadcut (exit of east-bound lane) in western Valley City, light-gray marlstone of the Gregory is overlain by black, bentonitic shale of the DeGrey Member, and the Crow Creek is absent. The uppermost 10 cm of the Gregory appears to be a disturbed zone below an unconformity. Centimeter-size clasts, veins, and irregular-shaped masses of black to dark-gray shale are mixed with Gregory marlstone. To the northwest near Cooperstown in Griggs County (Hoganson and others, 1996), light-gray marlstone of the Gregory is overlain by black bentonitic shale of the DeGrey, and the Crow Creek is absent.

Crandell (1952) and Schultz (1965) concluded that the disconformity below the Crow Creek was produced by submarine processes. In contrast, Witzke and others (1996) maintained that it was formed by subaerial processes. Baird (1978) proposed criteria for the identification of subaqueous unconformities as follows: 1) Lack of irregularities on the surface such as large channels, 2) Lack of soils and caliche associated with the surface, 3) Obliteration of primary sedimentary or fossil structures, 4) Presence of marine strata below and above, 5) Correlation with a continuous marine stratigraphic sequence, and 6) Presence of marine epizoan on reworked phosphatic pebbles. Application of Baird's (1978) criteria 1-5 to the sub-Crow Creek unconformity indicates that its is of submarine origin. We believe that criterion 4 is especially important because the unconformity separates normal marine rocks of the Gregory and Crow Creek Members, both of which contain suites of marine fossils.

The thickness of the Gregory Member below the sub-Crow Creek disconformity decreases from west to east across South Dakota as follows: 60 m about 100 km west of Chamberlain, 52 m at Pierre, 35 m at Chamberlain, 18 m at Rising Hail, 9 m at Wagner, 2 m at Fort Randall Creek, and 8 m near Niobrara State Park in northeast Nebraska. Near Yankton, the Gregory is not present and the Crow Creek rests on a few meters of the Sharon Springs Member. Schultz (1965) concluded that this eastward thinning results from both depositional processes and submarine erosion. He reasoned that most of the thinning is in the lower calcareous, rather than the upper noncalcareous part of the Gregory, because a zone containing thin bentonite beds persistently occurs in the upper Gregory from Chamberlain to near Platte. Nevertheless, the amount of erosion beneath the disconformity must be appreciable because the basal part of the Crow Creek Member pervasively contains ripup marine Cretaceous shale and marl chips.

In summary, the Crow Creek Member has many lithologic characteristics that have a bearing on its origin and distinguish it from other units in the Pierre Shale. The most significant is detrital grains of dolomite, calcite, and large shock-metamorphosed mineral grains in the member. Other important features include: sharp lower and upper contacts; hummocky cross stratification and large-scale ripple marks; general lack of bioturbation, small-scale ripple marks, and trace fossils; and ripup shale and marlstone chips (up to 11 cm) of marine Cretaceous units in the basal Crow Creek. Currents necessary to move such large fragments must have been vigorous, perhaps 2 m/s (see Twenhofel, 1961, p. 645). The marlstone chips in central South Dakota should be examined to determine if some might contain nannofossils of upper Niobrara Formation age. If confirmed, it would indicate that the marlstone chips were derived from erosion of Niobrara Formation of eastern South Dakota or Iowa at a time when the basal Crow Creek was being deposited.

Shocked Mineral Grains

Izett and others (1993b) found three sites in southeastern South Dakota where the Crow Creek Member contains an assemblage of detrital grains, some of which are quartz and feldspar (microcline and plagioclase) that have multiple sets of planar deformation features (PDF). Many of the microcline grains exhibit patchy extinction or mosaicism. In addition, heavy-mineral concentrates of the insoluble residue contain a few rounded zircon crystals containing sets of PDF's. Subsequently several more sites were found where the basal Crow Creek contains quartz and feldspar detrital grains having PDF's (Izett and Cobban, 1994.) Witzke and others (1996) listed two sites southwest of Chamberlain where the Crow Creek contains quartz and feldspar grains having PDF"s. However, the bedrock exposed at one of their sites, alleged to be along Black Dog Creek southwest of Chamberlain, consists not of the Crow Creek Member (and enclosing Gregory and DeGrey Members) but rather the Verendrye Member.

Currently, 19 places in South Dakota are known where the Crow Creek contains quartz and feldspar grains having PDF's, including sites in Brule, Buffalo, Davison, Gregory, Hughes, Lyman, and Yankton counties and three places in Nebraska in Cedar and Knox Counties, Nebraska (Table 2). The PDF's in quartz (Fig. 9) and feldspar grains in the Crow Creek are similar to those in shock-metamorphosed rocks at known impact structures (Chao, 1968). Such PDF's in quartz and feldspar grains are taken as evidence that the grains experienced high levels of shock associated with the impact of an asteroid or comet. The apparently can form only at pressures greater than 60 kb. Some of the shocked quartz grains lack strictly planar lamellae and instead contain widely spaced microfractures indicative of low-level shock. Others also exhibit pronounced shock-mosaic texture especially the potassium feldspar grains. We also identified millimeter-size, shocked composite grains of quartz and feldspar. The presence of PDF's in well-rounded quartz grains suggests that they were derived from a sedimentary rock.

The amount of shocked mineral grains in the Crow Creek was estimated by point-count techniques of oil-immersion mounts and thin sections. At 12 places (Table 2, Fig. 8) in the northwestern facies, the amount of shocked grains in the calcarenite is low, much less than 1% of the nonclay HCl-insoluble residue. Although shocked grains are most abundant in the lower part of the Crow Creek, nonetheless, they also occur in the overlying marlstone. For example at the Black Dog site (Table 2, Fig. 8), the clay-free HCl-insoluble residue from 1.0 kg of marlstone 1.0 m above the calcarenite contained 0.7 g of quartz and feldspar grains ~0.1 mm in diameter. Of the 0.7 g, about 1.0% have multiple sets of planar deformation features. At several sites in Lyman County, we observed that the relative amount of shocked grains is highest in the coarsest 5% of the basal part of the calcarenite.

At seven places in the southeastern facies of the Crow Creek, the amount of shocked quartz and feldspar grains is variable, but less than 20% of the clay-free HCl-insoluble residue. At most places, the amount is distinctly higher than that in the northwestern facies. The highest concentration of shocked grains recorded was at the Lake Marindahl site (Table 2, Fig. 8). There, the insoluble residue contains about 20% shocked quartz and feldspar grains. The lowest concentrations were found at two places in the southeastern facies. At the East of Niobrara site (Table 2, Fig. 8), the lowest part of the Crow Creek contained ~10% quartz and feldspar grains, of which about 15% have multiple sets of planar deformation features. At Fort Randall Cemetery (Table 2, Fig. 8) near the western limit of the southeastern facies, about 8% of the quartz and feldspar grains have shock lamellae.

The size of shocked mineral grains was estimated in oil-immersion mounts using a petrographic microscope equipped with a calibrated ocular micrometer. Table 2 lists the locations, size of the largest shocked mineral grains measured at each site, and the distance from each site to the MIS. The largest shocked grains (1.5-3.2 mm) are in southeastern South Dakota near Yankton. These sites range from 215 to 315 km from the MIS. By contrast, the smallest shocked mineral grains (0.4-1.2 mm) are from 410 to 475 km from the MIS. In general, the size of the largest shocked grains decreases from Yankton to Chamberlain away from the MIS, although large grains (0.7-0.8 mm) occur at several sites in Lyman County (Table 2).

The above evidence suggests that shocked mineral grains in the Crow Creek were derived from the east, probably at the MIS, and were transported either ballistically or by high level atmospheric winds and reworked at their deposition sites. The source for most of the unshocked mineral grains was partly from the erosion of older rocks of eastern continental North America adjacent to the Cretaceous seaway and partly from unshocked rocks at the MIS.

In summary, the calcarenite at the base of the Crow Creek Member records the abrupt appearance in the stratigraphic column of shocked mineral grains that we believe were derived from the MIS. Thus, the calcarenite is an event bed equivalent in age to the Manson impact, and it records an instant of geologic time.

Biostratigraphic Setting of the Crow Creek Member

The Pierre Shale, which includes the Crow Creek Member, underlies much of South Dakota, and to the casual observer it is a thick, monotonous sequence of dark gray shale. Stratigraphic studies, however, have shown that it consists of many types of shale, calcareous mudstone and shale, marlstone, numerous thin bentonite beds, many different types of concretions, and marine invertebrates and vertebrate fossils. The Pierre includes, in ascending order, the Sharon Springs, Gregory, Crow Creek, DeGrey, Verendrye, Virgin Creek, Mobridge, and Elk Butte Members. Four of these members (Gregory, Crow Creek, DeGrey, and Mobridge) contain beds of calcareous mudstone and shale and, in some instances marlstone.

Two features of the Pierre Shale are relevant to understanding the stratigraphic setting of the Crow Creek Member and its origin. First, the formation thins from west to east across South Dakota. At a principal reference section at Redbird, Wyo. (Gill and Cobban, 1966), the Pierre is 945 m thick. West of Pierre (T. 4 N., R. 18 E.), it is only 440 m thick (Rice, 1977). Near Pierre, the type area, the formation is 300 m thick, and 195 km east of Pierre at Iroquois, South Dakota, it is 76 m thick (Dyman and others, 1994). In northeastern Nebraska near Niobrara State Park where the upper part of the Pierre has been removed by erosion, the remaining Pierre may be ~200 m thick. The original eastern extent of the Pierre is not known, but Anderson and Witzke (1994, Fig. 8) questionably identified the Niobrara Formation and 21 m of lower Pierre above a normal Lower to Upper Cretaceous sequence of Dakota Formation, Graneros Shale, Greenhorn Shale, and Carlile Shale in cuttings from a well on the rim of the buried MIS. However, Witzke (oral commun., 1996) no longer believes the upper shale should be assigned to the Pierre.

Second, the Pierre contains numerous unconformities, including the one at the base of the Crow Creek Member and, in places, concentrations of stratigraphically associated phosphatic nodules. Baird (1978) noted that such nodules are commonly associated with unconformities. Several unconformities in the Sharon Springs Member were described by Martin (1996) and Stoffer and Chamberlain (1996) on the east flank of the Black Hills. Along the Cheyenne River southeast of Hot Springs, we observed concretions in a shale unit containing the ammonite *Didymoceras nebrascense* overlain by another shale unit shale containing *Baculites rugosus* and *Exiteloceras jenneyi*. The contact between the shale units is marked by a zone of phosphatic nodules. Shale that should contain *Didymoceras stevensoni* is not present. Elsewhere in the Western Interior, *D. stevensoni* is a fairly common ammonite, but it has not been found in the Pierre east of the Black Hills. Near Midland, S. Dak., we also found another unconformity near the base of the Mobridge Member, which is marked by phosphatic nodules and the local absence of one ammonite zone, *Baculites grandis.*

Biostratigraphic Age of the Crow Creek Member

The Crow Creek Member is sandwiched between fossil-bearing marine shale of the Gregory and DeGrey Members of the Pierre Shale (Fig. 10; Appendix C). In western South

Dakota and elsewhere in the Western Interior, ammonites are locally abundant in the Pierre, and their study has resulted in a biostratigraphic zonation (Cobban and others, 1994) of the formation. However, ammonites are not common in the Pierre of central South Dakota, and they are especially rare in the southeastern part of the state. Nevertheless, we have been able to extend some of the standard Cretaceous ammonite zones to the central and eastern parts of the state. Unfortunately, the Crow Creek Member is apparently devoid of ammonites and other megafossils useful for biostratigraphic dating, although rare fragments of bivalves, including *Inoceramus* and *Ostrea* and fish bones have been reported (Bretz, 1979).

In South Dakota, marine shale below the Crow Creek Member is assigned to either the Sharon Springs or Gregory Member of the Pierre Shale. At the Red Bird section in eastern Wyoming, the Sharon Springs contains *Baculites obtusus* and *B. mclearni*, and the overlying Mitten Black Shale Member contains *B. asperiformis* (Gill and Cobban, 1966). These three ammonites of early middle Campanian age have not been reported from eastern South Dakota. In the Chamberlain area, the lower part of the Gregory contains, in ascending order, the ammonites *B. perplexus* and *B. gregoryensis* of late middle Campanian age (Gill and Cobban, 1965). We identified the ammonite *Didymoceras cochleatum* and *B. gregoryensis* in collections along the Missouri River in Buffalo County from the middle Gregory Member. In the upper part of the Gregory Member, we collected *B. scotti* of latest middle Campanian age 3-15 m below the Crow Creek (compare Gill and Cobban, 1965). The normally coiled ammonite *Menuites oralensis* (Cobban and Kennedy, 1993) characteristically occurs in the upper part of the *B. scotti* zone, and we collected this taxon from the upper part of the Gregory within 4 m of the overlying Crow Creek near Chamberlain (Appendix C, D13444). In shale 10 m below the Crow Creek at Chamberlain, we collected the heteromorph ammonite *Didymoceras archiacianum* (d'Orbigby). We also collected a fragment of a juvenile whorl of a *Didymoceras*, possibly the species *puebloense* (Cobban and others, 1997) and a single individual of *Oxybeloceras* from shale 25 cm below the Crow Creek at a nearby site in Chamberlain. *D. puebloense* is only known from the zones of *D. nebrascense* and *D. stevensoni* in Colorado and Wyoming.

Marine shale above the Crow Creek constitutes the DeGrey Member, and black-weathering manganese-rich concretions in the lower part of the shale contain rare, poorly preserved ammonites. Although ammonites are rare in these unusual manganese-rich concretions, bivalves are numerous and well-preserved. A fragment of an ammonite found by R.E. Stevenson (retired) of the University of South Dakota (written commun., J.E. Fox, South Dakota School of Mines, 1996) or one of his students places an upper limit on the age of the Crow Creek Member. We identify the ammonite as *Didymoceras nebrascense* of earliest late Campanian age. We follow Kennedy and others (1992) who use a three-fold informal division of the Campanian. The field record slip indicates the fossil was collected in Buffalo County from the Oacoma zone of the DeGrey Member. Our visit to the alleged collection area showed that the Verendrye and Virgin Creek Members compose the surface rocks, and *Didymoceras nebrascense* should not occur in these members (Cobban and others, 1994). Although the exact geographic place in Buffalo County where *D. nebrascense* was collected may never be known, its stratigraphic position is certain on the basis of the stratigraphic information on the record slip and especially the type of matrix attached to the fossil. The matrix is identical to black, worm-bored, manganese-rich concretions in the lowest part of the Oacoma facies of the DeGrey Member.

