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Notes

Chemostratigraphic correlation of four fossil-bearing sections in southwestern North Dakota

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ABSTRACT

A -1.5‰ to -2‰ carbon isotope shift in surface ocean dissolved inorganic carbon has been reported stratigraphically above the Cretaceous-Tertiary (K-T) boundary in the global stratotype section at El Kef, Tunisia, and in many other marine sections worldwide. Because a change in the $\delta^{13}\text{C}$ value of paleoatmospheric CO_2 would have accompanied the marine surface carbon isotope shift, the K-T carbon isotope anomaly can be used to diagnose and correlate the boundary in terrestrial sections as well. We use this shift and other secular variation in the carbon isotope signature from the Hell Creek and Fort Union Formations to correlate four sections in Slope County, North Dakota. At Pyramid Butte, the K-T boundary is marked by a -2.8‰ carbon isotope shift approximately 10 cm above the iridium-bearing impact clay. This shift is larger in magnitude than might be expected by plant vital effects or taphonomic variation. The Pyramid Butte carbon isotope shift allowed correlation to the Bobcat Butte (shift of -2.1‰), Terry's Fort Union Dinosaur (shift of -1.1‰ to -1.6‰), and the New Facet Boundary (shift of -2.3‰) sections, where an iridium enrichment was not preserved. An interval of carbon isotope fluctuation marks the transition from HCII to HCIII megafloreal zones in the Bobcat Butte section, suggesting that some instability in the physical environment was correlated with the observed vegetation change. These data demonstrate that carbon isotope chemostratigraphy can provide an additional line of evidence that is largely independent from taxonomic extinction for correlating fossil-bearing terrestrial sections. Chemostratigraphic interpretations may also recognize incomplete or condensed sections, which will aid interpretation of the fossils found in them.

CHEMOSTRATIGRAPHY AS AN AID TO TERRESTRIAL CORRELATION

Testing mechanistic hypotheses of extinction and biotic change across the Cretaceous-Tertiary (K-T) boundary requires

that the stratigraphic position of the boundary be placed accurately. In the global stratotype section at El Kef, Tunisia, a clay layer containing above-background concentrations of iridium records the terminal-Cretaceous bolide impact and defines the K-T boundary (Keller et al., 1995; Keller and Lindinger, 1989).

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A similar clay layer, including impact-related shocked mineral and blast melt spherules, has been reported in terrestrial sediments and diagnoses the K-T boundary in these facies (eg., Alvarez, 1987; Bohor et al., 1984; Bohor et al., 1987a; Orth et al., 1981). However, these diagnostic components are uncommon and ephemeral in most subaerial depositional environments, making diagnosis of the K-T boundary difficult in terrestrial rocks.

To circumvent this problem, terrestrial stratigraphers have looked for other indicators associated with the K-T boundary clay, including lithostratigraphy, paleomagnetic zones, radioisotope dates, last appearances of characteristic Cretaceous pollen forms, and first appearances of typical Tertiary mammals. The first attempts to identify the K-T boundary were based on lithostratigraphy. In 1938, Roland Brown argued that both the Hell Creek–Fort Union formational contact and the chronostratigraphic K-T boundary should be placed at the base of the lowest regionally continuous lignite bed that stratigraphically overlay the last nonreworked dinosaur fossil (Brown, 1938, 1952). This approach proved unsatisfactory first because reference lignite units form diachronous complexes rather than a single isochronous datum (Archibald, 1982; Fastovsky, 1987; Lofgren, 1995), and second because a number of the basal Fort Union lignite beds have subsequently been determined to be Paleocene in age (see Clemens, this volume; Nichols and Johnson, this volume; Pearson et al., this volume; Johnson et al., 1989). Paleomagnetism documents periodic reversals of Earth's magnetic field. In the Western Interior region of North America, the K-T boundary occurs within the 29r magnetochron (Archibald et al., 1982; Butler and Lindsay, 1985; Johnson and Hickey, 1990; Johnson et al., 1989; Lerbekmo, 1985), which lasted less than one million years (Harland et al., 1989).

In terrestrial sections with high sedimentation rates, Lerbekmo et al. (1996) reported a brief (~50 k.y.) normal polarity subchron at the K-T boundary in Alberta and Saskatchewan. However, this subchron has not been reported in all sections studied, perhaps because of the discontinuous nature of terrestrial sedimentation in most locations. Radioisotope dates from volcanic ash provide isochronic markers within some sections (Swisher et al., 1993). Several ash beds were preserved within the Fort Union Formation in northeastern Montana, but no suitable volcanic ashes have been identified in southwestern North Dakota. Palynological and mammalian biostratigraphic markers diagnose the boundary by the last appearance of characteristic Maastrichtian pollen types (Hotton, 1988; Johnson et al., 1989; Nichols, 1990; Nichols and Fleming, 1990; Nichols et al., 1986; Nichols and Sweet, 1993; Orth et al., 1981; Tschudy et al., 1984) or first appearance of Paleocene mammals (Archibald et al., 1987; Lofgren, 1995; Woodburne, 1977; Woodburne and Swisher, 1995).

Biostratigraphy can leave zones of ambiguity and gaps where fossils were not preserved (Archibald, 1982) or give confusing results when fossils were reworked (Lofgren et al., 1990). Interpretation of biostratigraphic markers can also be

complicated by biogeographic variation (Nichols and Sweet, 1993). First- and last-occurrence data are also insensitive to unconformities, which may be important to interpreting rates of biotic change. More significantly, biostratigraphic time markers introduce circularity to inference of extinction and biotic change within an interval, particularly when one uses the same fossils both to define zones and to discuss biotic patterns. Therefore no biostratigraphic method alone provides the biologically independent, high-resolution correlations across the wide range of terrestrial facies needed to test hypotheses of pattern and process associated with biotic change before and at the K-T boundary.

