

Carbon Isotope Excursion in Atmospheric CO₂ at the Cretaceous-Tertiary Boundary: Evidence from Terrestrial Sediments

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A -1.5% to -2% carbon isotope excursion immediately above the clay layer that defines the Cretaceous-Tertiary (K/T) boundary has been reported in marine sediments world wide. This paper reports a similar -1.5% to -2.8% carbon isotope excursion recorded by C3 land plants from three temporally-controlled, stratigraphically-constrained terrestrial sections in the Western Interior of North America (Garfield County, Montana, and Slope County, North Dakota). Carbon isotope measurements of bulk sedimentary organic carbon were well-correlated with those of isolated plant cuticle, suggesting that the terrestrial organic carbon signature in these sediments parallels that of plant cuticle. Carbon isotope signatures were also independent of rock type and depositional environment, showing that the carbon isotope signature of plants, although altered, is not biased taphonomically. Because the signature in terrestrial facies records the isotope composition of paleoatmospheric CO₂, this record—combined with that from marine sections—offers additional insight into changes in carbon cycling underlying the K/T negative carbon isotope excursion. For example, radiometric age determinations from the Hell Creek Road locality in Montana bracket the atmospheric carbon isotopic recovery between 65.00 ± 0.05 Ma and 65.16 ± 0.04 Ma. This reflects a more rapid recovery for the terrestrial biosphere than for that of the marine realm, perhaps due to lower extinction rates in land plants than in marine primary producers.

INTRODUCTION

A -1.5% to -2% carbon isotope excursion in surface ocean dissolved inorganic carbon has been documented in association with the Cretaceous-Tertiary (K/T) boundary impact at the global stratotype section in El Kef, Tunisia (Keller and Lindinger, 1989; Keller et al., 1995), and at other marine sections world wide (Hsü et al., 1982; Perch-Nielsen et al., 1982; Zachos and Arthur, 1986; Stott and Kennett, 1989; Zachos et al., 1989; Stott and Kennett, 1990; Robin et al., 1991; Zachos et al., 1992; D'Hondt et al., 1998). Homogenization of the normally-positive marine surface-to-deep water carbon isotope gradient at the K/T boundary has been attributed to a geologically brief cessation of primary productivity in the surface ocean (Hsü et al., 1982; Zachos and Arthur, 1986; D'Hondt et al., 1998).

¹²C added to the surface ocean-atmosphere carbon reservoir also would be required to produce the negative anomaly (Kump, 1991; Ivany and Salawitch, 1993). Recovery of marine primary productivity during the early Paleocene drew down the negative surface ocean-atmosphere carbon isotope anomaly in a few hundreds of thousands of years (Keller and Lindinger, 1989; Robin et al., 1991; D'Hondt et al., 1998). However, the positive surface-to-deep ocean gradient was not restored fully for up to 3 Ma perhaps due to decreased organic matter transport into the deep ocean (D'Hondt et al., 1998).

This paper reports a similar -1.5% to -2.8% carbon isotope excursion recorded by C3 land plants from three temporally-controlled, stratigraphically-constrained fully terrestrial sections. Because the signature in terrestrial facies records the isotope signature of atmospheric CO₂ (Arens et al., 2000), this record—combined with that from the marine realm—constrains the possible mechanisms underlying the K/T surface ocean negative carbon isotope excursion.

GEOLOGIC SETTING

To document a terrestrial carbon isotope anomaly at the K/T boundary, sections must be examined where the boundary has been diagnosed by the iridium- and shocked mineral-bearing boundary clay that documents the terminal Cretaceous impact (Alvarez et al., 1980; Bohor et al., 1984; Robin et al., 1991; Alvarez et al., 1995). Two well-studied sections in Garfield County, Montana (Hell Creek Road and Hauso Flat), and one in Slope County, North Dakota (Pyramid Butte; Fig. 1), were chosen. Iridium- and shocked-mineral enriched boundary clays mark the terminal Cretaceous impact event at all three localities: Hauso Flat = 11.7 ppb iridium (Smit and van der Kaars, 1984) and 2.22 ppb iridium (Swisher et al., 1993), Hell Creek Road = 0.57 ppb iridium (Baadsgaard et al., 1988), Pyramid Butte = 0.72 ppb iridium (Johnson et al., 1989). The Hauso Flat section discussed herein is approximately one kilometer east of the section discussed by Swisher et al. (1993) and that of Smit and van der Kaars (1984). However, strata are exposed continuously over this distance and were traced laterally from the previous sample sites to the locality reported herein.