In black-weathering manganese-rich concretions from the middle and upper parts of the DeGrey shale, we found a late form of *Baculites rugosus* having a weakly ribbed venter and a baculite with a smooth ovate cross section (see also Gill and Cobban, 1965). These bacultes are restricted to rocks in the zone of *Didymoceras cheyennense* of middle late Campanian age. At DeGrey, South Dakota, we found a specimen of *D. cheyennense* near the base of the upper third of the member in a black manganese-rich concretion (Fig. 10). At several places east and south of Fort Thompson, poorly preserved specimens *of B. compressus* occur in the upper 2-3 m of the DeGrey. Locally, the lower and middle parts of the overlying Verendrye Member sequentially contain *B. cuneatus* and *B. reesidei* of latest Campanian age. The position of the Campanian-Maastrichtian boundary varies widely depending on which group of marine fossils is used. On the basis of ammonite correlations from North America to Europe, Kennedy and others (1992) placed the boundary between the zones of *Baculites jenseni* and *B. eliasi* of the Western Interior. The next overlying member, the Virgin Creek of earliest Maastrichtian age contains *Baculites eliasi* in the lower part and *B. baculus* in the upper part. The easternmost record of baculites (*B. eliasi)* in the Pierre Shale was found near Niobrara State Park in northeastern Nebraska. *Baculites grandis* is rare in central and eastern South Dakota, but this taxon was collected from the lowest part of the Mobridge Member of Maastrichtian age near Pierre. On the highest hills north of the White River and west of Chamberlain, white limestone concretions and shale of the Mobridge Member contain *Baculites clinolobatus*. The youngest Cretaceous marine shale in the Western Interior occurs south of Verdigre, Neb. There, sandstone concretions in shale equivalent to the Elk Butte Member contain the ammonite *Jeletskytes nebrascensis*, typically found in the Fox Hills Formation (Timber Lake Member) in its type area (Cobban and others, 1994). The Fox Hills of northcentral South Dakota grades southeastward into the upper part of the Pierre.

The Crow Creek contains a diverse microfossil fauna (Crandell, 1958) consisting of 29 genera of foraminifera (Bretz, 1979), nannofossils, radiolaria, ostracoda, calcispheres, and sponge spicules. A thorough study of the fauna has not been published. Watkins (1989, 1995) and Hammond and others (1995) identified two different calcareous nannofossil assemblages in the Crow Creek. To an indigenous assemblage (*Tranolithus phacelosus* zone), they assigned a late Campanian age compatible with the age ascribed to the Crow Creek Member by Izett and Cobban (1994). To a reworked assemblage (*Aspidolithus parcus* zone) concentrated in the lower part of the Crow Creek, they assigned and early Campanian age. Nannofossils constituting this late Santonian and earliest Campanian assemblage characteristically occur in the upper part of the Niobrara Formation that lies below the Sharon Springs Member. Hammond and others (1995) reported that these reworked nannofossil show little evidence of overgrowth or dissolution and are a major part of the assemblage in the lower part of the Crow Creek Member in Gregory County 60 km from the nearest possible Niobrara Formation source areas.

J. Self-Trail examined samples of the basal calcarenite of the Crow Creek Member from the Black Dog and Oacoma sites near Chamberlain (Table 2). She found that the calcarenite at the Oacoma site contains an abundant, well preserved assemblage of nannofossils, including *Aspidolithus parcus expansus, Eprolithus floralis, Eiffellithus eximius, Reinhardtites anthophorous, and (?)Lithastrinus grilli* (written commun., 1997). The nannofossils most likely are in the micrite matrix of the calcarenite as opposed to rare marl chips. We believe that the reworked assemblage of nannofossils was derived from erosion of upper Santonian and lowermost Campanian rocks, such as the upper Niobrara Formation of eastern South Dakota and Iowa.

Witzke and others (1996) suggested that the source of reworked nannofossils in the Crow Creek Member was from erosion of the Niobrara Formation around the Sioux Ridge of eastern

South Dakota during their postulated marine transgression. For easternmost South Dakota, their proposal seems reasonable because of the proximity of Niobrara Formation source areas to Crow Creek depositional areas. But these authors did not consider the ramifications of possible reworked Niobrara nannofossils in the Crow Creek of central South Dakota at least 100 km from the nearest Niobrara Formation source areas.

Summing up, the Crow Creek Member is underlain by shale of the upper part of the Gregory Member containing, in ascending order the ammonites *Baculites scotti* and *Menuites oralensis* followed by *Didymoceras puebloense(?)*. The Crow Creek is overlain by shale of the lowest part of the DeGrey Member containing *D. nebrascense,* followed successively by shale containing the zones of *D.* cheyennense and *B. compressus.* The Crow Creek is, thus, in the zone of *D. nebrascense* of earliest late Campanian age. A search for remains of the ammonites *Didymoceras stevensoni* and *Exiteloceras jenneyi*, which should occur in the middle of the DeGrey Member, failed to produce a single fragment of these common taxa. Either they are exceptionally rare or rocks of their age are missing in central and eastern South Dakota. The presence of reworked Niobrara-age nannofossils in the basal Crow Creek in the Chamberlain area provides important evidence bearing on the origin of this member.

$^{40}Ar/^{39}Ar$ Age of the Crow Creek Member

Ages of sanidine crystals recovered from bentonite beds in Cretaceous rocks of the Western Interior have been used to construct a useful and reliable $^{40}Ar/^{39}Ar$ time scale (Obradovich, 1994). This time scale, nevertheless, lacks sufficient detail to infer a reliable numerical age for the Crow Creek Member. To bracket its age and to compare the resulting ages with our $^{40}Ar/^{39}Ar$ age for the MIS, we collected kilogram-size samples of bentonite beds from shale below and above the Crow Creek for age analysis. Analytical data for the dated bentonite beds are available in Appendices B-2 to B-5.

A maximum age for the Crow Creek Member is based on $^{40}Ar/^{39}Ar$ analyses of a 10 cm thick bentonite formerly exposed in a bulldozer excavation 3 m below the Crow Creek Member at Oacoma 5 km west of Chamberlain (Table 1, Fig. 8). This bentonite bed is one of a set that generally occurs in the upper 7 m of the Gregory Member in central South Dakota. The ammonites *Baculites scotti* and *Menuites oralenis* were collected from shale and concretions 10-15 m below this bentonite. We made 13 sanidine analyses (three different irradiations) of this bentonite bed that have a weighted-mean age of 74.5 ± 0.1 Ma. From the same bentonite bed, we analyzed six groups of black, fresh-appearing, hexagonal biotite crystals that yielded an age of 75.0 ± 0.1 Ma, about 0.5 Ma older than the coexisting sanidine. We assume that the sanidine age of 74.5 ± 0.1 Ma accurately reflects the age of the bentonite bed because $^{40}Ar/^{39}Ar$ ages of sanidine are generally more consistent and reliable than biotite ages from the same volcanic rock unit. Moreover, our sanidine age of 74.5 Ma is consistent with a sanidine age of a bentonite bed in marine shale in the uppermost part of the *Baculites scotti* zone in New Mexico (J.D. Obradovich, unpub. data).

A minimum age for the Crow Creek Member is based on $^{40}Ar/^{39}Ar$ analyses of three bentonite beds, each about 10 cm thick, in the lower part of the DeGrey Member of central South Dakota. Searight (1937) divided the shale equivalent to the DeGrey in central South Dakota into two units. The lower, composed of brittle fissile shale, he named the "Agency shale," and the upper, composed of bentonitic black shale containing abundant black manganese-rich concretions, he called the "Oacoma zone." Gries (1942) suggested that his Agency shale thickens from south to north in central South Dakota at the expense of his Oacoma zone.

The first bed, which is near the base of the "Agency shale" of the DeGrey Member, is here called the Lower Agency bentonite. Near Rousseau 24 km east of Pierre (Table 2, Fig. 8), the bed is 3.0 m above the top of the Crow Creek Member. The second bed, called the Lower Oacoma bentonite, is exposed in an excavation 4.5 m above the top of the member at Oacoma 5 km west of Chamberlain. This bed may be slightly higher in the stratigraphic section owing to local slumping of the enclosing bentonite-rich shale. The third bed, about 35 m above the Crow Creek Member (Crandell, 1958, p. 14), may be the Lower micaceous bentonite (LMB) of Gries (1947, Table 1, p. 20), and it marks the base of the Oacoma facies at Fort Pierre.

Five $^{40}Ar/^{39}Ar$ analyses of biotite from the Lower Agency bentonite bed have a weighted-mean age of 74.4 ± 0.3 Ma. The ages show considerable scatter, about 1.0 Ma, and therefore the overall age has a large uncertainty. Evidence that this bentonite bed may be slightly contaminated with older minerals is provided by analyses of coexisting plagioclase. Three analyses gave scattered ages of 102.2, 89.1, and 81.8 Ma, much older than the biotite ages from the same sample. The group of analyzed plagioclase crystals may have contained a few detrital grains significantly older than the pyrogenic plagioclase crystals of the bentonite.

Sanidine and biotite analyses for the Lower Oacoma bentonite bed produced identical ages of 73.6 ± 0.1 and 73.6 ± 0.2 Ma (Table 1). The ages of the sanidine and biotite are tightly grouped, consistent with our 74.1 ± 0.1 Ma inferred age of the underlying Crow Creek, and provide a reliable and precise age for this bentonite bed, although its exact stratigraphic position may be somewhat higher owing to slumping as previously suggested.

The third bentonite bed, probably the Lower micaceous bed of Gries (1947), yielded essentially identical sanidine and biotite ages of 73.8 ± 0.1 Ma and 73.8 ± 0.2 Ma (Table 1). The essentially identical ages of the sanidine and biotite (17 and 6 analyses, respectively) establish a reliable and precise age for this bed and the zone of *Didymoceras cheyennense*.

In a previous section of this paper we proposed that the Manson impact event occurred in the zone of *Didymoceras nebrascense* at 74.1 ± 0.1 Ma. To test this proposition we collected two bentonite beds for $^{40}Ar/^{39}Ar$ age analysis from sites in Colorado, one from the zone of *Didymoceras nebrascense* and a second from the zone of *Exiteloceras jenneyi*. Sanidine and biotite from these bentonites were irradiated adjacent to and in the same package with the analyzed Manson sanidine (Table 1).

The first bentonite, exposed along a west service road for highway I-25 (south of Exit 23) north of Trinidad, Colo., is underlain and overlain by shale containing many fragments of *Didymoceras nebrascense*. Ten $^{40}Ar/^{39}Ar$ analyses (two irradiations) of plagioclase from this bentonite bed have a weighted-mean age of 74.1 ± 0.1 Ma (Table 1). The age of this bentonite is statistically indistinguishable from our age for the Manson impact event (74.1 Ma). The second bentonite bed, along the north lane of I-25 (Milepost 126) south of Colorado Springs, Colo., is immediately overlain by concretions that contain an unnamed as yet early form of *Exiteloceras jenneyi*. Along Tom Hollow east of Pueblo, Colorado, rare concretions at the base of the *E. jenneyi* zone contain both this taxon and *Didymoceras stevensoni*. Twenty-one analyses (four irradiations) of sanidine from the bentonite at Milepost 126 have a weighted-mean age of 74.2 ± 0.1 Ma (Table 1), slightly older but statistically the same as our age for the bentonite from the zone of *Didymoceras nebrascense* and the Manson impact event.

In summary, the $^{40}Ar/^{39}Ar$ ages for the bentonite beds below and above the Crow Creek Member are consistent statistically with their stratigraphic position and our 74.1 Ma age for the Manson impact event. The most reliable bentonite ages imply that shale in the zone of

Didymoceras nebrascense was deposited in the span 74.5 ± 0.1 to 73.8 ± 0.1 Ma in the earliest late Campanian. This time interval is surely much less than 0.7 Ma because the dated bentonites beds are 3.0 m and 35 m below and above the Crow Creek, respectively. Additionally, the ages of bentonites in the zones of *Didymoceras nebrascense* and the lower part of the zone of *Exiteloceras jenneyi* from Colorado are consistent with our age for the MIS.

Origin of the Crow Creek Member

Recently two origins for the Crow Creek Member have been advanced. One invokes normal marine processes, whereas the second appeals to an unusual process, largely speculative, involving tsunami waves and subsequent seiche activity initiated by the Manson impact event. Crandell (1952) concluded that the member is of normal marine origin and that it was deposited in response to a local uplift of the Sioux Ridge and the surrounding seafloor of eastern South Dakota. He suggested that this uplift exposed the Codell Sandstone Member of the Carlile Shale of Late Cretaceous age to erosion and that the Codell provided the source of unusually large detrital grains in the calcarenite at the base of the Crow Creek Member. However, Witzke and others (1983) pointed out that no areas are known in eastern South Dakota where the sub-Crow Creek unconformity can be seen cutting across the Niobrara Formation onto the subjacent Codell Sandstone.

Izett and others (1993b) proposed an unconventional idea for the origin of the Crow Creek Member following their determination of a new $^{40}Ar/^{39}Ar$ age for the MIS and discovery of large shocked mineral grains in the member. They briefly speculated that a Manson-impact-triggered tsunami could account, in part, for some of the anomalous stratigraphic and biostratigraphic features of the Crow Creek. However, a genetic relationship between tsunami deposits and asteroidal or cometary impacts is tenuous (Bourgeois, 1994). Izett and Cobban (1994) further noted that the calcarenite at the base of the Crow Creek may be a reworked tsunami deposit. Steiner and Shoemaker (1996) concurred with Izett and others (1993b) that Crow Creek Member was deposited during a Manson-impact-generated tsunami.