Carbon isotope chemostratigraphy may provide an additional line of evidence for correlating terrestrial sections. As outlined by Williams et al. (1988), isotope chemostratigraphy relies on temporal variability in the primary isotope composition of a chemical system. To develop a correlation, the pattern of this temporal variability is matched among stratigraphic sections. Williams et al. (1988) defined five general features of isotope records that provide useful correlation tie points: (1) cycles, which may be of uniform or variable duration, are repeating quasiperiodic components of an average signal; (2) events are unique excursions distinctly above or below the average signal; (3) shifts are abrupt (<200 k.y.) deviations from the average signal, which may be reversible; (4) long-term trends are steady directional changes in isotope composition through time; and (5) absolute magnitude may establish boundary conditions, distinguish among shifts, or identify events. Williams et al. (1988) proposed that correlations initially be developed by matching secular patterns among sections. When secular variation for an interval is well understood, new sections may be matched to a regional or global reference curve.

Isotope chemostratigraphy has been extensively applied to marine sediments. For example, Adabi (1997) compared single carbonate carbon isotope absolute magnitudes from the Renison Formation, Tasmania, to global compilation curves (Derry et al., 1992; Kaufman and Knoll, 1995) to conclude that these rocks were Neoproterozoic, not Cambrian, in age. Bartley et al. (1998) used organic carbon isotope curves to correlate the Platonovskaya Formation of Russia with other Siberian Platform sections. Body and trace fossils could bracket Platonovskaya sedimentation only between Vendian and earliest Cambrian. Comparison of both isotope shifts and absolute magnitudes through the section allowed correlation to the Precambrian-Cambrian (Yudomian-Nemakit–Daldynian) boundary and highlighted the probable presence of an unconformity at that horizon. At the Permian-Triassic boundary in Spitsbergen, Wignall et al. (1998) used organic carbon isotope data to diagnose the P-T boundary and to recognize that the sequence was indeed conformable, contrary to previous conclusions based on biostratigraphy (Nakamura et al., 1987; Stemmerik, 1988; Szaniawski and Malkowski, 1979). At the Cenomanian-Turonian boundary, Gale et al. (1993) demonstrated the synchronous nature of biostratigraphic and carbon isotope chemostratigraphic

markers in both Sussex, England, and Pueblo, Colorado. Their data confirm the global character of late Cenomanian and early Turonian carbon isotope trends. Koch et al. (1992) extended the approach to terrestrial sediments by correlating the marine carbon isotope curve at the Paleocene-Eocene boundary with that developed from mammalian tooth enamel. At the K-T boundary global stratotype section in El Kef, Tunisia, a reversible carbon isotope shift has been documented in association with the iridium anomaly (Keller et al., 1995; Keller and Lindinger, 1989). This shift, along with other features of the latest Cretaceous and Paleocene (Shackleton, 1985; Shackleton and Hall, 1984) carbon isotope curve may provide similar resolution to correlations both at the K-T boundary and within the Hell Creek and Fort Union Formations.

K-T BOUNDARY CARBON ISOTOPE SHIFT IN THE ATMOSPHERE LINKS MARINE AND TERRESTRIAL RECORDS

A -1.5‰ to -2.0‰ reversible carbon isotope shift in surface ocean dissolved inorganic carbon has been documented in association with the K-T boundary impact at the global stratotype section (Keller et al., 1995; Keller and Lindinger, 1989) and at other marine sections worldwide (D'Hondt et al., 1998; Hsü et al., 1982; Perch-Nielsen et al., 1982; Robin et al., 1991; Smit, 1982; Stott and Kennett, 1989, 1990; Zachos and Arthur, 1986; Zachos et al., 1989; Zachos et al., 1992). Most workers attribute homogenization of the normally positive surface-to-deep water carbon isotope gradient to breakdown of marine primary productivity in the wake of the bolide impact (D'Hondt et al., 1998; Hsü et al., 1982; Zachos and Arthur, 1986). However, homogenization of marine carbon isotope values alone does not fully account for the negative carbon isotope shift observed in planktonic foraminifera after the boundary impact. The negative shift requires input of ^{12}C from another source. Based on modeling results, Kump (1991) concluded that fluvial input into a postimpact ocean devoid of life would cause the surface ocean to come into isotope equilibrium with weathering input over non-steady-state time scales ($<10^5$ years). This would produce a -5‰ carbon isotope shift, much larger than that observed in marine sediments (Kump, 1991). Also based on model results, Ivany and Salawitch (1993) proposed that burning $\sim 25\%$ of terrestrial biomass in postimpact wildfires introduced $^{12}\text{CO}_2$ into the atmosphere, lowering atmospheric carbon isotope values and raising pCO_2 . However, apart from soot found in several K-T boundary sections (Wolbach et al., 1985), there is little independent evidence for a conflagration of this magnitude. Whatever the source, $^{12}\text{CO}_2$ added to the atmosphere would rapidly come into mass and isotope equilibrium with the surface ocean (Schlesinger, 1997; Siegenthaler and Sarmiento, 1993), thus lowering marine carbon isotope values, and producing the observed K-T boundary shift. Such a mechanism would produce carbon isotope shifts of similar sign and magnitude in both the surface ocean and atmosphere. Since

land plants sample atmospheric CO_2 directly and discriminate in a known way during C_3 carbon fixation (Farquhar et al., 1989), the isotope composition of plant tissue reflects the composition of the atmosphere under which it was fixed (Arens et al., 2000; 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, the post-Cretaceous carbon isotope shift recorded in surface ocean dissolved inorganic carbon should also be preserved in the plant tissue and terrestrial organic carbon derived from it.