Several additional lines of evidence converge to chronostratigraphically bracket the K/T boundary in the chosen sections. At all three localities, the K/T boundary occurs

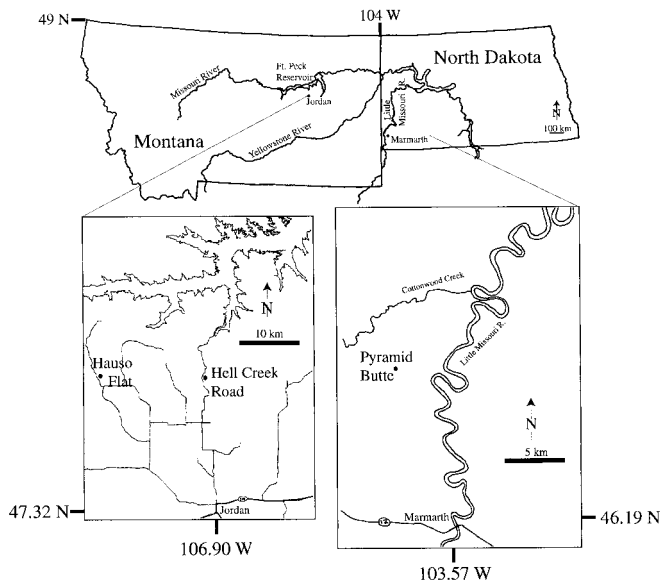


FIGURE 1—The locations of Hauso Flat, Hell Creek Road, and Pyramid Butte sections in Montana and North Dakota, U.S.A.

within the 29R magnetochron (Archibald, 1982; Johnson and Hickey, 1990; Swisher et al., 1993). Both eastern Montana sections also contain volcanic ash beds that provide radiometric control on the timing and duration of the carbon isotope excursion (Fig. 2; Swisher et al., 1993). In the Hell Creek Road section (equal to Hell Creek Recreation Area Access Road locality), the K/T boundary clay is preserved at the base of a 145 cm thick lignite, which defines the local contact between the Hell Creek Formation and the Tullock Member of the Fort Union Formation (Collier and Knechtel, 1939). At Hell Creek Road, this lignite preserves several ash beds, from which radiometric dates have been obtained, stratigraphically above the iridium-bearing boundary clay (radiometric dates from Swisher et al., [1993] are reproduced on Fig. 2). At Hauso Flat, the iridium-bearing boundary clay is preserved at the base of a 7–9 cm lignite, which also contains a volcanic ash from which radiometric dates have been obtained (Smit and van der Kaars, 1984; Swisher et al., 1993). Palynostratigraphy in all three sections documents the characteristic loss of Cretaceous indicator pollen taxa (e.g., *Wodehouseia spinata* and *Aquilapollenites* spp.) over the 50–300 cm surrounding the iridium-enriched boundary clay (Smit and van der Kaars, 1984; Hotton, 1988; Johnson et al., 1989). At Hauso Flat, Hotton (1988) reported the stratigraphically highest occurrence of typical Cretaceous *Aquilapollenites* species about 3 m below the iridium-bearing lignite, roughly coincident with the highest *in situ* dinosaur remains. Smit and van der Kaars (1984) reported typically Cretaceous palynomorphs up to 50 cm below the iridium enrichment. At Hell Creek Road, typical Cretaceous pollen taxa are present to the base of the lignite (Hotton, 1988).

Four main lithofacies were encountered in the sections studied: cross-stratified and ripple laminated sandstone facies, siltstone facies, carbonaceous mudstone facies, and a facies dominated by organic accumulation (lignite). Sandstone facies were generally fine to very fine-grained;

grains were moderately well-rounded and well-sorted. Individual beds were planar to lenticular and may be structureless or characterized by cross-stratification, ripple or plane-parallel lamination. Fossil lags are not uncommon on basal scour surfaces or on the concave surface of trough or ripple laminae. Fossil hash including small fragments of coalified wood may be present within the sandstone. Root traces are rare but present in sandstone facies. Iron-rich (siderite or limonite) concretions are common. Sandstone facies in these sections are interpreted as overbank and crevasse splay deposits (Fastovsky, 1987). None of the sections described herein contain channel sandstone bodies.

Siltstone facies dominate the upper Hell Creek Formation in the three sections studied. Rocks are buff to gray, generally bedded, structureless, with variable fractions of silt, sand, and clay-size particles. Clays are generally mixed-layer smectite/illite, which accounts for the characteristic “popcorn weathering” associated with this lithology (Fastovsky and Dott, 1986). Root traces are common, but other evidence of soil development (ped structure, slicken slides, clay skins) are rare, suggesting that depositional events were frequent. Siltstone facies are interpreted as deposition on a moderately well-drained vegetated floodplain (Fastovsky and Dott, 1986). Most siltstone facies show some degree of pedogenic development. In the sections studied, most would be classified as the Maka pedotype of Retallack (1994). This pedotype preserves root traces, but soil formation has not progressed far enough to generate pedogenic structure or the formation of clay skins. Retallack (1994) estimated that these soils formed rapidly (100–500 years), which suggests high depositional rates in the upper Hell Creek Formation. Some paleosols showing greater degrees of pedogenic development (Sapakot, Otssko, and Spatsiko pedotypes of Retallack, 1994) are present in the upper Hell Creek Formation, but they are not present in the sections described herein.

Organic-rich facies (carbonaceous mudstone and lignite) were present in the uppermost Hell Creek and at the Hell Creek—Fort Union formational contact, suggesting higher water tables up section. Carbonaceous shales commonly are laminated and occur in relative thin beds (commonly less than 20 cm). Carbonaceous shales are laterally continuous at the scale of ten of meters, but are not regional stratigraphic markers. Lignites, including the Z-coal that marks the base of the Fort Union Formation, range from 10 cm to 2 m in thickness. Lignites are also laterally continuous at the outcrop scale but time transgressive regionally (Fastovsky and Dott, 1986; Fastovsky, 1987; Johnson et al., 1989; Swisher et al., 1993). Lignites (Sik pedotype of Retallack, 1994) may be interbedded with very fine grained clastic sediment and volcanic ash, which has yielded radiometric dates. The iridium-bearing K/T boundary clay is preserved only in lignite facies. Carbonaceous mudstone in the Hell Creek Formation likely represent deposition in flood plain swales, with organic accumulation truncated by renewed clastic input. The planar geometry and fine-grained underclays associated with lignites in these sections suggests accumulation on the floodplain under conditions of low clastic sedimentary influx and relatively high water table.