Hammond and others (1995) and Witzke and others (1996) criticized the speculation of Izett and others (1993b) that the Crow Creek Member might be a tsunami deposit and offered an alternate proposal for its origin. They suggested, as had Bretz (1979), that the Crow Creek was deposited during a transgression (Bearpaw marine cycle of Gill and Cobban, 1973) of the Late Cretaceous epicontinental sea onto the Sioux Ridge of eastern South Dakota. These authors further suggested that a western-sloping subaerial surface developed mainly on the Gregory and Sharon Springs Members (also locally the Niobrara Formation) of eastern South Dakota.

The marine transgression hypothesis is consistent with some features of the Crow Creek Member; nevertheless, other observations and data seem to us to be inconsistent with the hypothesis as follows:

1) The Crow Creek Member generally lacks sedimentary features such as bioturbated sediments, pervasive small-scale ripple marks, and megafossil lag deposits commonly found in shallow water marine transgressive deposits.

2) Stratigraphic and biostratigraphic data suggest that the disconformity at the base of the Crow Creek is of submarine and not subaerial origin because it separates shale of the Gregory Member from carbonate-rich rocks of the Crow Creek Member both of which contain marine fossils.
3) Biostratigraphic and isotopic ages indicate that the Crow Creek was deposited during a

brief episode in one ammonite zone, *Didymoceras nebrascense.* Such a transgression of the Late Cretaceous sea in such a brief interval is possible but not likely in our view.

4) The marine transgression hypothesis seems to require a coincidence of the Manson impact event and the start of the eastward marine transgression to explain the ubiquitous presence of shocked minerals concentrated in the basal Crow Creek throughout its extent. This coincidence is possible but not probable in our judgment.
5) The marine transgression hypothesis assumes that MIS-derived shocked mineral grains were deposited on a subaerial erosion surface that stretched from 100 km west of Chamberlain to as far east as Yankton. We believe that the shocked mineral grains would be thoroughly dispersed and diluted below their detection limit by the hypothetical eastward transgressing Late Cretaceous sea.
6) According to Witzke and others (1996), the provenance of the reworked assemblage of nannofossils in the Crow Creek Member of southeastern South Dakota was from erosion of "nearby" outcrops of the Niobrara Formation during a marine transgression of that area. Their proposal seems reasonable for easternmost South Dakota because of the proximity of Niobrara Formation source areas. However, their proposal fails to explain the presence of reworked Niobrara nannofossils in the Crow Creek Member of central South Dakota by any mechanism related to an eastward marine transgression for the following reason. The nearest source for the Niobrara nannofossils in the Crow Creek Member of the Chamberlain, South Dakota, area is at least 100 km to the east.

Two possible sedimentologic analogues for the Crow Member have been described. Hay (1960) reported that near Tampico, Mexico, a marked faunal change is coincident with an unconformity separating uppermost Cretaceous shale of the Mendez Formation from lowermost Paleocene shale of the Velasco Formation, both fairly deep-water marine deposits. The Velasco contains a basal spherule bed (now known to be altered tektites) followed by a pair of sandstone beds just above the faunal change. He concluded that the faunal change was best explained by a scenario involving complete withdrawal of the sea (250-500 m depth), creation of a subaerial erosion surface, and immediate transgressive return of the sea. In contrast, Smit and others (1994) proposed that the spherule bed and overlying sandstone beds just above the K-T boundary record the passage of primary tsunami waves and subsequent seiche activity. Keller and others (1994), however, argued that the K-T deposits were not produced by a catastrophic impact-generated tsunami but rather by normal marine processes.

A second possible analogue of the Crow Creek Member was reported by Kastens and Cita (1981). They described a 7-m-thick marl bed of Holocene age in the western part of the Mediterranean Sea composed mainly of calcareous nannofossils. The marl is so uniform that they informally named it "homogenite." Both the Crow Creek and the Holocene "homogenite" have a thin basal clastic bed overlain by a uniform massive marl. Kastens and Cita (1981) maintained that the deposits were formed by refracted tsunami waves spawned by the collapse of the Santorini caldera.

In conclusion, two origins for the Crow Creek Member have been proposed, normal marine transgression and impact-triggered tsunami. Both hypotheses have flaws, and neither hypothesis is consistent with all of the present data, nevertheless we judge that the majority of the evidence at this time favors an impact spawned tsunami origin for the Crow Creek Member. Our model assumes that at the instant of the Manson impact the Western Interior Late Cretaceous sea covered much of South Dakota, perhaps as far as central Iowa. We speculate that impact triggered tsunami waves and subsequent seiche activity mobilized Late Cretaceous seafloor sediments of eastern South Dakota and scoured Niobrara Formation and Pierre Shale sediments along the eastern margin of the Cretaceous seaway. These entrained sediments were 1) mixed with MIS impact ejecta, indigenous microfossils, and upper Niobrara Formation nannofossil, 2) transported by tsunami and seiche waves, and 3) deposited as the Crow Creek Member throughout eastern South Dakota.

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Appendix A. Analytical methods.

$^{40}Ar/^{39}Ar$ ages were determined in the USGS laboratory in Menlo Park, CA, using procedures described by Dalrymple and Lanphere (1971), Dalrymple and Duffield (1988), and Dalrymple (1989). A technical feature of this dating method is that materials of unknown age are irradiated with fast neutrons next to a fleuence-monitor mineral, or standard, having an accepted isotopic age. Thus, the method is a relative one; ages of unknown materials are relative to ages of a selected fluence-monitor mineral. The measured $^{40}Ar/^{39}Ar$ ratios of the fluence-monitor mineral are used with its known age to calculate a conversion-efficiency factor, J, which is a measure of the fraction of ^{39}K converted to ^{39}Ar by the fast neutron reaction ($^{39}K(n,p)^{39}Ar$). The factor J is then used in the age equation to calculate ages for materials of unknown age. Obviously, the precision of the fluence-monitor mineral calibration has a significant effect on the precision of the ages calculated for materials of unknown age.

$^{40}Ar/^{39}Ar$ analyses of the sanidine from the melt rock of the Manson impact structure were made using single fragments (~0.3 mg). For sanidine and biotite from the bentonite beds, small groups (3-20) crystals were analyzed, and for the plagioclase, several milligrams were used. The analyzed sanidine and plagioclase crystals of the bentonite beds were 100-120 mesh, but biotite crystals were 60-80 mesh. To prepare the bentonite beds for analysis we collected 10-25 kg because the clay generally contains only a minuscule amount of volcanic crystals. To recover the sanidine and plagioclase from the sticky, swelling clay, we used the following scheme: Samples were dried at 125 °C, disaggregated in buckets of water, and further disaggregated using a cocktail blender. The resulting slurry was washed and sieved, and the +120 mesh fraction recovered. This fraction was ultrasonically scrubbed in HCl to remove gypsum, and next etched in 10% HF for 3 minutes. Concentrates of feldspar were obtained by conventional heavy-liquid separation techniques. A few milligrams of these minerals were hand picked under a microscope at 90X, and the purity of the concentrate checked repeatedly in refractive index oils.

We emphasize that not all bentonite beds are amenable to $^{40}Ar/^{39}Ar$ analysis. Some are contaminated with older detrital minerals and others do not contain enough pyrogenic crystals of sufficient size. For example, plagioclase and biotite from a 25-cm-thick bentonite bed at Fort

Pierre, probably the Big bentonite bed of Gries (1947), yielded seven inconsistent older plagioclase ages in the range 76.53 to 74.23 Ma and three inconsistent biotite ages in the range 72.52 to 74.00 Ma

In this study minerals were irradiated in 1993-96 in the core of the USGS's TRIGA reactor in seven different experiments (irradiation packages GLN9, JDO13, GLN12, GLN13, JDO21, JDO22, and JDO24; Table 1). Six of these were irradiated for 30 hours and received fast neutron doses of 3.0×10^{18} nvt; the sixth one (GLN9) was irradiated for 24 hours and received a dose of 2.4×10^{18} nvt. The nuclear reactor fluence attributes, irradiation procedures, and methods for measuring corrections for interfering Ar isotopes induced by undesirable nuclear reactions with K and Ca were described by Dalrymple and others (1981) and Dalrymple (1989). To achieve added calibration, samples of Manson M-1 sanidine were placed next to other packets containing minerals from bentonite beds near the Crow Creek Member of the Pierre Shale to achieve added relative calibration. Figure A-1 shows the data and J-curve for JDO22 in the series of seven irradiations.

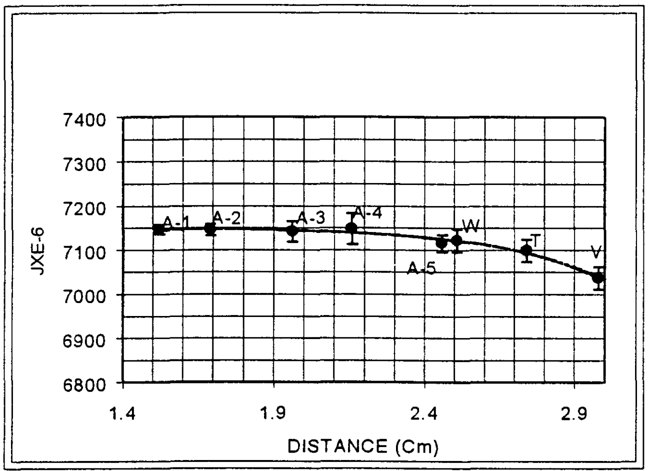


Figure A-1. Diagram for irradiation experiment JDO22 showing plotted position (mm) of reactor package containing sanidine monitor-mineral Taylor Creek Rhyolite (27.92 Ma) and associated J Values. Error bars (2 sigma) are calculated precision of J.

A typical irradiation packet consisted of selected minerals loaded into a 9-mm-diameter aluminum-foil cup and covered with a 9-mm aluminum foil cap. The flattened pancake-like packets were sandwiched between similar packets of neutron fluence-monitors, arranged in a vertical stack in a 10-mm-diameter quartz glass tube, and the position of the packets measured. The distance between adjacent packet centers typically was 0.2-0.3 mm. The neutron fluence within the radiation package was measured by analyzing 5-7 lots of 2-3 sanidine crystals for each fluence monitor.

We used sanidine from the Taylor Creek Rhyolite (TCR) of New Mexico (Dalrymple and Duffield, 1988) as a fluence-monitor mineral because it has been shown to be uniform in K and Ar content, and its isotopic age (27.92 Ma) is within an acceptable range of the suspected ages of minerals to be dated. This age for TCR is based on an internally consistent set of laboratory standard values measured in the Menlo Park laboratory, and it results in a measured age of 513.9 Ma for the widely used Mmhb-1 hornblende (Lanphere and others, 1990), which is 1.26% younger than the internationally adopted mean value of 520.4 Ma (Samson and Alexander, 1987). Dalrymple and others (1993) gave reasons for using 27.92 Ma for TCR sanidine, its effect on interlaboratory comparisons, and an exact algebraic conversion formula. New ages reported in this paper can be normalized approximately to an age of 520.4 Ma for Mmhb-1 hornblende by multiplying the ages by 1.0143.

Two methods were used to date the samples: total fusion using an argon-ion laser to melt individual or small groups of crystals and incremental heating using an internal resistance furnace. For laser total-fusion analysis, which results in a single $^{40}Ar/^{39}Ar$ age, the minerals were loaded in wells in a copper disk. Zero-age degassed basalt glass chips were used to facilitate melting of sanidine and plagioclase. The samples were heated and melted in ultrahigh vacuum for as long as 4 minutes with a 5-W continuous Ar-ion laser at the maximum temperature attainable, ~1,500 °C. The gas released from the samples during melting was cleaned with Zr-Al getters, and the isotopic composition of the Ar released was analyzed with a Mass Analyzer Products 216 rare-gas mass spectrometer with Baur-Signer source and 20-stage electron multiplier (Dalrymple and Duffield, 1988; Dalrymple, 1989).

For incremental heating analysis, which results in a series of ages as a function of increasing temperature (an age spectrum), individual fragments (~1 mg) of M-1 sanidine or melt rock were heated in a double pumped resistance furnace. The gas released during each heating step was cleaned and analyzed with the same cleanup system and mass spectrometer described above.

Errors given for individual $^{40}Ar/^{39}Ar$ ages are estimates of the analytical precision at the 1 σ level and include a conservative error of 0.5% in J. Mean values for a group of analyses are weighted means $\pm \sigma_{best}$, where weighting is by the inverse of the variance (Taylor, 1982). The standard deviation of the means for groups of ages are at the 95% confidence level. All ages reported herein were calculated using decay constants recommended by the Subcommission on Geochronology of the IUGS (Steiger and Jäger, 1977).

Appendix B-1. Laser total-fusion $^{40}Ar/^{39}Ar$ analytical data for sanidine clasts (melted microcline) from the 374.8 and 470.0 m depths of the Manson M-1 core. Ages are relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based an age of 513.4 Ma for Mmhb-1 hornblende. Age in parentheses relative to a fluence-monitor age for Mmhb-1 hornblende of 520.4 Ma. Analysts G.A. Izett and G.B. Dalrymple. *, radiogenic.