A reversible carbon isotope shift has been reported in terrestrial K-T boundary sections in northeastern Montana and southwestern North Dakota (Arens and Jahren, 2000). At the Hell Creek Road and Hauso Flat localities in Garfield County, Montana, carbon isotope shifts of -1.6‰ and -2.5‰ have been reported $\sim 2\text{--}3$ cm above the iridium-enriched K-T boundary clay (Arens and Jahren, 2000). At the Pyramid Butte locality in North Dakota, a shift of -2.8‰ was recorded 5 cm above the K-T boundary clay (Arens and Jahren, 2000). A similar isotope shift was reported at several localities in the Raton Basin of New Mexico and Colorado (Holmes and Bohor, 1994; Lomax et al., 1998; Schimmelmann and DeNiro, 1984). At York Canyon, New Mexico, a -1.8‰ carbon isotope shift occurs 10 cm above the peak iridium enrichment (there is no clear boundary clay preserved at this locality; Schimmelmann and DeNiro, 1984). At two other localities, reversible negative carbon isotope shifts were accompanied by anomalies in sulfur isotopes, interpreted as sulfuric acid rain in the aftermath of impact (Holmes and Bohor, 1994; Lomax et al., 1998). A negative shift of similar magnitude (-1.4‰ to -1.8‰) was also reported in extracted $n\text{-C}_{29}$ alkane, a land plant biomarker, from marine rocks in Caravaca, Spain (Arinobu et al., 1999).

Therefore, an atmospheric carbon isotope shift stratigraphically above the K-T boundary offers a tool for correlating among terrestrial sections, particularly in the absence of an iridium-enriched boundary clay. This tool provides a time-stratigraphic framework in which to correlate fossil-bearing sections using data that are not autocorrelated to the biotic hypotheses that one may wish to test. Because plant-derived organic carbon is common in terrestrial rocks of this region, the technique can be applied to most fossil-bearing sections. More important, the absence of a carbon isotope shift within a suspected K-T boundary section offers powerful evidence of paraconformity—a common point of debate in many discussions of extinction and biotic change (Lofgren, 1995). Furthermore, a carbon isotope anomaly of similar magnitude and sign has been documented at the global K-T stratotype section at El Kef (Keller and Lindinger, 1989), thus making the carbon isotope signature directly diagnostic of the boundary event.

METHODS

Sediment samples for carbon isotope analysis were collected in July 1998 at localities where stratigraphic sections had

been measured by K.R. Johnson. Sediment samples for bulk carbon isotope analysis were identified only by generic collection numbers in order to preclude any possibility of bias because of stratigraphy or field location. For bulk organic carbon measurements, a 1–5 g sample of the rock was acidified in 1M HCl overnight to remove carbonate. A standard weight sample of the remaining bulk organic residue was analyzed for $\delta^{13}\text{C}$ value in triplicate. Uncertainty in each carbon isotope measurement associated with mass spectrometry is $\pm 0.05\%$. All samples were prepared for stable isotope analysis by combustion in sealed tubes containing Cu, CuO, and Ag (Minagawa et al., 1984). Released CO_2 was purified cryogenically and collected for $^{13}\text{C}/^{12}\text{C}$ 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 $R_{\text{standard}} = ^{13}\text{C}/^{12}\text{C} = 0.011237$. Weight percent organic carbon content of sediment samples was calculated from these data.

Plant cuticle for carbon isotope analysis was picked by hand from demineralized (54% HF for three days) sediment; cuticle morphology, including the characteristic pattern of epidermal cell imprints, was verified under $100\times$ magnification. Cuticle is the waterproof coating on aerial surfaces of vascular land plants. Although it is of mixed chemical composition (De Leeuw et al., 1995), carbon isotope values of plant cuticle parallel those of whole leaf tissue with a standard error of less than $\pm 1\%$ (Upchurch et al., 1997).

SECTIONS STUDIED

To date, we have analyzed four K-T boundary crossing sections in the vicinity of Marmarth, North Dakota (Fig. 1). Stratigraphic sections figured below were based on Johnson's unpublished data and additional observations in the field. Stratigraphic positions are described with reference to the Hell Creek–Fort Union formational contact.

The Pyramid Butte section (KRJ Sections 86160 and 86161, Slope County, North Dakota; Johnson et al., 1989) has been extensively studied because it is the only section in southwestern North Dakota to yield an iridium (0.72 ppb Ir) anomaly at the K-T boundary. At Pyramid Butte, the iridium-bearing boundary clay overlies a lignite, the base of which marks the contact between the Hell Creek Formation and the Ludlow Member of the Fort Union Formation. Characteristic Cretaceous pollen taxa including *Wodehouseia spinata* and several species of *Aquilapollenites* become rare in the 20 cm above the iridium-bearing clay (see also Nichols and Johnson, this volume; Johnson et al., 1989). Plant macrofossils typical of the FUI floral zone (Johnson and Hickey, 1990), including "*Populus*" *nebrascensis*, *Dicotylophyllum anomalum*, and *Paranymphaea crassifolia*, are common in the mudstone immediately

overlying the iridium-bearing clay (Johnson et al., 1989). We sampled for carbon isotope stratigraphy at 5 cm intervals through the 2 m that embrace the iridium-bearing boundary clay. The complete section at Pyramid Butte is ~ 60 m thick. Outside the recognized boundary interval, we sampled at 50 cm to 100 cm intervals in the Hell Creek Formation and, where possible (commonly 2–3 m intervals), in the overlying Ludlow Member. At Pyramid Butte, the Ludlow Member of the Fort Union Formation was relatively sandy and highly weathered; sandy sediments preserve little organic matter, and extensive weathering may alter carbon isotope values. Therefore, we focused our sampling on organic or silt-rich horizons within the Ludlow Member.