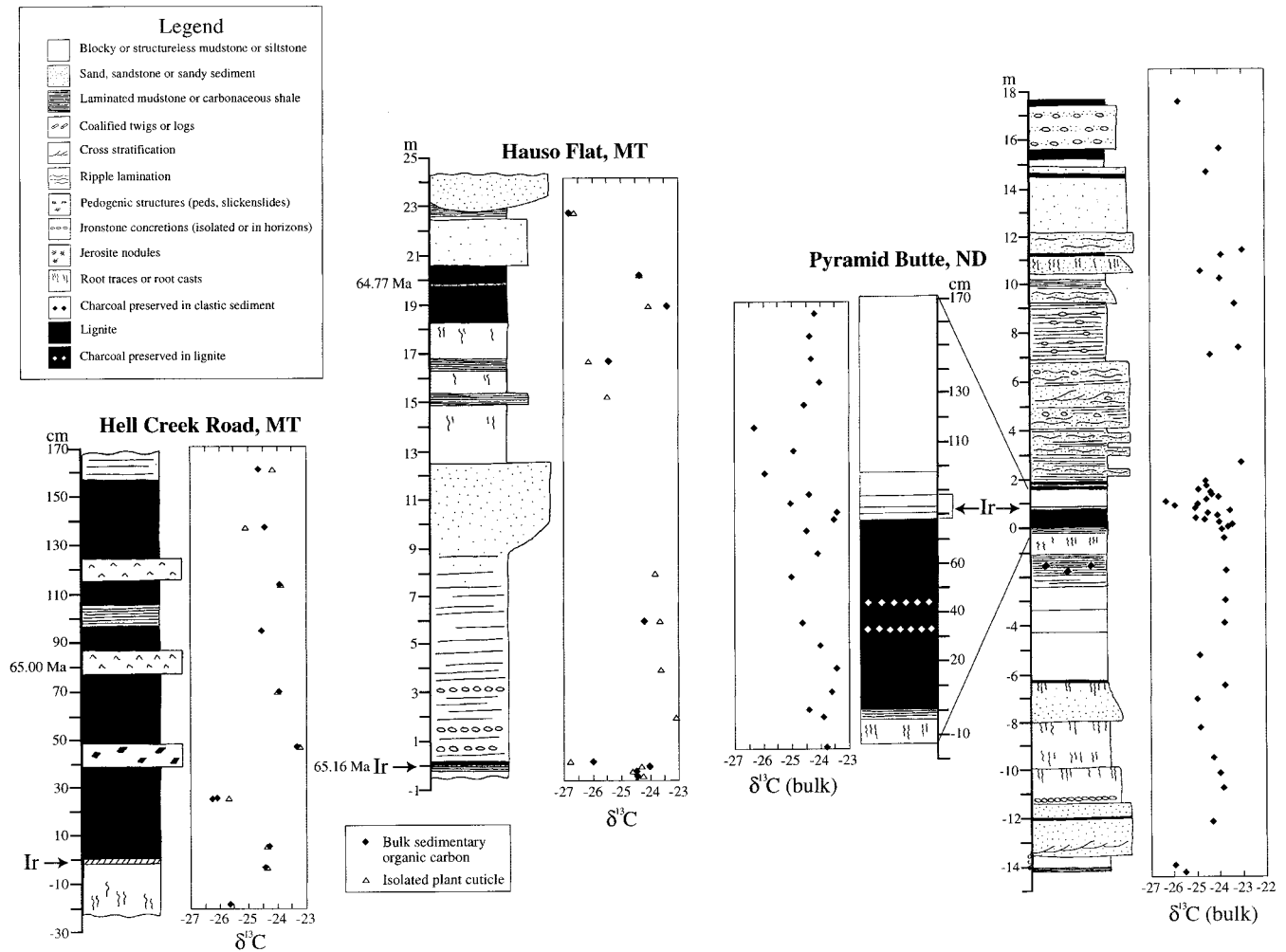


FIGURE 2—Carbon isotope data from bulk organic carbon plotted against measured sections from Hell Creek Road and Hauso Flat, Montana, and Pyramid Butte, North Dakota. The stratigraphic positions of the iridium-bearing boundary clay (Iridium) and radiometrically dated volcanic ash layers (Swisher et al., 1993) are noted in each section. Measurement error bars for carbon isotope data are smaller than plot symbols in all cases.