Exp. No.	Irrad. No.	Depth ft	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar*/^{39}Ar$	^{40}A %	Age Ma	Error 1 sigma
93Z0767	GLN9-5	374.8	6.21E-02	4.208E-02	2.805E-02	6.7605	89.	74.17	0.50
93Z0808	JDO13-9	470.0	7.56E-02	1.098E-01	3.922E-02	5.5144	82.	73.66	0.47
93Z0873B	JDO13-9	470.0	7.56E-02	9.326E-02	3.389E-02	5.5549	84.	74.18	0.45
93Z0873C	JDO13-9	470.0	7.56E-02	1.297E-01	1.790E-02	5.4856	91.	73.39	0.49
93Z0873D	JDO13-9	470.0	7.56E-02	4.571E-02	1.185E-02	5.6014	94.	74.79	0.56
93Z0873E	JDO13-9	470.0	7.56E-02	1.360E-01	8.124E-03	5.5213	95.	73.74	0.46
93Z0873F	JDO13-9	470.0	7.56E-02	1.313E-01	1.041E-02	5.5607	94.	74.26	0.45
93Z0873H	JDO13-9	470.0	7.56E-02	7.635E-02	7.561E-03	5.5930	96.	74.68	0.48
93Z0873I	JDO13-9	470.0	7.56E-02	3.883E-02	2.378E-02	5.5756	88.	74.45	0.49
93Z0873J	JDO13-9	470.0	7.56E-02	5.208E-02	6.269E-02	5.5259	74.	73.80	0.56
94Z0212	GLN12-3	374.8	6.92E-02	2.261E-02	2.276E-02	5.9814	89.	73.19	0.83
94Z0213	GLN12-3	374.8	6.92E-02	3.616E-02	9.036E-03	6.0615	95.	74.19	0.46
94Z0214	GLN12-3	374.8	6.92E-02	2.459E-02	8.805E-03	6.0938	95.	74.58	0.66
94Z0215	GLN12-3	374.8	6.92E-02	4.592E-02	9.161E-03	6.0911	95.	74.54	0.87
94Z0216	GLN12-3	374.8	6.92E-02	3.066E-02	2.033E-02	6.0766	90.	74.36	0.61
95Z0351	GLN13-6	374.8	7.18E-02	3.920E-02	7.278E-03	5.8746	96.	74.49	0.52
95Z0352	GLN13-6	374.8	7.18E-02	4.698E-02	8.558E-03	5.8272	95.	73.90	0.46
95Z0353	GLN13-6	374.8	7.18E-02	4.035E-02	1.529E-02	5.8216	92.	73.83	0.48
95Z0354	GLN13-6	374.8	7.18E-02	5.617E-02	1.232E-02	5.8411	94.	74.07	0.44
95Z0355	GLN13-6	374.8	7.18E-02	3.403E-02	1.003E-02	5.8498	95.	74.18	0.47
95Z0356	GLN13-6	374.8	7.18E-02	4.088E-02	8.328E-03	5.8824	95.	74.59	0.48
95Z0624	JDO21-X	374.8	7.11E-02	4.578E-02	2.163E-02	5.8339	90.	73.33	0.53
95Z0625	JDO21-X	374.8	7.11E-02	4.177E-02	2.564E-02	5.9356	88.	74.58	0.59
95Z0626	JDO21-X	374.8	7.11E-02	3.622E-02	2.675E-02	5.9083	88.	74.24	0.47
95Z0627	JDO21-X	374.8	7.11E-02	4.105E-02	2.483E-02	5.9368	88.	74.59	0.48
95Z0628	JDO21-X	374.8	7.11E-02	4.278E-02	2.570E-02	5.8864	88.	73.97	0.47
95Z0629	JDO21-X	374.8	7.11E-02	2.974E-02	3.508E-02	5.9042	85.	74.19	0.45
96Z0176	JDO22-Y	374.8	7.08E-02	6.139E-02	1.839E-02	5.9319	91.	74.16	0.65
96Z0177	JDO22-Y	374.8	7.08E-02	4.600E-02	2.290E-02	5.9560	89.	74.46	0.51
96Z0178	JDO22-Y	374.8	7.08E-02	3.750E-02	1.310E-02	5.9342	93.	74.19	0.66
96Z0179	JDO22-Y	374.8	7.08E-02	4.520E-02	1.910E-02	5.9072	91.	73.86	0.55
96Z0180	JDO22-Y	374.8	7.08E-02	4.680E-02	1.400E-02	5.9088	93.	73.88	0.52
96Z0203	JDO22-Y	374.8	7.08E-02	1.440E-01	1.720E-02	5.8936	91.	73.70	0.48
96Z0204	JDO22-Y	374.8	7.08E-02	4.303E-02	1.880E-02	5.9003	91.	73.78	0.53
					Weighted mean			**74.11**	**(75.17)**
					Sigma Best			**0.09**	

Appendix B-2. Laser total-fusion $^{40}Ar/^{39}Ar$ analytical data for sanidine and biotite from a bentonite bed in the upper part of the Gregory Member (Zone of *Baculites scotti*) at Oacoma, South Dakota. Ages are relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based an age of 513.4 Ma for Mmhb-1 hornblende. Ages in parentheses relative to a fluence-monitor age for Mmhb-1 hornblende of 520.4 Ma. Analyst G.A. Izett. *, radiogenic.

Exp. No.	Irrad. No	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar^*/^{39}Ar$	$^{40}Ar^*$ %	Age Ma	Error 1 sigma
Sanidine								
94Z0266	GLN12-2	6.92E-02	7.876E-02	7.530E-03	6.0992	96.4	74.54	0.49
94Z0267	GLN12-2	6.92E-02	7.517E-02	3.214E-04	6.0833	99.7	74.34	0.53
94Z0269	GLN12-2	6.92E-02	7.930E-02	8.599E-04	6.0678	99.4	74.16	0.51
94Z0270	GLN12-2	6.92E-02	8.119E-02	1.459E-06	6.1071	99.9	74.63	0.42
94Z0271	GLN12-2	6.92E-02	8.437E-02	3.568E-04	6.0969	99.7	74.51	0.47
94Z0247	GLN12-11	6.95E-02	1.031E-01	6.194E-04	6.0480	99.6	74.28	0.44
94Z0248	GLN12-11	6.95E-02	8.913E-02	5.634E-04	6.0394	99.6	74.17	0.44
94Z0249	GLN12-11	6.95E-02	9.070E-02	1.880E-05	6.1016	99.9	74.92	0.42
94Z0250	GLN12-11	6.95E-02	8.965E-02	6.814E-04	6.0712	99.5	74.56	0.46
95Z0321	GLN13-2	7.15E-02	6.733E-02	1.018E-03	5.9093	99.3	74.61	0.50
95Z0322	GLN13-2	7.15E-02	8.961E-02	2.725E-03	5.8988	98.5	74.48	0.50
95Z0323	GLN13-2	7.15E-02	7.355E-02	1.110E-03	5.9035	99.3	74.54	0.43
95Z0324	GLN13-2	7.15E-02	8.676E-02	7.508E-04	5.9025	99.4	74.53	0.45
				Weighted mean			**74.49**	**(75.55)**
				Sigma best			**0.13**	
Biotite								
96Z0538	JDO24-O	6.70E-02	1.340E-01	4.070E-03	6.3298	98.0	74.90	0.51
96Z0539	JDO24-O	6.70E-02	6.540E-02	4.730E-03	6.3462	97.7	75.10	0.50
96Z0561	JDO24-O	6.70E-02	9.150E-02	5.090E-03	6.3601	97.6	75.30	0.55
96Z0562	JDO24-O	6.70E-02	1.013E-01	7.750E-03	6.3346	96.4	75.00	0.56
96Z0563	JDO24-O	6.70E-02	1.890E-01	5.730E-03	6.3279	97.3	74.90	0.45
96Z0564	JDO24-O	6.70E-02	1.320E-01	4.550E-03	6.3372	97.8	75.00	0.62
				Weighted mean			**75.02**	**(76.10)**
				Sigma best			**0.06**	

Appendix B-3. Laser total-fusion $^{40}Ar/^{39}Ar$ analytical data for sanidine and biotite from the Lower Oacoma bentonite bed (Zone of *Didymoceras nebrascense*) in the lower part of the DeGrey Member at Oacoma, South Dakota, and biotite from the Lower Agency bentonite bed in the lower part of the DeGrey Member near Rousseau, South Dakota. Ages are relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based an age of 513.4 Ma for Mmhb-1 hornblende. Ages in parentheses relative to a fluence-monitor age for Mmhb-1 hornblende of 520.4 Ma. Analyst G.A. Izett. *, radiogenic.

Exp. No.	Irrad. No.	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar^*/^{39}Ar$	$^{40}Ar^*$ %	Age Ma	Error 1 sigma
Sanidine and biotite from the Lower Oacoma bentonite bed								
96Z0154	JDO22-X	7.12E-02	9.329E-02	8.577E-04	5.8573	99.4	73.72	0.56
96Z0155	JDO22-X	7.12E-02	2.560E-01	1.278E-03	5.8364	99.2	73.46	0.48
96Z0156	JDO22-X	7.12E-02	1.502E-01	7.766E-04	5.8568	99.5	73.71	0.43
96Z0157	JDO22-X	7.12E-02	1.659E-01	5.257E-04	5.8638	99.6	73.80	0.45
96Z0158	JDO22-X	7.12E-02	8.361E-02	1.205E-03	5.8412	99.3	73.52	0.47
96Z0159	JDO22-X	7.12E-02	4.081E-01	7.602E-04	5.8630	99.5	73.79	0.46
96Z0160	JDO22-X	7.12E-02	1.996E-01	2.145E-04	5.8642	99.8	73.80	0.45
96Z0161	JDO22-X	7.12E-02	4.493E-01	1.753E-04	5.8696	99.8	73.87	0.45
96Z0162	JDO22-X	7.12E-02	2.073E-01	2.816E-03	5.8363	98.5	73.46	0.48
96Z0170	JDO22-X	7.12E-02	8.104E-02	2.189E-03	5.8376	98.8	73.47	0.44
96Z0171	JDO22-X	7.12E-02	8.721E-02	1.008E-03	5.8615	99.4	73.77	0.50
96Z0172	JDO22-X	7.12E-02	8.807E-02	2.737E-03	5.8533	98.5	73.67	0.46
96Z0173	JDO22-X	7.12E-02	3.724E-01	1.195E-03	5.8368	99.3	73.47	0.47
96Z0174	JDO22-X	7.12E-02	8.290E-02	2.110E-03	5.8428	98.8	73.54	0.45
96Z0175	JDO22-X	7.12E-02	1.153E-01	4.481E-03	5.8445	97.7	73.56	0.48
	-			Weighted mean			**73.64**	**(74.69)**
				Sigma best			**0.12**	
Biotite from the Lower Oacoma bentonite bed								
96Z0181	JDO22-S	7.10E-02	1.025E-01	4.158E-03	5.8685	97.8	73.62	0.45
96Z0182	JDO22-S	7.10E-02	1.151E-01	2.928E-02	5.8823	87.1	73.81	0.51
96Z0183	JDO22-S	7.10E-02	1.055E-01	4.632E-03	5.8781	97.6	73.76	0.48
96Z0184	JDO22-S	7.10E-02	1.908E-01	4.783E-03	5.8358	97.5	73.24	0.45
96Z0185	JDO22-S	7.10E-02	1.435E-01	5.826E-03	5.8811	97.0	73.80	0.44
				Weighted mean			**73.64**	**(74.69)**
				Sigma best			**0.12**	
Biotite from the Lower Agency bentonite bed								
96Z0519	JDO24-L	6.70E-02	3.370E-02	4.080E-03	6.2776	98.0	74.32	0.81
96Z0520	JDO24-L	6.70E-02	2.570E-01	2.170E-03	6.3273	98.9	74.90	0.50
96Z0521	JDO24-L	6.70E-02	4.850E-02	5.880E-03	6.2436	97.2	73.93	0.59
96Z0527	JDO24-L	6.70E-02	1.240E-01	4.960E-03	6.2557	97.6	74.07	0.58
96Z0529	JDO24-L	6.70E-02	2.130E-01	4.260E-03	6.2911	97.9	74.48	0.61
				Weighted mean			**74.38**	**(75.44)**
				Sigma best			**0.27**	

Appendix B-4. Laser total-fusion $^{40}Ar/^{39}Ar$ analytical data for sanidine and biotite from the Lower micaceous bentonite bed (Zone of *Didymoceras cheyennense*) in the upper part of the DeGrey Member at Fort Pierre, South Dakota. Ages are relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based an age of 513.4 Ma for Mmhb-1 hornblende. Ages in parentheses relative to a fluence-monitor age for Mmhb-1 hornblende of 520.4 Ma. Analyst G.A. Izett. *, radiogenic.