The Bobcat Butte section (KRJ Section 86148, Slope County, North Dakota) offered a long, continuous, well-exposed Hell Creek Formation section. A K-T boundary interval in the mudstone overlying the lowest lignite bed of the Fort Union Formation has been recognized by the loss of characteristic Cretaceous pollen types of the *Wodehouseia spinata* Assemblage Zone (see Nichols and Johnson, this volume). No iridium anomaly has been identified. Bobcat Butte also offered direct stratigraphic ties to the HCII–HCIII megafloreal transition (see Johnson, this volume). Across the palynological K-T boundary transition, samples for carbon isotope analysis were taken at 3–10 cm intervals. Throughout the Hell Creek portion of the section, we sampled at 1 m intervals except between the highest reported HCII floral locality and the lowest HCIII horizon, where we sampled at 50 cm intervals. In the Ludlow Member, we were again hampered by poor exposure and sampled primarily at organic-rich horizons.

Terry's Fort Union Dinosaur section (KRJ Section 9653 = Dean Pearson Section 128 = DJN-96-7, Slope County, North Dakota; see also Nichols and Johnson, this volume) was of particular interest because a partially articulated ceratopsian frill, horn, and rib pieces were preserved in this section at +80 cm (KRJ Section 96530), which is stratigraphically within the Fort Union Formation. Precise positioning of the chronostratigraphic K-T boundary is particularly important in this section because many workers in the dinosaur-rich lands of northeastern Montana, eastern Wyoming, and southern Canada have noted a 2–3 m gap between the highest stratigraphic occurrence of dinosaur remains and the chronostratigraphic K-T boundary (Archibald, 1982). This observation led some to hypothesize that dinosaurs may have been extirpated in this region before the terminal Cretaceous catastrophe (Archibald, 1987). We sampled at 1 cm intervals through the exposed fossil-bearing section.

New Facet Boundary (KRJ Section 9847, Slope County, North Dakota; see also Nichols and Johnson, this volume) was newly opened during the 1998 field season. We sampled at 10 cm intervals across the Hell Creek–Ludlow contact to determine whether and where the chemostratigraphic K-T boundary was preserved.

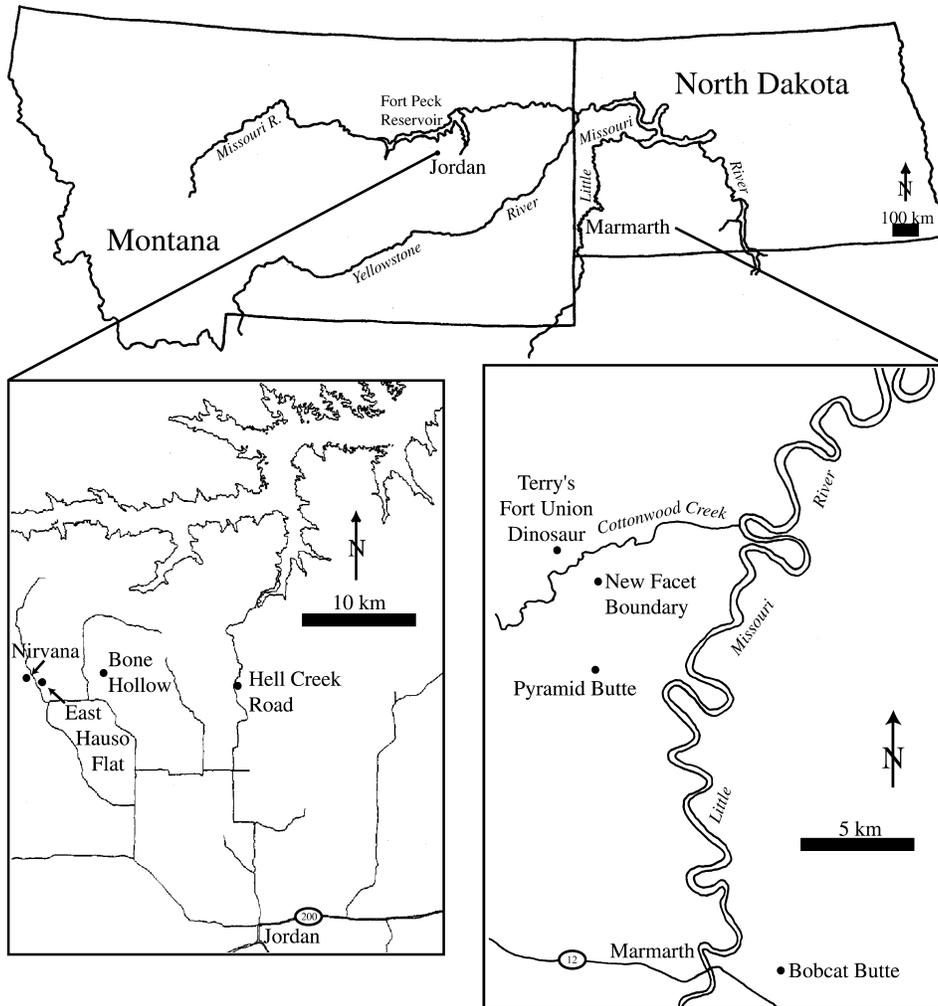


Figure 1. Location of sections discussed in this paper.