TAPHONOMY AND SAMPLING OF TERRESTRIAL ORGANIC CARBON

Sediment samples for carbon isotope analysis were collected from measured sections (Fig. 2) made in 1997 and 1998. The Pyramid Butte, North Dakota, section was sampled within the context of sections measured by K.R. Johnson, but stratigraphy was verified in the field by N.C.A. Samples were taken from floodplain siltstone and organic-rich facies (carbonaceous mudstone and lignite). Sandstone preserves little organic material and was seldom sampled. Sampling of well-developed paleosols (those bearing clearly-developed pedogenic structure, clear soil horizons, and clay skins such as the Sapakot, Otssko and Spatsiko pedotypes of Retallack, 1994) was avoided because of possible alteration of the organic carbon isotope signal due to within-soil carbon recycling (Wedin et al., 1995).

Because C3 land plants sample atmospheric CO_2 during photosynthesis and fractionate it in a known way during carbon fixation (Farquhar et al., 1989), the isotope composition of plant tissue reflects the composition of the at-

mosphere under which it was fixed (Medina et al., 1986; Van der Merwe and Medina, 1989), with a quantifiable error due to physiological and ecological vital effects (Arens et al., 2000). Therefore, secular variation observed in the isotope signature of fossil plant tissue recorded variation in paleoatmospheric carbon isotope composition with a statistically-derived error due to plant vital effects (Arens et al., 2000). The correlation between carbon isotope composition of plant tissue and atmospheric CO_2 was strongest when the carbon isotope composition of several co-occurring plants was sampled and averaged (Lloyd and Farquhar, 1994; Arens et al., 2000). For most moist climates, this sampling regime effectively minimizes variation due to ecological and physiological vital effects (Arens et al., 2000). In the geologic record, it is assumed that the organic carbon of many plants is averaged spatially as it is incorporated into floodplain sediments. Therefore, measuring the isotope composition of bulk organic carbon in these sediments approximates the many-species average sample that best minimizes physiological vital effect bias.

Carbon isotope samples were analyzed using only ge-

TABLE 1—Carbon isotope data for bulk sedimentary organic carbon and isolated plant cuticle for the Hauso Flat, Hell Creek Road, Montana, sections, and the Pyramid Butte, North Dakota, section. Data are listed in relative stratigraphic position.

Hauso Flat			Hell Creek Road			Pyramid Butte	
Sample ID	$\delta^{13}\text{C}_{\text{bulk}}$	$\delta^{13}\text{C}_{\text{cuticle}}$	Sample ID	$\delta^{13}\text{C}_{\text{bulk}}$	$\delta^{13}\text{C}_{\text{cuticle}}$	Sample ID	$\delta^{13}\text{C}_{\text{bulk}}$
MT97-011	-24.62	-24.15	MT97-099	-26.62	-26.80	ND98-197	-25.74
MT97-010	-24.42	-25.10	MT97-098	-24.65	-24.35	ND98-196	-23.94
MT97-009	-23.91	-24.33	MT97-097	-24.04	-23.38	ND98-195	-24.49
MT97-008		-23.86	MT97-096	-26.12	-25.41	ND98-194	-22.94
MT97-007	-24.51		MT97-095	-25.49		ND98-193	-23.88
MT97-006	-23.94	-24.00	MT97-094	-23.82		ND98-192	-24.80
MT97-005	-23.30	-23.21	MT97-093	-23.65	-24.19	ND98-191	-23.94
MT97-004	-26.08	-25.67	MT97-092	-23.62		ND98-190	-23.31
MT97-003	-24.26	-24.36	MT97-091	-23.09		ND98-189	-22.90
MT97-002	-24.42	-25.67	MT97-090	-26.77	-25.99	ND98-188	-23.12
MT97-001	-25.62	-25.49	MT97-089	-24.31	-24.01	ND98-187	-24.35
			MT97-088	-24.63	-24.47	ND98-186	-23.01
			MT97-087	-24.24	-24.45	ND98-185	-24.56
						ND98-184	-24.53
						ND98-183	-24.88
						ND98-182	-24.18
						ND98-181	-24.35
						ND98-180	-24.30
						ND98-179	-23.99
						ND98-178	-24.54
						ND98-177	-26.28
						ND98-176	-24.92
						ND98-175	-25.92
						ND98-174	-24.37
						ND98-173	-25.02
						ND98-172	-23.41
						ND98-171	-23.51
						ND98-170	-22.61
						ND98-169	-24.48
						ND98-168	-24.06
						ND98-167	-24.99
						ND98-166	-24.61
						ND98-165	-23.97
						ND98-164	-23.42
						ND98-163	-23.58
						ND98-162	-24.39
						ND98-161	-23.87
						ND98-160	-23.77
						ND98-159	-23.69
						ND98-158	-23.74
						ND98-157	-23.78
						ND98-156	-24.84
						ND98-155	-23.77
						ND98-154	-24.96
						ND98-153	-24.83
						ND98-152	-24.27
						ND98-151	-23.99
						ND98-150	-23.82
						ND98-149	-24.30
						ND98-148	-25.96
						ND98-147	-25.47

neric collection number identification. At the Hell Creek Road locality 11 bulk sedimentary organic carbon samples were analyzed over an 185 cm K/T boundary-crossing section; at Hauso Flat nine bulk sedimentary organic carbon samples were analyzed over a 24 m interval; at Pyramid Butte, 51 bulk sedimentary organic carbon samples were analyzed over a 32 m interval, with 22 samples spanning a 160 cm K/T boundary-crossing section (Table 1). For bulk organic carbon measurements, a 1–5 g sample of the rock was acidified in 1M HCl overnight to remove carbonate. The remaining bulk organic residue was analyzed for car-

bon isotope value in triplicate. All samples were prepared for stable isotope analysis by combustion in sealed tubes containing Cu, CuO, and Ag (Minagawa et al., 1984). Released CO₂ was purified cryogenically, and collected for ¹³C/¹²C measurement on an Optima mass spectrometer. All isotope values are reported in the $\delta^{13}\text{C}$ notation:

$$\delta = \frac{R_{\text{sample}} - R_{\text{standard}}}{R_{\text{standard}}} \times 1000[\text{‰}]$$

where the reported standard is PDB limestone with $R_{\text{standard}} = {}^{13}\text{C}/{}^{12}\text{C} = 0.011237$.