Exp. No.	Irrad. No.	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar^{*}/^{39}Ar$	$^{40}Ar^{*}$ %	Age Ma	Error 1 sigma
Sanidine								
95Z0598	JDO21-Y	7.12E-02	9.238E-01	1.725E-03	5.8604	99.1	73.77	0.44
95Z0599	JDO21-Y	7.12E-02	8.097E-01	1.450E-03	5.8586	99.2	73.74	0.45
95Z0600	JDO21-Y	7.12E-02	7.334E-01	1.049E-03	5.8916	99.4	74.15	0.61
95Z0601	JDO21-Y	7.12E-02	2.225E-00	4.608E-03	5.8496	97.9	73.63	0.54
95Z0602	JDO21-Y	7.12E-02	8.179E-01	1.868E-03	5.8799	99.0	74.01	0.45
95Z0603	JDO21-Y	7.12E-02	9.956E-01	3.497E-03	5.8801	98.2	74.01	0.51
95Z0604	JDO21-S	7.11E-02	8.723E-01	2.988E-03	5.8885	98.5	73.97	0.44
95Z0605	JDO21-S	7.11E-02	7.950E-01	2.433E-03	5.8573	98.7	73.59	0.44
95Z0606	JDO21-S	7.11E-02	1.396E-00	1.358E-03	5.8740	99.3	73.79	0.43
95Z0607	JDO21-S	7.11E-02	3.952E-01	9.967E-08	5.9012	99.9	74.13	0.42
95Z0608	JDO21-S	7.11E-02	2.173E-01	4.023E-03	5.8225	97.9	73.16	0.58
95Z0609	JDO21-S	7.11E-02	1.059E-00	1.375E-07	5.9151	99.9	74.30	0.42
95Z0618	JDO21-S	7.10E-02	8.910E-01	3.010E-03	5.8321	98.5	73.20	0.45
95Z0619	JDO21-S	7.10E-02	3.920E-01	7.870E-04	5.8866	99.5	73.87	0.46
95Z0620	JDO21-S	7.10E-02	1.737E-00	3.660E-03	5.8594	98.3	73.54	0.45
95Z0621	JDO21-S	7.10E-02	8.160E-01	3.390E-03	5.8720	98.3	73.69	0.48
95Z0622	JDO21-S	7.10E-02	1.359E-00	9.360E-04	5.8816	99.6	73.81	0.43
				Weighted mean			**73.80**	**(74.85)**
				Sigma best			**0.11**	
Biotite								
95Z0652	JDO21-R	7.10E-02	1.957E-01	1.268E-02	5.8713	93.9	73.64	0.51
95Z0653	JDO21-R	7.10E-02	6.024E-02	9.524E-03	5.8429	95.3	73.29	0.66
95Z0654	JDO21-R	7.10E-02	9.741E-02	1.297E-02	5.8934	93.8	73.91	0.57
95Z0655	JDO21-R	7.10E-02	1.528E-01	1.186E-02	5.8848	94.3	73.80	0.46
95Z0656	JDO21-R	7.10E-02	1.243E-01	8.019E-03	5.9155	96.0	74.18	0.44
95Z0657	JDO21-R	7.10E-02	1.392E-01	9.979E-03	5.8706	95.1	73.63	0.46
	-			Weighted mean			**73.79**	**(74.84)**
				Sigma best			**0.20**	

Appendix B-5. Laser total-fusion $^{40}Ar/^{39}Ar$ analytical data for sanidine from a bentonite bed in the Zone of *Didymoceras nebrascense* north of Trinidad, Colo. (Lat 37° 17' 23" N., Long 104° 31' 24"), and sanidine from another bentonite bed in the lower part of the Zone of *Exiteloceras jenneyi* (Pierre Shale) south of Colorado Springs, Colo. (Lat 38° 38' 47" N., Long 104° 41' 41"). Ages are relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based an age of 513.4 Ma for Mmhb-1 hornblende. Ages in parentheses relative to a fluence-monitor age for Mmhb-1 hornblende of 520.4 Ma. Analyst G.A. Izett. *, radiogenic.

Sanidine from a bentonite bed in the *Exiteloceras jenneyi* zone in Colorado

Exp. No.	Irrad. No.	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar^{*}/^{39}Ar$	$^{40}Ar^{*}$ %	Age Ma	Error 1 sigma
94Z0217	GLN12-4	6.94E-02	5.944E-02	1.653E-06	6.0593	99.9	74.27	0.42
94Z0218	GLN12-4	6.94E-02	6.550E-02	1.930E-06	6.0758	99.9	74.47	0.21
94Z0219	GLN12-4	6.94E-02	6.520E-02	1.900E-06	6.0929	100.0	74.56	0.50
94Z0245	GLN12-4	6.94E-02	8.063E-02	6.460E-04	6.0460	99.5	74.11	0.47
94Z0246	GLN12-4	6.94E-02	7.625E-02	4.084E-04	6.0290	99.7	73.90	0.45
94Z0272	GLN12-12	6.96E-02	5.874E-02	3.066E-06	6.0427	99.9	74.30	0.42
94Z0273	GLN12-12	6.96E-02	1.739E-01	5.907E-04	6.0189	99.6	74.01	0.44
94Z0274	GLN12-12	6.96E-02	1.777E-01	9.889E-04	6.0122	99.4	73.93	0.45
94Z0275	GLN12-12	6.96E-02	7.176E-02	1.017E-03	6.0115	99.4	73.92	0.45
94Z0276	GLN12-12	6.96E-02	8.128E-02	7.207E-04	6.0040	99.5	73.83	0.44
95Z0329	GLN13-8	7.18E-02	8.368E-02	1.019E-03	5.8562	99.3	74.30	0.48
95Z0330	GLN13-8	7.18E-02	6.690E-02	3.999E-03	5.8774	97.8	74.57	0.49
95Z0331	GLN13-8	7.18E-02	7.971E-02	3.030E-03	5.8516	98.3	74.24	0.44
95Z0332	GLN13-8	7.18E-02	2.378E-01	4.860E-04	5.8632	99.6	74.39	0.46
95Z0333	GLN13-8	7.18E-02	2.394E-02	7.040E-03	5.7461	96.3	72.93	0.52
95Z0624	JDO21-X	7.11E-02	4.580E-02	2.160E-02	5.8339	90.0	73.33	0.53
95Z0625	JDO21-X	7.11E-02	4.180E-02	2.560E-02	5.9356	88.6	74.58	0.59
95Z0626	JDO21-X	7.11E-02	3.620E-02	2.680E-02	5.9083	88.1	74.24	0.47
95Z0627	JDO21-X	7.11E-02	4.110E-02	2.480E-02	5.9368	88.9	74.59	0.48
95Z0628	JDO21-X	7.11E-02	4.280E-02	2.570E-02	5.8864	88.5	73.97	0.47
95Z0629	JDO21-X	7.11E-02	2.974E-02	3.510E-02	5.9042	85.0	74.19	0.45
				Weighted mean			**74.18**	**(75.24)**
				Sigma best			**0.09**	

Plagioclase from a bentonite bed in the *Didymoceras nebrascense* zone in Colorado

Exp. No.	Irrad. No.	J	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	$^{40}Ar^{*}/^{39}Ar$	$^{40}Ar^{*}$ %	Age Ma	Error 1 sigma
94Z0277	GLN12-14	6.95E-02	8.972E-00	1.115E-02	6.0606	95.8	74.43	0.45
94Z0278	GLN12-14	6.95E-02	9.840E-00	7.326E-03	6.0319	97.6	74.08	0.43
94Z0279	GLN12-14	6.95E-02	9.849E-00	8.868E-03	6.0465	96.9	74.26	0.44
94Z0280	GLN12-14	6.95E-02	1.074E+01	7.546E-03	6.0238	97.6	73.99	0.45
94Z0281	GLN12-14	6.95E-02	8.967E-00	9.977E-03	6.0183	96.3	73.92	0.45
95Z0336	GLN13-3	7.16E-02	1.012E+01	1.076E-02	5.8636	95.9	74.19	0.47
95Z0337	GLN13-3	7.16E-02	9.962E-00	1.050E-02	5.8577	96.0	74.12	0.46
95Z0339	GLN13-3	7.16E-02	9.609E-00	1.124E-02	5.8570	95.6	74.11	0.47
95Z0340	GLN13-3	7.16E-02	9.133E-00	2.919E-02	5.8553	87.9	74.09	0.50
95Z0341	GLN13-3	7.16E-02	9.547E-00	9.911E-03	5.8589	96.2	74.13	0.46
				Weighted mean			**74.13**	**(75.19)**
				Sigma best			**0.14**	

Appendix C. Ammonite collections from the Pierre Shale in South Dakota, Nebraska, and Wyoming. Identifications by W.A. Cobban.

Loc. No.	Location in South Dakota except where noted	Collectors and years	Remarks
22181	NW 1/4 Sec. 16, T. 30 N., R. 6 W., Knox Co., Neb.	W.A. Cobban, 1941	*Jeletzkytes nebrascensis* from concretions in silty shale in Elk Butte Member
D13495	SE 1/4 Sec. 9, T. 3 S., R. 24E., Jackson Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D13496	Southline Sec. 18, T. 2 S., R. 26 E., Jones Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D13497	NE 1/4 Sec. 31, T. 45 N., R. 27 E., Jones Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D13498	NW 1/4 Sec. 32, T. 2 S., R. 27 E., Jones Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D13598	C E 1/2 W 1/2 Sec. 20, T. 44 N., R. 29 E., Jones Co.	G.A. Izett, 1994	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D13597	SW 1/4 Sec. 23, T. 42 N., R. 29 W., Mellette Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites clinolobatus* from white limestone concretions in Mobridge Member
D4971	NE 1/4 Sec. 21, T. 6 N., R. 29 E., Stanley Co.	W.A. Cobban, 1965	*Baculites grandis* from zone of phosphatic nodules at the base of the Mobridge Member
D877	NW 1/4 Sec. 11, T. 1 N., R. 24 E., Haakon Co.	W.A. Cobban, 1956; G.A. Izett and W.A. Cobban, 1996	*Baculites baculus* and *Inoceramus typicus* in Virgin Creek Member
Black Hills Institute	SE 1/4 Sec. 13, T. 105 N., R. 77 W., Lyman Co.	L. Marken, Reliance S. Dak.	*Baculites baculus* in shale of Virgin Creek Member
D13500	NW 1/4 Sec. 32, T. 3 S., R. 29 E., Jones Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites eliasi* in Virgin Creek Member
D13599	SW 1/4 Sec. 5, T. 4 S., R. 29 E., Jones Co.	G.A. Izett, 1994	*Baculites eliasi* in Virgin Creek Member
D13499	SW 1/ 4 Sec. 24, T. 44 N., R. 9 E., Mellette Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites eliasi* in Virgin Creek Member
D13501	NW 1/ sec. 33, T. 3 S., R. 29 E., Jones Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites eliasi* in Virgin Creek Member
D13502	SW 1/4 Sec. 17, T. 43 N., R. 28 W., Mellette Co.	G.A. Izett, 1994	*Baculites eliasi* in Virgin Creek Member
D13578	NE 1/4 Sec. 12, T. 32 N., R. 7 W., Knox Co., Neb.	G.A. Izett and W.A. Cobban, 1995	*Baculites eliasi* in Virgin Creek Member
D13789	NW 1/4 Sec. 5 T. 3 n., r. 29 E., Stanley Co.	G.A. Izett, 1997	*Baculites reesidei* in shale of the upper Verendrye Memberi
D13782	SW 1/4 Sec. 10, T. 56 N., R. 69 W., Campbell Co. Wyo.	G.A. Izett and W.A. Cobban, 1997	*Baculites reesidei* in gray limestone concretions in Pierre Shale
D4965	NE 1/4 Sec. 13, T. 102 N., R. 73 W., Lyman Co.	W.A. Cobban, 1965	*Baculites reesidei* and *Jeletzkytes brevis* from middle part of Verendrye Member
D882	NE 1/4 Sec. 33, T. 5 N., R. 31 E., Stanley Co.	W.A. Cobban	*Baculites cuneatus* from lower part of the Verendrye Member
D844	Sec. 2, T. 110 N., R. 79 W., Hughes Co.	W.A. Cobban	*Baculites cuneatus* from lower part of Verendrye Member
D13506	SW 1/4 Sec. 33, T. 104 N., R. 74 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites compressus* from black manganese-rich concretions near top of DeGrey Member (Oacoma facies)

Appendix C (contd.)

Loc. No.	Location in South Dakota except where noted	Collectors and years	Remarks
D13507	NE 1/4 Sec. 17, T.107 N., R. 71 W., Buffalo Co.	G.A. Izett, 1996-97	*Baculites compressus, Baculites* sp. (Gulf migrant), and *Jeletzkytes brevis* from black manganese-rich concretions in upper 2-3 m of DeGrey Member (Oacoma Facies)
D13790	SE 1/4 SE 1/4 Sec. 9, T. 107, R. 71 W., Buffalo Co.	G.A. Izett and W.A. Cobban, 1997	*Baculites compressus, Baculites* sp. (Gulf migrant*), Placenticeras* sp., *and Eutrephocera* sp., *and Jeletskytes brevis*
D13791	SW 1/4 NE 1/4 Sec. 36, T. 110, R. 76 W., Hughes Co.	G.A. Izett and W.A. Cobban, 1997	*Didymoceras?* sp., *Baculites* sp., *Placenticeras* sp., *and Jeletskytes brevis*
D13507	SE 1/4 Sec. 15, T.106 N., R. 71 W., Buffalo Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites compressus* from black concretions at top of DeGrey Member (Oacoma facies)
D13508	C S 1/2 Sec. 13, T .105 N., R. 71 W., Brule Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites compressus* from black concretions at top of DeGrey Member (Oacoma facies)
D13579	NW 1/4 Sec. 8, T. 103 N., R. 73 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites sp.* and *Scaphites sp.* from black concretions in upper part of DeGrey Member (Oacoma facies)
D4963	C E 1/2 E 1/2 Sec. 15, T. 104 N., R. 72 W., Lyman Co.	W.A. Cobban, 1965	*Didymoceras* sp. *and Baculites* sp. (stout elliptical cross section and widely spaced arcuate flank ribs) from black concretions in the upper 2-3 m of the DeGrey Member (Oacoma facies)
D13641	SW 1/4 Sec. 6, T. 109 N., R. 75 W., Hughes Co.	W.A. Cobban, 1995	*Didymoceras cheyennense* from black concretions in the upper part of the DeGrey Member (Oacoma facies)
61S50	Center Sec. 9, T. 104 N., R. 72 W., Lyman Co.	R.E. Stevenson, 1961?	*Didymoceras cheyennense* from DeGrey Member (Oacoma facies) at the Manganese Pilot Plant north of I-90 west of Oacoma
D13784	N 1/2 Sec. 21, T. 9 N., R. 8 E., Butte Co.	G.A. Izett and W.A. Cobban, 1997	*Exiteloceras jenneyi* and *Baculites* sp. from buff-gray concretion along Mud Elm Creek
D13785	W 1/2 Sec. 9, T. 12 N., R. 3 E., Butte Co.	G.A. Izett and W.A. Cobban, 1997	*Exiteloceras jenneyi*
D13786	Section line between 19 and 20, T. 9 N., R.83 E., Butte Co.	G.A. Izett and W.A. Cobban, 1997	*Exiteloceras jenneyi, Baculites rugosus,* and *Placenticeras* sp.
D13440 D13441 D13577 D13583	E 1/2 Sec. 34 and W 1/2 Sec. 35, T. 107 N., R. 72 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1993-94	*Baculites rugosus* (late form) from black concretions in upper part of DeGrey Member (Oacoma facies)
D13584	N 1/2 Sec. 9, T. 104 N., R. 72 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1994	*Baculites* n. sp. (Gulf migrant) from black concretions in upper part of the DeGrey Member (Oacoma facies)
D13692	SE 1/4 Sec. 11, T. 107 N., R. 74 W., Lyman Co.	G.A. Izett, 1995; G.A. Izett and W.A. Cobban 1997	*Baculites compressus, Baculites* n. sp. (Gulf migrant) from black concretions in upper part of DeGrey Member (Oacoma facies) at the type locality of the DeGrey Member
D13686	NW 1/4 Sec. 19, T. 108 N., R. 72 W., Buffalo Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites* n. sp. (Gulf migrant), *Scaphites* sp., and pachydiscid ammonite from DeGrey Member (Oacoma facies) concretions
D13443	SW 1/4 Sec. 25, T. 103N., R. 73 W., Black Dog Township, Lyman Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites* n. sp. (Gulf migrant) from manganese-rich concretions in DeGrey Member (Oacoma facies)
D13748	SE 1/4 Sec. 30, T. 140 N., R. 58 W., Barnes Co., North Dakota	G.A. Izett, 1996	*Baculites* n. sp. (Gulf migrant) from black concretions in lower part of DeGrey Member (Oacoma facies) 3 m above top of Gregory Mbr.