CARBON ISOTOPE VARIATION IN C₃ PLANTS

Plant vital effects

A variety of plant vital effects influence the carbon isotope composition of C₃ plant tissue (see related discussion in Arens et al., 2000) and could, therefore, influence chemostratigraphic interpretation of secular variation in terrestrial organic carbon isotopes. For example, extreme light limitation can drive plant tissue $\delta^{13}\text{C}$ values as much as 5‰–6‰ lower than those of plants growing in full sun (Ehleringer et al., 1986; Zimmerman and Ehleringer, 1990). Water stress can push plant tissue to higher values, with $\delta^{13}\text{C}$ values in desert plants up to 6‰ greater than those in mesic environments (Ehleringer et al., 1992; Ismail et al., 1994; Johnson and Tieszen, 1993; Tieszen, 1991). Nutrient stress (Toft et al., 1989) and low temperature (Tieszen, 1991) can decrease leaf $\delta^{13}\text{C}$ values by as much as 4‰. Juvenile woody plants tend to have lower $\delta^{13}\text{C}$ values than do reproductive-age plants (Donovan and Ehleringer, 1991). Brooks et al. (1997a) found that growth form influenced plant

$\delta^{13}\text{C}$ values, particularly when different growth forms encounter differing proportions of respired CO₂ (Broadmeadow et al., 1992; Brooks et al., 1997b; Medina et al., 1986). However, analysis of a larger data set failed to show differences in $\delta^{13}\text{C}$ value among growth forms (Arens et al., 2000). Some lineage effects emerge. The average isotope composition of conifers is about 1‰ greater than that of angiosperm trees and herbs, which have similar values (Arens et al., 2000). Among woody angiosperms, deciduous forms have lower average $\delta^{13}\text{C}$ values (<1‰) than evergreen taxa (Marshall and Zhang, 1994). However, this effect may also be related to source CO₂ because all of the carbon used to build deciduous leaves in the mid latitudes (where much of the relevant research was conducted) comes from the early spring atmosphere, which has a $\delta^{13}\text{C}$ value about 1‰ lower than the annual average (Ciais et al., 1995). The annual average atmospheric $\delta^{13}\text{C}$ value characterizes evergreen leaves, which build tissue throughout the growing season (this effect is discussed in more detail in Arens et al., 2000). C₃ grasses have average $\delta^{13}\text{C}$ values up to 2‰ greater than those of herbaceous dicots (Arens et al., 2000; Mooney et al., 1989).

However, Poaceae (grasses) does not appear in the fossil record until the late Paleocene (Crepet and Feldman, 1991).

Clearly, variation in carbon isotope composition because of plant vital effects may be large for individual plants under extreme conditions. However, organic carbon from many individual plants of different species is likely to be averaged as organic material enters the sedimentary record. Therefore, single-species data or that from plants under extreme conditions gives only partial insight into the vital effect variation that might be expected in the sedimentary record. A more realistic approach considers ranges of variation in carbon isotopic composition of many species within a single community. To quantify within-community variation in plant $\delta^{13}\text{C}$ value, we analyzed data from two modern communities in which $\delta^{13}\text{C}$ values were measured on a variety of species. Mooney et al. (1989) measured $\delta^{13}\text{C}$ values of leaf tissue from 18 species of angiosperm tree and shrub from a tropical woodland in the state of Jalisco, western Mexico. Both deciduous and evergreen species were represented in their sample and the ecosystem experienced seasonal drought. Smedley et al. (1991) studied 36 species of dicot herbs in a temperate grassland in Utah. These species showed $\delta^{13}\text{C}$ value variation due to lineage, life history (annual versus perennial) and phenology (Smedley et al., 1991). This community also experienced seasonal drought. We chose these communities because they differed in basic structure and composition and, therefore, encompass a wide range of possible vital effects.

The tropical woodland community displayed a mean $\delta^{13}\text{C}$ value of -27.3‰ with a standard deviation of 0.8‰ (Fig. 2, data from Mooney et al., 1989); the temperate grassland community had an average $\delta^{13}\text{C}$ value of -26.8‰ with a standard deviation of 0.8‰ (data from Smedley et al., 1991). An F-test for equality of variances failed to reject the null hypothesis that the variances in these samples are equal ($p = 0.46$). Residuals for both samples were normally distributed (Lilliefors $p = 0.19$ for the grassland community and $p = 0.15$ for the tropical woodland). Analysis of variance showed that the mean $\delta^{13}\text{C}$ values for these communities were not significantly different ($p = 0.06$). Given the similarity in both mean and standard deviation for these different plant communities, we propose the sample standard deviation ($\pm 0.8\text{‰}$) as a reasonable figure for expected vital effect variation within a plant community. This degree of variation might also be expected in $\delta^{13}\text{C}$ value for plant-derived organic carbon in the sedimentary record. Therefore, in chemostratigraphic interpretations, we disregard variation in the carbon isotope curve of $< \sim 1\text{‰}$, as this could be reasonably attributed to vital effects within the ancient plant community.

Choice of biogeochemical substrate

In using variation in carbon isotopic composition of C_3 plants to chart variation in the $\delta^{13}\text{C}$ value of paleoatmospheric CO_2 , we must consider what organic carbon substrate to mea-

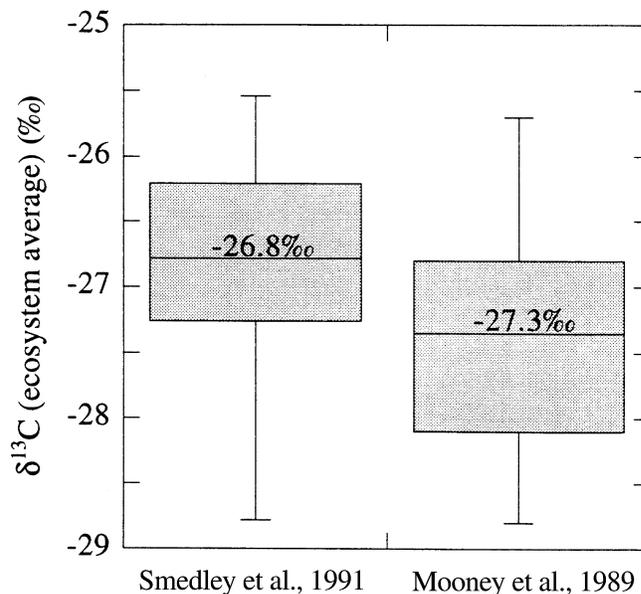


Figure 2. A comparison of whole-leaf $\delta^{13}\text{C}$ measurements from 18 species of tree and shrub from a tropical woodland in the state of Jalisco, western Mexico (Mooney et al., 1989), and 36 species of dicot angiosperms from a temperate grassland community in Utah (Smedley et al., 1991). The mean $\delta^{13}\text{C}$ value for each community is represented by the line bisecting each box, magnitude of the mean is noted above this line; one standard deviation is included within the gray boxes, this value is 0.8‰ for both communities; two standard deviations are denoted by brackets. An analysis of variance showed that these means do not differ significantly.