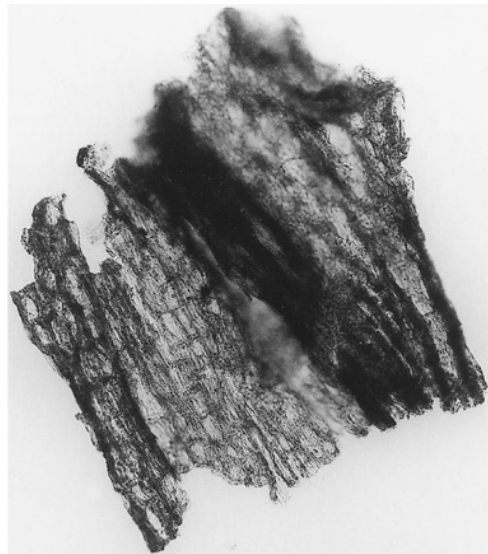
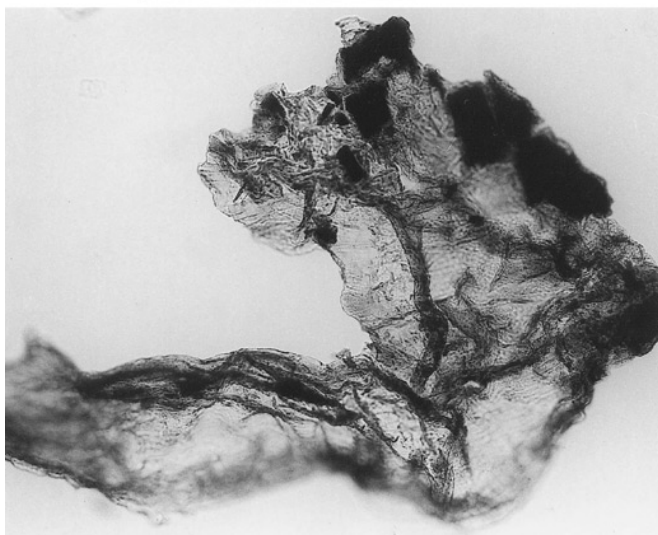
50 μm 

FIGURE 3—Photomicrographs of plant cuticle fragments isolated from sediment samples from the Hauso Flat, Montana section: MT97-090 (above); MT97-100 (below). Note ranks of epidermal cell imprints.

Before attempting to interpret a terrestrial carbon isotope signature, possible taphonomic biases that might influence the $\delta^{13}\text{C}$ value of sedimentary organic material must be evaluated. Organic material in sediment will contain carbon derived from organisms other than plants. To segregate the C3 plant signal, isolated plant cuticle also was sampled. Cuticle is the biopolymer coating the epidermis of vascular land plants to inhibit water loss. In a study of more than 10 angiosperm and conifer species, including C3, C4, and CAM physiologies (G.R. Upchurch pers. comm. 1998), cuticle carbon isotope values were 3‰ to 3.5‰ less than whole plant carbon isotope composition with a consistent error of less than $\pm 1\%$ (Upchurch et al., 1997).

However, plant cuticle cannot be isolated from all sediment samples. Therefore, it is necessary to know the rela-

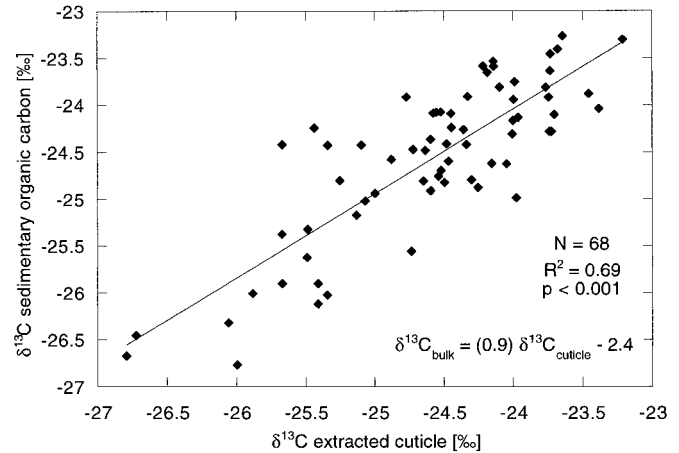


FIGURE 4—Regression of bulk sedimentary organic carbon $\delta^{13}\text{C}$ values onto $\delta^{13}\text{C}$ values from mixed samples of plant cuticle isolated from the same sediment sample. The resulting relationship was statistically significant ($R^2 = 0.69$, $p < 0.001$). The slope was statistically significant ($p < 0.001$, $\text{SE} = 0.1$); the y-intercept was not statistically significant ($p = 0.2$).