Appendix C (contd.)

Loc. No.	Location in South Dakota except where noted	Collectors and years	Remarks
D13444	SW 1/4 Sec. 25, T. 103 N., R. 73 W., Lyman Co.	G. A. Izett and W.A. Cobban, 1993	*Menuites oralensis* and *Baculites scotti* from concretions 4 m below base of Crow Creek Member
62-15-393	Approximate, N 1/2 SE 1/4 Sec. 22, T. 107 N., R. 7 0 W., Buffalo Co.	R.E. Stevenson, 1962?, 62H1, SD 47	Didymoceras nebrascense. The locality as given on the collecting slip is underlain by the Verendrye Member, which does not contain this ammonite.
D4899	NE 1/4 Sec. 28, T. 9. N., R. 7 E., Butte Co.	G.A. Izett and W.A. Cobban, 1997	Didymoceras nebrascense
D13787	SW 1/4 Sec. 16, T. 12 N., R. 3 E., Butte Co.	G.A. Izett and W.A. Cobban, 1997	Didymoceras nebrascense from gray limestone concretion
D13585 D13638 D13756	SE 1/4 Sec. 15, T. 104 N., R. 71 W., Brule Co.	G.A. Izett, 1994	Fragment of *Didymoceras puebloense?*, *Oxybeloceras* sp., *Baculites scotti*, and *Baculites* sp. from 25 cm below Crow Creek Member
	SE 1/4 Sec. 15, T. 104 N., R. 71 W., Brule Co.	G.A. Izett, 1994	*Baculites scotti* on shale slope 3 m below Crow Creek Member
D13582 D13475	SW 1/4 Sec. 6, T. 109 N., R. 75 W., Hughes Co.	G.A. Izett, 1993	*Baculites scotti* from shale slope 10 m below base of Crow Creek Member in bluff along Lake Sharpe
D13476	SE 1/4 Sec. 13, T. 104 N., R. 72 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites scotti* and *Menuites oralensis* from shale 10-15 m below base of Crow Creek Member
D13505	SE 1/4 Sec. 15, T. 104 N., R. 71 W., Brule Co.	G.A. Izett, 1993	*Baculites scotti* and *Didymoceras archiacianum* from shale near base of bluff 10 m below base of the Crow Creek Member.
D13757	SE 1/4 Sec. 14, T. 104 N., R. 72 W., Brule Co.	G.A. Izett, 1996	One small ammonite *(Menuites?)*
D13753	SE 1/4 Sec. 6 and NE 1/4 Sec. 7 T. 109 N., R. 76 W., Hughes Co.	G.A. Izett, 1996	*Baculites gregoryensis* and *Didymoceras* sp.
D13476	SE 1/4 Sec. 13, T. 104 N., R. 72 W., Lyman Co.	G.A. Izett and W.A. Cobban, 1993	*Baculites gregoryensis* from Gregory Member about 18 m below base of Crow Creek Member
D4966	SW 1/4 Sec. 3, T. 103 N., R. 73 W., Lyman Co.	W.A. Cobban, 1965	*Baculites perplexus* from lower part of the Gregory Member

Figure 1. A, Shocked microcline (below) and partially melted microcline (above) from the 206.8 m (678.5 ft) and 128.8 m (422.5 ft) level of the M-1 core of the Manson, Iowa, impact structure. B, Sanidine (Precambrian melted microcline clast) and adjacent melt matrix from the 114 m (374.8 ft) of the M-1 core of the Manson, Iowa, impact structure used in $^{40}Ar/^{39}Ar$ analyses.

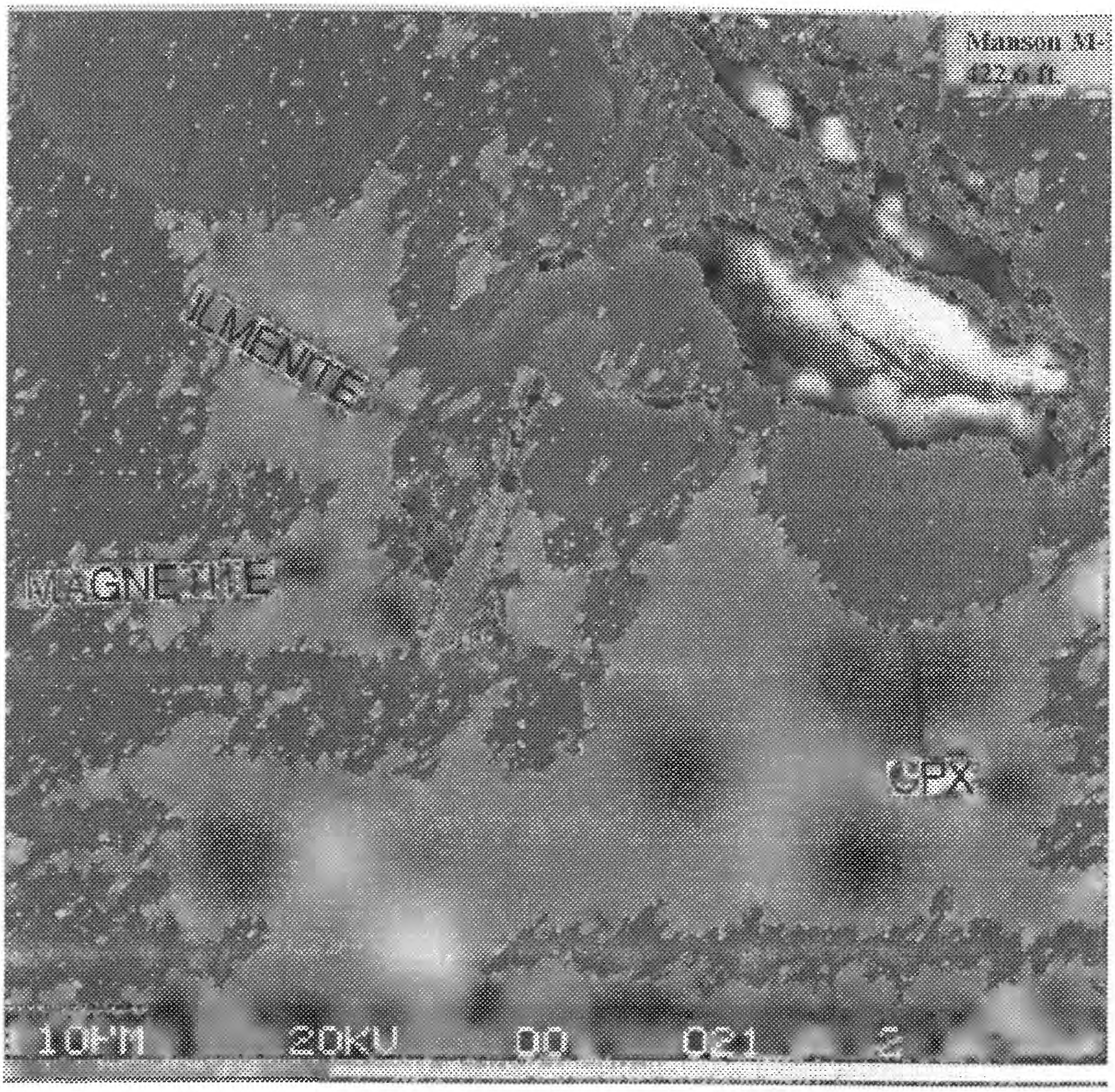


Figure 2. Coexisting titanomagnetite and ilmenite crystals in quartz halo surrounding a shock-metamorphosed quartz grain in core (422.6 ft.) of the M-1 core of the melt layer of the Manson impact structure

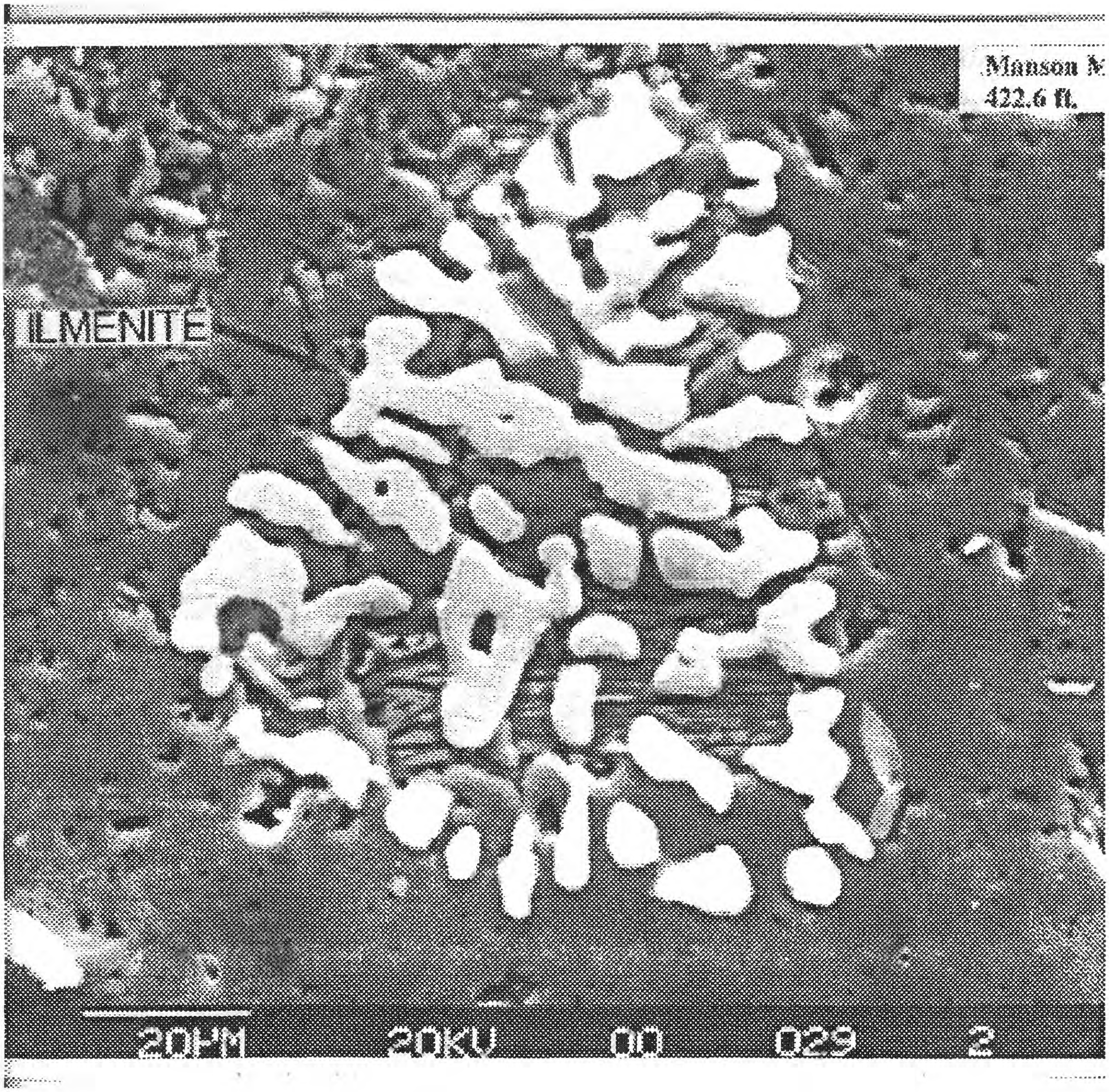


Figure 3. Skeletal ilmenite crystal in the matrix of the melt layer in the M-1 core (422.6 ft).