sure. Whole-fossil carbon isotopic values may be difficult to interpret because early diagenetic changes may alter the carbon isotope signal up to 4‰ (Benner et al., 1987). A variety of plant-derived macromolecules can be isolated from sedimentary rocks, and several have been proposed as useful substrates. Isolated cellulose, which ranges from 0.3‰ to 3.0‰ greater than whole tissue $\delta^{13}\text{C}$ values (Marino and McElroy, 1991; O'Leary, 1988), is preferred by many workers studying Pleistocene and Holocene materials (eg., Dupouey et al., 1993; Marino et al., 1992; Van de Water et al., 1994). Lignin ranges from 4.2‰ to 2.0‰ less than whole plant tissue values (Benner et al., 1987). Lipids, although popular for biogeochemical analysis, can be highly isotopically variable, depending on the tissue and synthetic pathway from which they were derived. In *Fagus sylvatica*, the $\delta^{13}\text{C}$ value of $n\text{-C}_{27}$ alkane varied by more than 1‰ with leaf orientation (Lockheart et al., 1997). For sitosterol, the difference between north- and south-facing leaves was 3‰ . In *Quercus robur* sun leaves, the $\delta^{13}\text{C}$ value of $n\text{-C}_{27}$ alkane and $n\text{-C}_{29}$ alkane was $\sim 6\text{‰}$ less than whole leaf tissue values. Values from shade leaves varied between 4‰ and 6‰ less than those of bulk tissue (Lockheart et al., 1997). Individual leaves also showed seasonal variation of an additional 3‰ in $n\text{-C}_{27}$ and $n\text{-C}_{29}$ alkanes (Lockheart et al., 1997). Bulk tissue was less variable in all cases (Lockheart et al., 1997). Despite this eco-

phenotypic variability, a -1.4‰ to -1.8‰ shift in $n\text{-C}_{29}$ alkane has been interpreted as the K-T boundary in Spain (Ari-nobu et al., 1999).

We have chosen to evaluate isolated cuticle because it appears to directly parallel whole leaf carbon isotope values with low variation ($<1\text{‰}$, Upchurch et al., 1997). However, cuticle cannot be isolated from all sediment samples. Therefore, we must ask how well bulk organic carbon in sediment reflects the isotope composition of plants growing on the landscape. Medina et al. (1986) noted that the average $\delta^{13}\text{C}$ value of forest litter equaled that of canopy leaves, which were consistently 2‰ less than the associated soil organic matter. These data suggest that organic matter in soil has a consistent isotope relationship with the associated vegetation. To test this hypothesis, in the rock record, we regressed bulk sedimentary organic carbon isotope values onto the $\delta^{13}\text{C}$ value of a mixed sample of plant cuticle isolated from the same sediment ($n = 68$). We recovered a significant relationship between the $\delta^{13}\text{C}$ values of bulk sedimentary organic carbon and those of isolated cuticle (Fig. 3, $R^2 = 0.69$, $p < 0.001$). The residual average for this

analysis was $\pm 0.4\text{‰}$. Residuals for this analysis include variation due to both plant vital effects and taphonomic variation. A slope approximately equal to one was statistically significant ($p < 0.001$), but the associated y-intercept (~ -2) was not ($p = 0.2$). However, the y-intercept was similar to the difference between the carbon isotope composition of canopy leaf tissue and that of soil organic material reported by Medina et al. (1986), suggesting that the actualistic relationship holds for ancient sediment. This shows that the bulk carbon isotope signature in sediments was related to the isotope composition of cuticle and chemically related decay-resistant materials, rather than by within-sediment carbon recycling or isotopically different compounds as previously suggested (Brooks et al., 1997b; Ke and Lin, 1997; Wedin et al., 1995). We further conclude that bulk carbon isotope measurements reflect the average signal of the ancient vegetation and, therefore, the best approximation of the paleoatmosphere. Therefore, bulk sedimentary organic carbon can be used for chemostratigraphic reconstruction in terrestrial rocks. However, we have consistently avoided sampling well-developed paleosols in the field to avoid data