tionship between bulk sedimentary organic carbon isotope composition and that of isolated plant cuticle. To explore this question, bulk sedimentary organic carbon isotope values were regressed onto those derived from a mixed sample of plant cuticle isolated from the same sediment ($N = 68$). Cuticle was hand picked from demineralized (54% HF, three days) sediment; cuticle was identified by its characteristic morphology, including the regular pattern of epidermal cell imprints (Fig. 3). Carbon isotope analyses of isolated cuticle were performed as above. Regression residuals were normally distributed (Lilliefors $p = 0.452$) with a mean close to zero (mean = 0.06); variances were homoscedastic. A significant relationship could be seen between the carbon isotope values of bulk sedimentary organic carbon and those from isolated cuticle ($R^2 = 0.69$, $p < 0.001$, Fig. 4). The slope of this relationship was statistically significant ($p < 0.001$) and approximately equal to one, implying a single step chemical conversion between leaves (from which cuticle was derived) and sedimentary organic matter. A similar relationship between plant tissue and soil organic matter has been observed in both tropical (Medina et al., 1986) and temperate forests (Balser et al., 1993). The y-intercept (-2.4%) was not statistically significant ($p = 0.2$), but was similar in magnitude and sign to the difference between the carbon isotope composition of soil organic material and canopy leaves in a modern tropical forest (Medina et al., 1986). This suggests that the bulk organic carbon isotope signature sampled from these floodplain facies was produced mainly by cuticle and chemically-related, decay-resistant plant material. Therefore, carbon isotope measurements of bulk sedimentary organic material should track secular variation in the carbon isotope composition paleoatmospheric CO_2 . However, analysis of cuticle should be conducted wherever possible as a “check” for unusual alteration of bulk sedimentary organic carbon values.

To evaluate the possible effect of depositional environment variation (e.g., peat-forming wetlands vs. well-drained flood plains) on carbon isotope value, 411 bulk or-

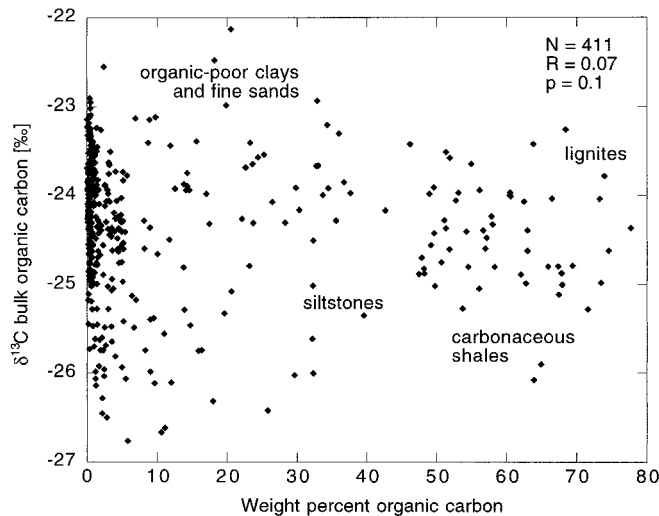


FIGURE 5—Bulk carbon isotope data from 411 sediment samples collected from sections in eastern Montana and southwestern North Dakota plotted against weight percent organic carbon content. Weight percent organic carbon content is a proxy for rock-type variation in these sections and corresponds to field descriptions. Sandstones and some clay-rich rocks have low organic carbon content, lignites and carbonaceous shale may be up to 80% carbon by weight.

ganic carbon isotope measurements were regressed onto weight-percent organic carbon values (Fig. 5). Weight percent organic carbon content was a useful proxy for depositional environment in these facies: Carbon-poor samples correspond to floodplain siltstone facies, while carbon-rich samples generally represent carbonaceous mudstone and lignite. Regression residuals were distributed normally (Lilliefors $p = 0.143$) with a mean close to zero (mean = 0.003). Variances were homoscedastic. There was no significant correlation ($R = 0.07$, $p = 0.1$) between carbon isotope value and weight-percent organic carbon content, suggesting that carbon isotope signature does not vary systematically with depositional environment.

To detect any relationship between rock type and carbon isotope value, each carbon isotope measurement was coded for rock type (lignite, carbonaceous shale, claystone/mudstone, siltstone, and sandstone) based on field observations. Because each rock type was represented by a different number of measurements, a balanced sample ($N = 39$ for each rock type) of lignite, carbonaceous shale, claystone/mudstone, and siltstone values was selected randomly for analysis. The nonparametric Kruskal-Wallis test was chosen for this analysis because carbonaceous shale failed the homogeneity of variance requirement for parametric methods. The Kruskal-Wallis test failed to reject ($p = 0.219$) the null hypothesis that all measurements were sampled from the same population. Analysis of five randomly selected samples of lignite, carbonaceous shale, claystone/mudstone, and siltstone yielded similar results ($p = 0.160$). Hence, carbon isotope value does not vary systematically with rock type. Together, these analyses show that there is no significant or systematic taphonomic bias in carbon isotope measurements of terrestrial organic material in the system studied. Therefore, secular variation observed in organic carbon $\delta^{13}\text{C}$ values

from sedimentary sections can be attributed primarily to the carbon isotope signature of the atmosphere.