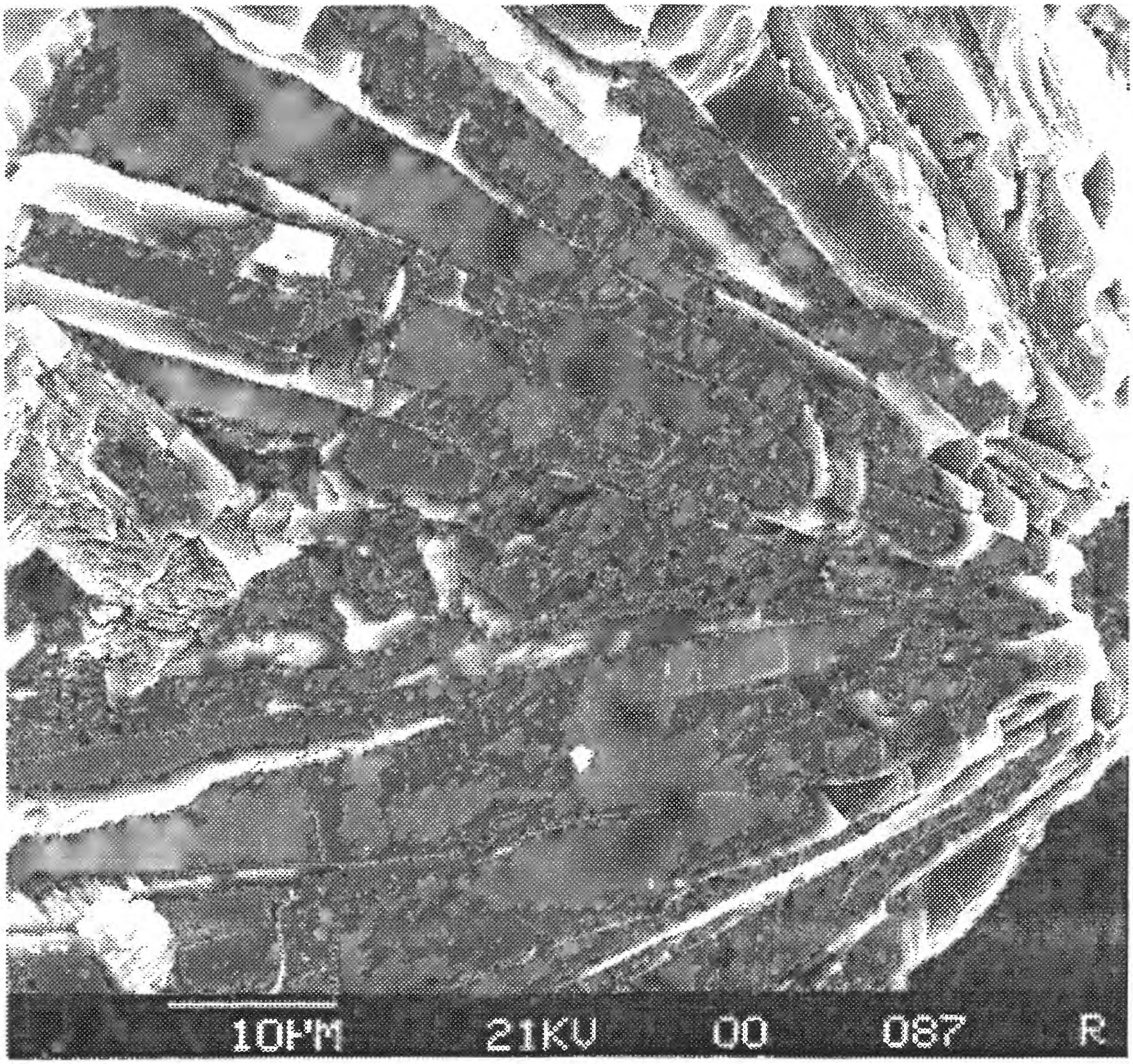


Figure 4. Scanning electron microscope image of spherulitic sanidine used in $^{40}Ar/^{39}Ar$ analyses of Appendix B-1.

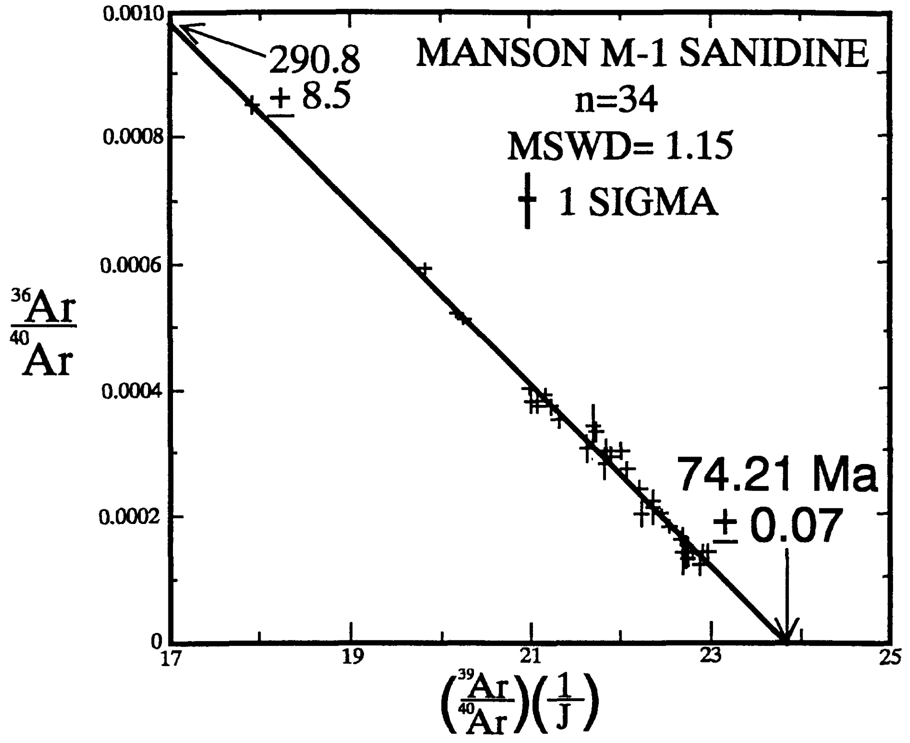


Figure 5. Inverse correlation diagram of data from 34 laser total-fusion $^{40}Ar/^{39}Ar$ analyses of sanidine from the 114 and 143 m levels of the M-1 core of the Manson, Iowa, impact structure.

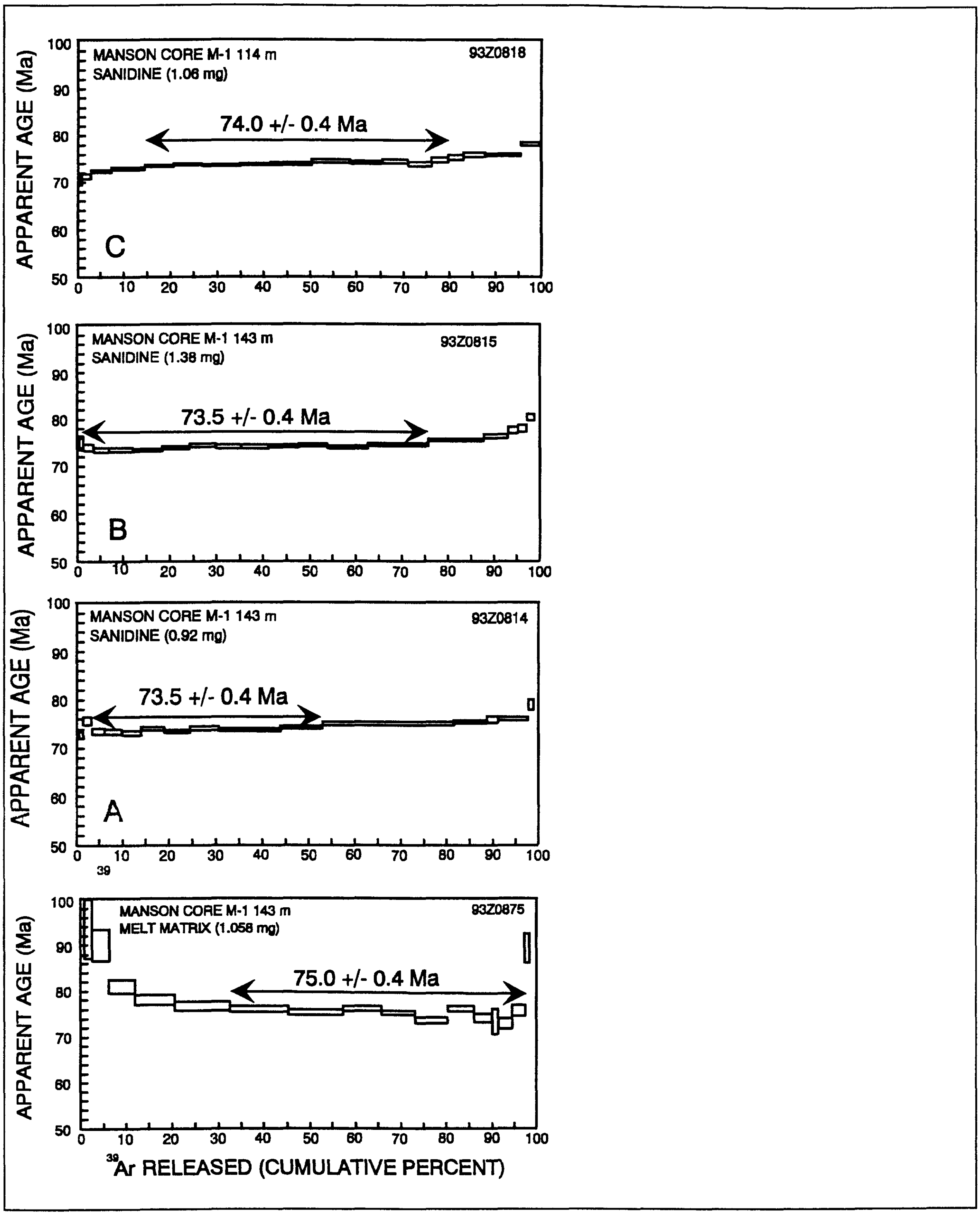


Figure 6. Incremental-heating $^{40}Ar/^{39}Ar$ age spectra of sanidine and melt matrix from the 114 and 143 m levels of the M-1 core of the Manson, Iowa, impact structure.

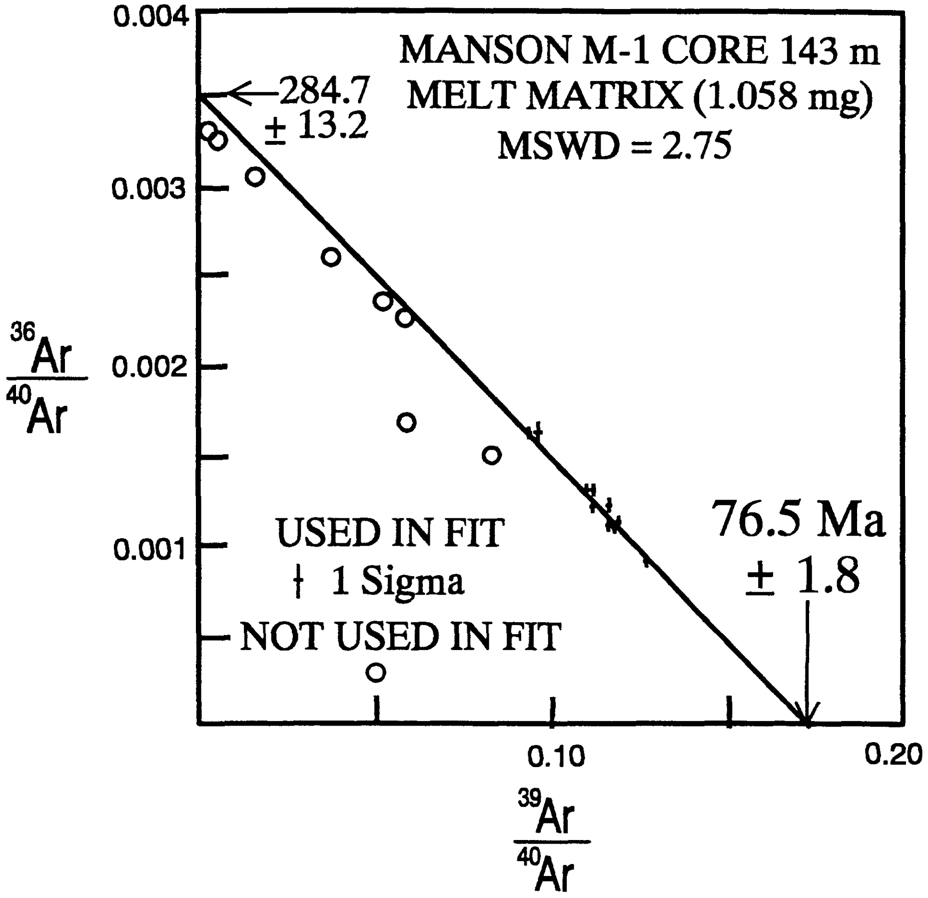


Figure 7. Inverse-correlation diagram of data from incremental heating $^{40}Ar/^{39}Ar$ analyses of melt matrix adjacent to a sanidine clast from the 143 m level of the M-1 core of the Manson, Iowa, impact structure.

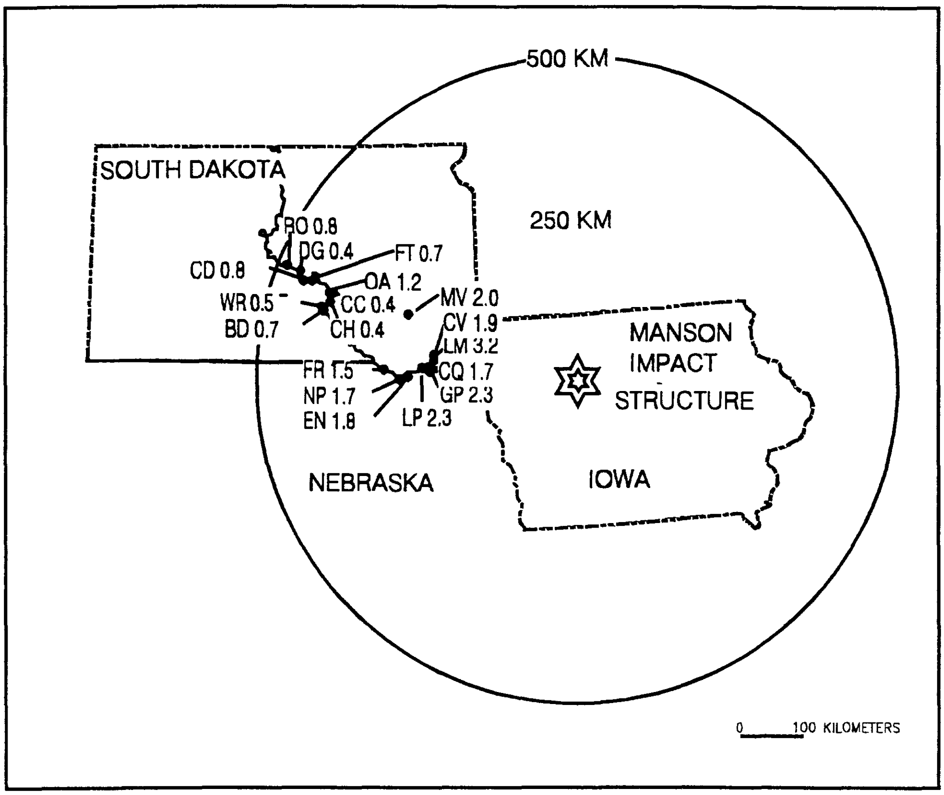


Figure 8. Map showing size (mm) of largest shock-metamorphosed quartz and feldspar grains in the Crow Creek Member of the Pierre Shale (Upper Cretaceous) and distance to the suspected source of the grains, the Manson, Iowa, impact structure. Precise geographic locations and size of shocked grains listed in Table 2. LM, Lake Marindahl; CV, Clay Valley; CQ, Cement Quarry; GP, Gavins Point; LP, Lakeport; EN, East of Niobrara; NP, Niobrara State Park; MV, Mount Vernon; FR, Fort Randall Cemetery; CH, Chamberlain; CC, Crow Creek; OA, Oacoma; BD, Black Dog; WR, White River; FT, Fort Thompson; CD, Cedar Creek; DG, DeGrey; RO, Rousseau.