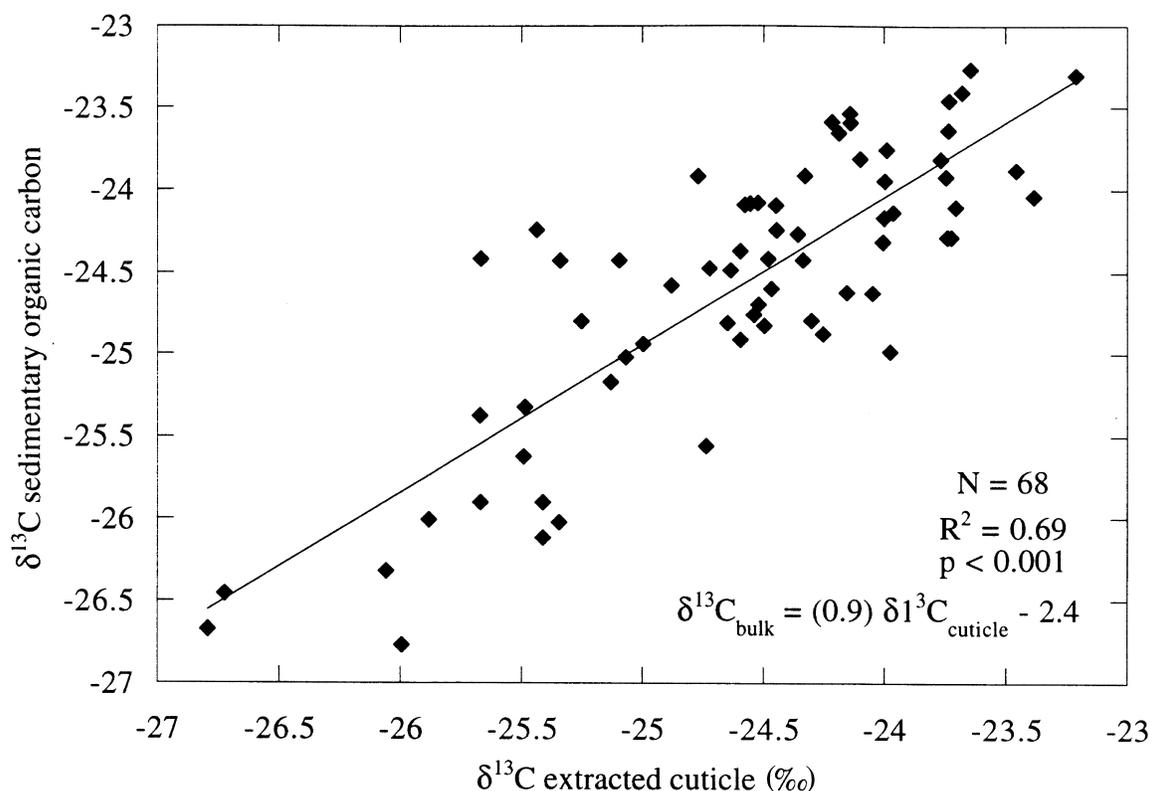


Figure 3. Regression of $\delta^{13}\text{C}$ value of sedimentary organic material onto $\delta^{13}\text{C}$ value of a mixed sample of vascular plant cuticle isolated from the same sediment. The relationship was statistically significant ($R^2 = 0.69$, $p < 0.001$). Slope ($p < 0.001$) was approximately equal to one, suggesting that bulk sedimentary organic material varied in tandem with associated plant cuticle. The regression intercept was not statistically significant ($p = 0.2$), but was approximately equal to 2‰ , which was the difference between the carbon isotope composition of soil organic material and canopy leaves observed in ever-wet forests (Medina et al., 1986). This result suggests that bulk sedimentary organic material reliably parallels the isotope composition of plant tissue and thus can be used to evaluate the isotope composition of paleoatmospheric CO_2 .

biased by extensive within-soil carbon recycling (Wedin et al., 1995).

TAPHONOMIC CONSIDERATIONS

Depositional environment and rock type

As discussed above, water status influences the $\delta^{13}\text{C}$ value of C_3 plants (eg., Beerling and Woodward, 1995; Farquhar and Richards, 1984; Marshall and Zhang, 1994). Consequently, one might suspect depositional environment to bias the $\delta^{13}\text{C}$ value of sedimentary organic carbon. Specifically, plants growing in better-drained substrates might experience more water stress than those growing in wetlands. Several authors have also suggested that differential diagenetic oxidation of organic material could systematically alter carbon isotopic ratios in fossils and sediment (Bartley et al., 1998; Benner et al., 1987). To test whether organic content of sediment systematically affects carbon isotope composition, we regressed the $\delta^{13}\text{C}$ values of bulk organic carbon in sediments upon weight percent organic material from the same samples (Fig. 4, $n = 411$). Weight percent organic carbon variation roughly paralleled rock-type variation observed in the field, ranging from mudstone and sandstone samples with very low organic content to lignite and carbonaceous shale samples with higher carbon content. Weight percent

carbon content also serves as a crude proxy for facies variation in this context, with higher-carbon sediments corresponding to wetlands and low-carbon samples to better-drained substrates. Unlike the vital effect tests discussed above, this analysis is not designed to document a range of variation in carbon isotope composition because of facies variation. Because the isotope composition of the atmosphere is assumed to vary among samples in this analysis, the regression only tests whether there is a systematic relationship between $\delta^{13}\text{C}$ value and organic carbon content of the rock. Correlation between carbon isotope composition and percent organic carbon content ($R = 0.07$, $p = 0.1$) was not statistically significant. Bartley et al. (1998) suggested that when diagenetic alteration of the carbon isotopic signal is minimal, $\delta^{13}\text{C}$ values should be uncorrelated with total organic carbon values as in Figure 4. This result demonstrates that organic carbon content of terrestrial rocks does not systematically bias $\delta^{13}\text{C}$ values. We can, therefore, compare $\delta^{13}\text{C}$ values across this range of terrestrial depositional environments without introducing systematic bias.

Another way to approach the question of facies variation is to examine rock type. To detect a relationship between rock type and carbon isotope value, we coded each $\delta^{13}\text{C}$ value for rock type (lignite, carbonaceous shale, claystone-mudstone, siltstone, and sandstone) based on field descriptions by Arens. Since each rock type was represented by a different number of