RESULTS

Figure 2 presents carbon isotope data from the Hauso Flat, Hell Creek Road, and Pyramid Butte sections in which the K/T boundary was diagnosed by an iridium- and shocked mineral-enriched impact clay. In both Montana sections, bulk sedimentary organic carbon $\delta^{13}\text{C}$ values for samples taken below the impact layer were approximately -24.3‰ . In both sections, a similar value was recorded in organic material sampled immediately above the impact clay. Values immediately above the boundary clay may represent vegetation killed by impact-related disturbance. Because these plants fixed their carbon under the pre-impact atmosphere, their tissue retains the latest Cretaceous atmospheric signature. Alternatively, these samples may document a temporal lag between the impact event and its consequences for the terrestrial carbon cycle.

Samples 2–3 cm above the impact clay have $\delta^{13}\text{C}$ values of -25.9‰ and -26.8‰ at Hell Creek Road and Hauso Flat, respectively. At Hauso Flat, organic carbon $\delta^{13}\text{C}$ values recovered to -23.9‰ within 220 cm above the impact clay. At Hell Creek Road, $\delta^{13}\text{C}$ values recovered to -23.3‰ 22 cm above the boundary. The stratigraphic difference in recovery interval at these sections resulted primarily from sampling interval, which was much finer at the Hell Creek Road locality (Fig. 2). At Pyramid Butte in North Dakota, samples immediately below the boundary clay had bulk $\delta^{13}\text{C}$ values of approximately -23.5‰ ; 5 cm above the iridium-enriched clay a $\delta^{13}\text{C}$ value of -26.3‰ was recorded. Values return to -24.5‰ within about 40 cm of the iridium-bearing clay, and to -24.0‰ at 50 cm.

CHANGES IN CARBON CYCLING ASSOCIATED WITH THE K/T BOUNDARY

The homogenization of the marine surface-to-deep water carbon isotope gradient immediately following the K/T boundary event has been attributed to cessation of primary productivity in the surface ocean, the so-called "Strangelove ocean" (Keller and Lindinger, 1989; Stott and Kennett, 1989; Zachos et al., 1989; Zachos et al., 1992; D'Hondt, 1998; D'Hondt et al., 1998). Under normal conditions, photosynthetic plankton preferentially take up ^{12}C , leaving surface ocean dissolved inorganic carbon with greater $\delta^{13}\text{C}$ values than those of deeper strata, where isotopically depleted organic matter is oxidized (Kroopnick et al., 1970; Kroopnick, 1980). Immediately after the terminal-Cretaceous impact, mortality and extinction in photosynthetic plankton erased the difference between surface and deep ocean carbon isotope signatures (D'Hondt et al., 1998). Although plankton populations recovered quickly from K/T boundary mortality, initial recovery of the surface-to-deep water carbon isotope gradient occurred over approximately 500,000 years (D'Hondt et al., 1998), while reestablishment of mechanisms that transport reduced carbon into the deep ocean may not have occurred for 1–3 Ma (Keller and Lindinger, 1989; Zachos et al., 1989; Zachos et al., 1992; D'Hondt et al., 1998). However, homogenization of marine carbon isotope values alone does not account for the negative $\delta^{13}\text{C}$ excursion recognized after the

boundary impact. The negative anomaly requires input of ^{12}C from some external source.

A variety of sources have been proposed. Kump (1991) argued that river flux into a surface ocean devoid of life following the terminal Cretaceous impact would produce a -5% carbon isotope excursion. However, this anomaly is larger than that observed in either the marine or terrestrial records. Furthermore, studies of the plankton biota show that marine productivity recovered rapidly after the terminal Cretaceous event (Keller, 1988; Keller and Lindinger, 1989; D'Hondt et al., 1998). Li and Keller (1999) suggested that negative excursions in the $\delta^{13}\text{C}$ value of the surface ocean carbon reservoir could be produced by increased organic carbon transfer from continents to the ocean during times of low sea level and cool climate. They have noted several negative excursions in surface ocean $\delta^{13}\text{C}$ values that correspond with Maastrichtian lowstands inferred from lithological and paleontological data (Li and Keller, 1999; Li et al., 1999). Ivany and Salawitch (1993) proposed that burning approximately one quarter of Maastrichtian terrestrial biomass in post-impact wildfires introduced $^{12}\text{CO}_2$ into the atmosphere. There is evidence for wildfire at the K/T boundary (Wolbach et al., 1985). However, there is little independent evidence for the magnitude of destruction proposed (e.g., Keller and MacLeod, 1993).

The recognition of a negative $\delta^{13}\text{C}$ excursion in post-impact terrestrial land plants helps refine hypotheses surrounding the mechanism and timing of changes in global carbon cycling associated with the K/T boundary. In a high-stratigraphic-resolution marine section at El Kef, Tunisia, the negative carbon isotope anomaly at El Kef begins 2 cm below the maximum iridium enrichment and coincides with an increase in organic carbon deposition (Keller and Lindinger, 1989). Furthermore, the planktonic foraminifera *Pseudotextularia deformis*, which became extinct at the K/T boundary, showed a negative carbon isotope change in its last recorded occurrence (Keller and Lindinger, 1989). These observations suggest that lower $\delta^{13}\text{C}$ values in the surface ocean may be related to input of continentally-derived organic material (Li and Keller, 1999) coupled with reduced primary productivity in the surface ocean (Li et al., 2000). Li and colleagues concluded that these conditions were linked to climate-driven sea level change that preceded the terminal-Cretaceous event, rather than with the bolide impact itself (Li and Keller, 1999; Li et al., 1999).