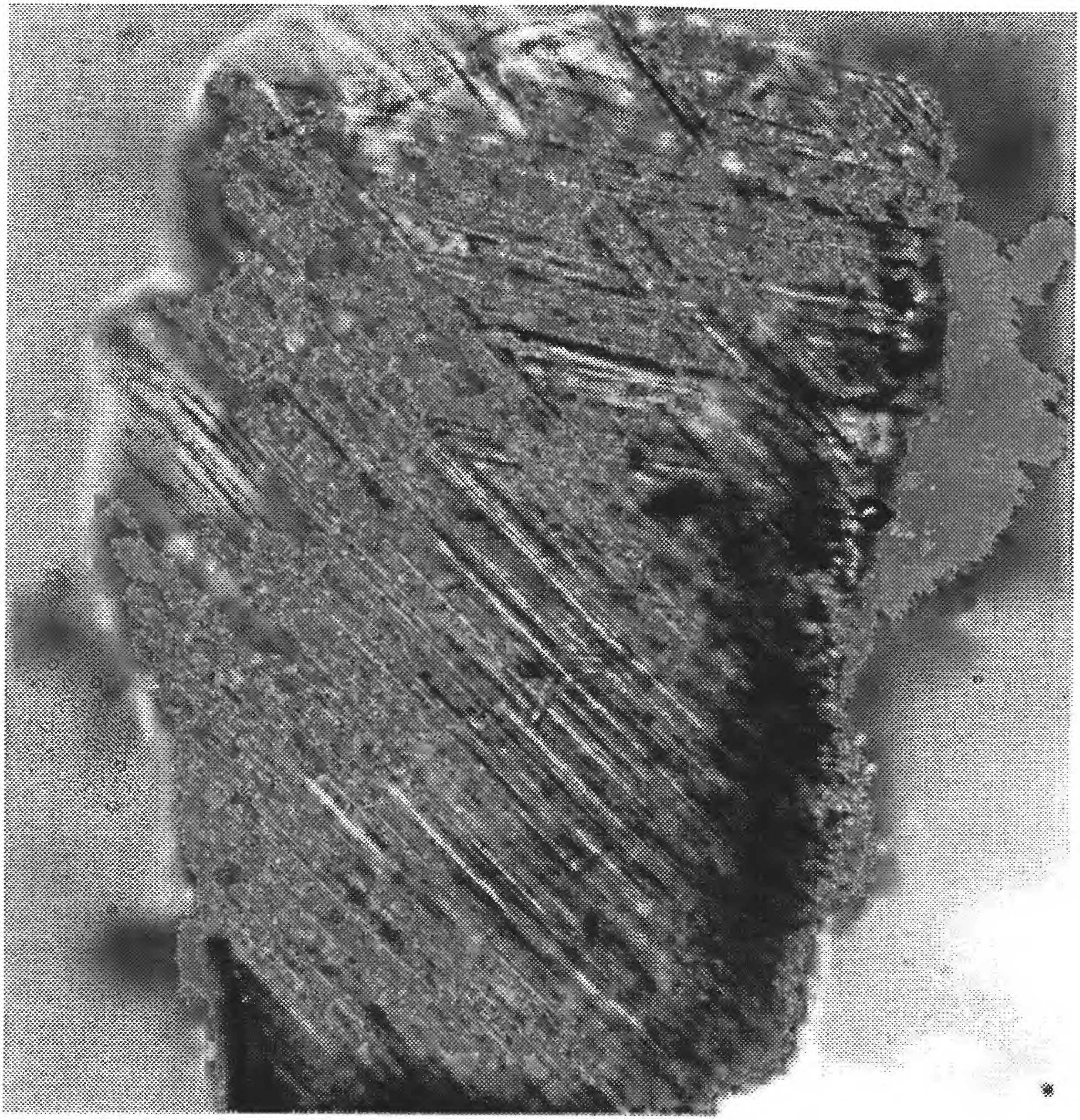
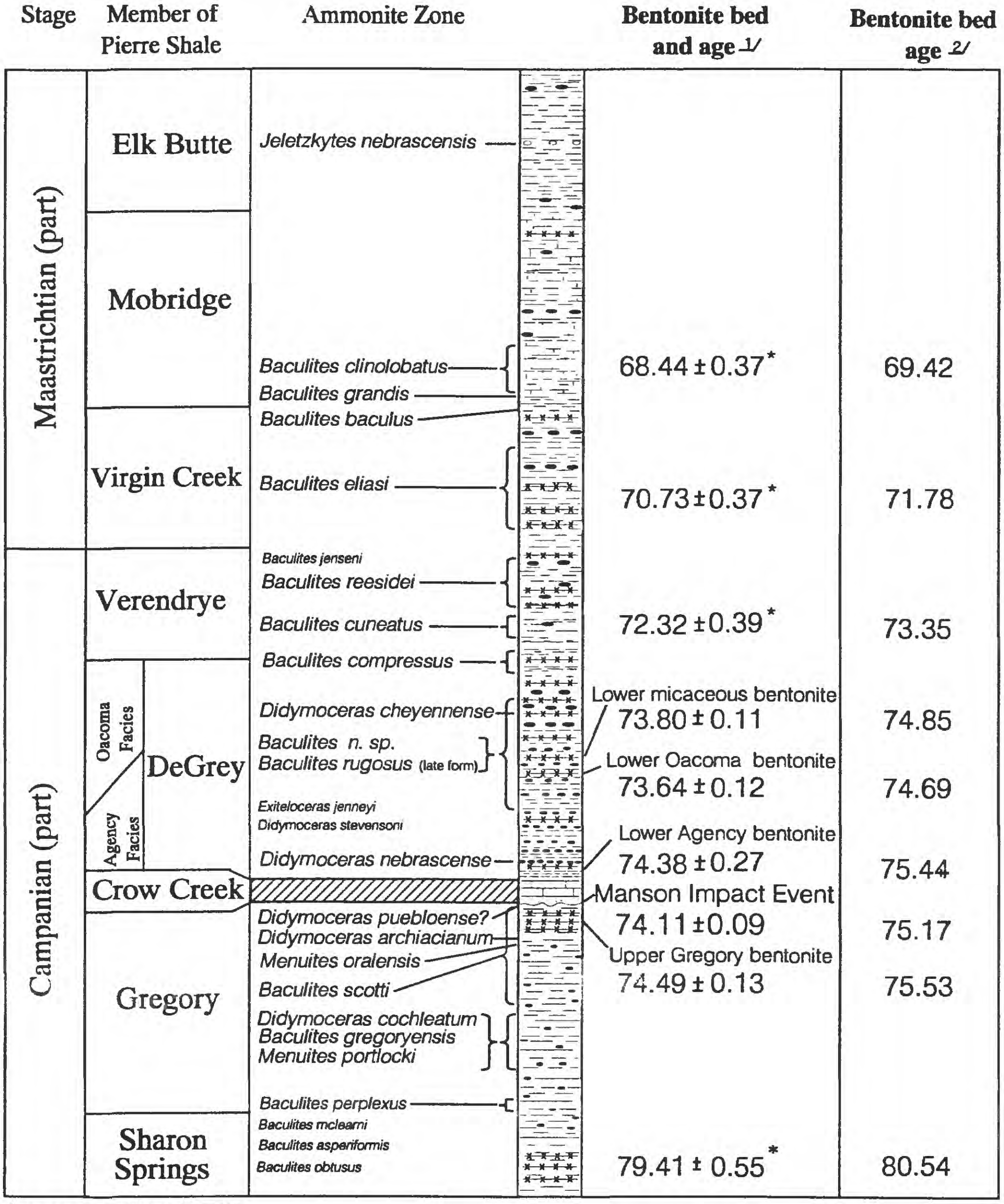


Figure 9. Shock-metamorphosed quartz grain (0.55 mm long from the lowest 4 cm of the Crow Creek Member of the Pierre Shale at the Lakeport [formerly referred to as the House of Mary Shrine locality (Izett and others , 1993b)] west of Yankton, S. Dak. Two sets of planar deformation features in this position. Mounted on tip of a spindle and photographed in plane polarized light.



*Obradovich [unpublished data, 1997]
1/ Age relative to age of Mmhb-1 hornblende 513.9 Ma
2/ Age relative to age of Mmhb-1 hornblende 520.4 Ma
Note: Ammonite names in small font not found in this study

Figure 10. Generalized stratigraphic column for the Pierre Shale (Upper Cretaceous) of central South Dakota showing ammonite zones and $^{40}Ar/^{39}Ar$ ages of sanidine and biotite obtained from bentonite beds below and above the Crow Creek Member of the Pierre Shale.

Table 1. Summary of laser total-fusion $^{40}Ar/^{39}Ar$ ages of 1) sanidine from melt rock of the Manson, Iowa, impact structure (M-1 core), 2) minerals from bentonite beds stratigraphically below and above the Crow Creek Member of the Pierre Shale in central South Dakota, and 3) bentonite beds of late Campanian age in the Pierre Shale of Colorado

Location Stratigraphic position	Mineral	Irrad. No.	No. of Analyses	Age ± σ_{best} (Ma)
Oacoma, S. Dak.				
Lower Oacoma bentonite	Sanidine	JDO22	15	73.64 ± 0.12
Bentonite 4.3 m above base of DeGrey Mbr.	Biotite	JDO22	5	73.64 ± 0.21
Fort Pierre, S. Dak.				
Lower micaceous bentonite	Sanidine	JDO21	17	73.80 ± 0.11
Base of Oacoma facies of DeGey Mbr.	Biotite	JDO21	6	73.79 ± 0.20
Rousseau, S. Dak.				
Lower Agency bentonite	Biotite	JDO24	5	74.38 ± 0.27
Bentonite 3.0 m above base of DeGrey Mbr.				
South of Colorado Springs, Colo.				
Bentonite, base of *Exiteloceras jenneyi* zone	Sanidine	GLN12	10	74.22 ± 0.12
in Pierre Shale		GLN13	5	74.13 ± 0.21
		JDO21	6	74.15 ± 0.20
		TOTAL	21	74.18 ± 0.09
Manson, Iowa				
M-1 core, 114m	Sanidine	GLN9	1	74.17 ± 0.50
		GLN12	5	74.22 ± 0.28
		GLN13	6	74.16 ± 0.19
		JDO21	6	74.15 ± 0.20
		JDO22	7	73.98 ± 0.21
M-1 core, 143 m	Sanidine	JDO13	9	74.09 ± 0.16
		TOTAL	34	74.11 ± 0.09
North of Trinidad, Colo.				
Bentonite in *Didymoceras nebrascence* zone	Oligoclase	GLN12	5	74.14 ± 0.20
in Pierre Shale		GLN13	13	74.13 ± 0.21
		TOTAL	18	74.13 ± 0.14
Oacoma, S. Dak.				
Bentonite near top of *Baculites scotti* zone	Sanidine	GLN12	9	74.47 ± 0.15
in the Gregory Mbr.		GLN13	4	74.54 ± 0.23
		TOTAL	13	74.49 ± 0.13
	Biotite	JDO24	6	75.02 ± 0.06

Note: Ages relative to a fluence-monitor age of 27.92 Ma for Taylor Creek Rhyolite based on an age for Mmhb-1 hornblende of 513.9 Ma. To convert ages relative to Mhhb-1 of 520.4, Ma multiply the ages of this table by 1.0143.

Table 2. Size of the largest shock-metamorphosed mineral grains in the Crow Creek Member of the Pierre Shale (Upper Cretaceous) and the distance (km) to the Manson, Iowa, impact structure

County	Site	Location		Size	Distance (km)
		Lat °N	Long °W	(mm)	from MIS
Yankton	Lake Marindahl	43 02 15	97 15 14	3.24	215
Yankton	Clay Valley	43 04 06	97 16 42	1.90	220
Yankton	Cement Quarry	42 52 51	97 28 44	1.72	235
Cedar*	Gavins Point	42 50 27	97 29 07	2.28	245
Yankton	Lakeport	42 52 54	97 32 13	2.30	250
Knox*	East of Niobrara	42 44 52	97 59 17	1.79	275
Knox*	Niobrara State Park	42 46 21	98 04 03	1.70	285
Davison	Mount Vernon	43 38 21	98 16 03	2.00	305
Gregory	Fort Randall Cemetery	43 02 33	98 34 11	1.50	315
Lyman	Chamberlain	43 48 21	99 19 26	0.45	410
Buffalo	Crow Creek	43 58 06	99 18 33	0.43	410
Lyman	Oacoma	43 48 24	99 23 24	1.20	415
Lyman	Black Dog	43 41 33	99 31 23	0.70	415
Lyman	White River	43 44 31	99 36 02	0.46	425
Lyman	Cedar Creek	44 05 23	99 43 02	0.26	455
Hughes	DeGrey	44 16 14	99 53 55	0.43	460
Hughes	Rousseau	44 18 29	100 03 32	0.80	475

Note: Locations determined by a Global Positioning System hand-held calculator
* Located in Nebraska, all others in South Dakota