The marine results were corroborated in terrestrial facies at York Canyon, New Mexico, where the trend toward more negative $\delta^{13}\text{C}$ values begins several centimeters below the peak iridium enrichment (Schimmelmann and DeNiro, 1984). Lowest $\delta^{13}\text{C}$ values in this section occurred 10 cm above the peak iridium enrichment, as in other Western Interior localities discussed herein. In the Pyramid Butte, North Dakota, section (Fig. 2), instability in the terrestrial $\delta^{13}\text{C}$ signature was noted below the iridium-bearing clay. However, this pattern was not repeated in either Montana locality, where the K/T boundary clay also is preserved in association with lignite facies. However, a taphonomic cause for this difference cannot be ruled out. In both of the Montana localities, the iridium- and shocked mineral-bearing K/T boundary clay is preserved at the base of associated lignite bed. In contrast, the K/T bound-

ary clay caps the lignite at Pyramid Butte, North Dakota. The latest Cretaceous carbon isotope instability recorded in the lignite at Pyramid Butte may have been lost by non-deposition or erosion at the facies transition in eastern Montana. However, there is no strong sedimentological evidence for erosion (scours, lags, load casts, etc.) at the base of the Hell Creek Road and Hauso Flat lignites. Also, the similarity in absolute magnitude of the last $\delta^{13}\text{C}$ value stratigraphically below the K/T boundary clay at all three localities argues against this interpretation (Hell Creek Road $\delta^{13}\text{C} = -24.4$, Hauso Flat $\delta^{13}\text{C} = -24.3$, Pyramid Butte $\delta^{13}\text{C} = -24.5$, Arens and Jahren, in press). Further examination of other sections will test this hypothesis.

Insight into the pace of post-K/T carbon cycle recovery comes from terrestrial sections for which there is radiometric time control. A volcanic ash directly overlying the iridium-bearing boundary clay at Hauso Flat was dated radiometrically at 65.16 ± 0.04 Ma (Swisher et al., 1993). Assuming that the boundary clay represents a single, instantaneous event, this age determination can be extended to the Hell Creek Road and Pyramid Butte boundary clays as well. A volcanic ash 110 cm above the Hell Creek Road boundary clay was dated radiometrically to 65.95 ± 0.05 Ma (Swisher et al., 1993). No dateable ash deposits were preserved in North Dakota. In the Hell Creek Road section, $\delta^{13}\text{C}$ value of the paleoatmosphere returned to pre-boundary or greater values within 40–60 cm above the impact clay—at least 20 cm below the 65.00 ± 0.05 Ma volcanic ash preserved in this section. Therefore, carbon isotope recovery can be bracketed radiometrically between 65.16 ± 0.04 Ma (the iridium-bearing impact bed) and 65.03 ± 0.05 Ma volcanic ash (Swisher et al., 1993)—a maximum interval of $130,000 \pm 5,000$ years. Thus, atmospheric carbon isotope rebound was not only similar in magnitude, but also in duration to that observed in the “early recovery” phase of the surface ocean signature reported by D'Hondt et al. (1998), Keller et al. (1989), and Hsü et al. (1982). Rapid rebound of the atmospheric carbon isotope signal contrasts with D'Hondt et al.'s (1998) conclusion that the normal marine surface-to-deep-carbon isotope gradient did not reestablish for approximately 3 Ma. The more rapid rebound of the atmospheric signature suggests that recovery of the terrestrial vegetation, and the carbon cycling it controls, may have been much faster than that of the marine trophic structure. This was likely due to lower extinction rates in land plants (single-locality extinction based on palynomorphs averaged about 30%; Nichols et al., 1986; Johnson et al., 1989) relative to marine microplankton (up to 80% for planktonic foraminifera; Keller, 1988).

The high time-stratigraphic control and resolution in some terrestrial sections—such as that on the Hell Creek Road—helps constrain mechanistic hypotheses of carbon flux during the period of global carbon cycle disruption following the K/T boundary impact. The geologically rapid (130–170 Ka) recovery of the atmosphere-surface ocean-carbon isotope signal suggests that processes driving this excursion were decoupled from primarily oceanographic processes, such as deep-ocean mixing. This conclusion agrees with the modeled results of Kump (1991) and interpretations from empirical data by D'Hondt et al. (1998). Furthermore, it is now necessary to look for sources of ^{12}C input to the atmosphere and surface ocean pools to ac-

count for the negative $\delta^{13}\text{C}$ excursion at the K/T boundary. Because atmospheric pCO_2 for the late Cretaceous and Paleocene have been estimated (Bernier, 1991; 1994; 1998) and the carbon isotope values of various carbon sources and sinks are known, or can be reconstructed (Siegenthaler and Sarmiento, 1993; Schlesinger, 1997; D'Hondt, 1998; Arens et al., 2000), these sources and sinks can now be modeled quantitatively and tested against the time scales associated with the brief, post-Cretaceous atmospheric carbon isotope anomaly.

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