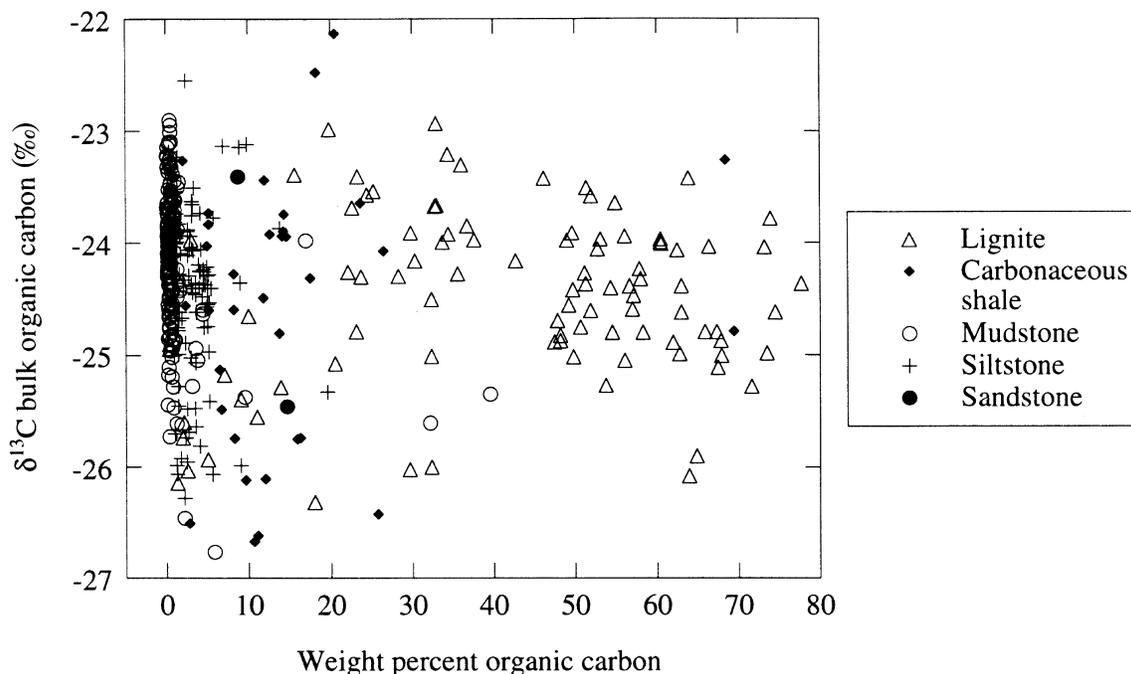


Figure 4. Regression of $\delta^{13}\text{C}$ values of bulk sedimentary organic material onto percent organic carbon content by weight. Data from several Hell Creek and Fort Union Formation localities in eastern Montana and southwestern North Dakota. There was no significant correlation ($R = 0.07$, $p = 0.1$) between these variables, suggesting that carbon isotope values of sedimentary organic material are not systematically influenced by the amount of organic carbon preserved. As weight-percent organic content was a proxy for facies variation, this result also shows that $\delta^{13}\text{C}$ values are not biased by depositional environment.

measurements, we randomly selected a balanced sample ($n = 39$) of lignite, carbonaceous shale, claystone-mudstone and siltstone values for analysis. We chose the nonparametric Kruskal-Wallis test for this analysis because carbonaceous shale failed the homogeneity of variance requirement for parametric ANOVA. 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, siltstone, and sandstone yielded similar results ($p = 0.160$). We conclude that carbon isotopic value does not vary systematically with rock type (see further discussion in Arens and Jahren, 2000). Together, these analyses show that there is no significant or systematic taphonomic bias in carbon isotopic measurements of terrestrial organic material. Therefore, secular variation observed in $\delta^{13}\text{C}$ values can be attributed primarily to the carbon isotopic signature of plant-derived organic material and thus the paleoatmosphere.

Lateral variation in carbon isotope value

If the carbon isotopic composition of C_3 plant-derived sedimentary organic material faithfully parallels the carbon isotopic composition of paleoatmospheric CO_2 , one would expect that samples taken from the same chronostratigraphic horizon in different locations would have similar absolute magnitudes. To test this hypothesis, we evaluated four K-T boundary-crossing sections, three in northeastern Montana and one in southwestern North Dakota (Table 1). In each of these sections, the K-T boundary is identified by an iridium-bearing boundary clay that is interpreted as the fallout of the terminal Cretaceous bolide impact (Alvarez, 1983). The assumption of temporal co-

incidence of the impact clay horizon at these localities hinges on the iridium-enriched clay representing a single, instantaneous event (Alvarez et al., 1980; Alvarez et al., 1995). Table 1 shows that the $\delta^{13}\text{C}$ values in the sample underlying the iridium-enriched impact clay are similar, regardless of rock type and location. Even across considerable distance, the absolute magnitude of $\delta^{13}\text{C}$ measurements from the same chronostratigraphic horizon correspond. This suggests that despite all of the possible plant vital effect and taphonomic variability, a meaningful chemostratigraphic signal emerges from these data.

Taken together, these analyses show that bulk sedimentary organic $\delta^{13}\text{C}$ values retain the carbon isotope composition of paleoatmospheric CO_2 . Therefore, carbon isotope cycles, trends, and shifts greater than that associated with vital effects (0.8‰ discussed above) can be used to correlate between terrestrial sections. Because plant-derived carbon is ubiquitous in the floodplain sediments that characterize terrestrial deposition in the Hell Creek and Fort Union Formations, this method will ostensibly work for most fossil-bearing sections. Because the method does not rely on biostratigraphic distribution of fossil forms, it offers an independent correlation, including placement of the K-T boundary, which will enhance the inference drawn from fossil distributions in space and time.

RESULTS

Pyramid Butte section

Because the Pyramid Butte section contains an iridium-bearing K-T boundary clay in addition to biostratigraphic markers (Johnson et al., 1989), it served as the reference for regional carbon isotope correlation. Figure 5 suggests cyclic variation of -24.0% to -25.5% in $\delta^{13}\text{C}$ value of bulk sedimentary organic material between -1500 cm and -200 cm (KRJ Section 86161, -365 cm to -1235 cm). We have observed similar quasiperiodic variation at the Bone Hollow locality (Los Angeles County Museum locality 7245, Garfield County, Montana, Trumbo Ranch 7.5' Quadrangle, sec. 34, T. 21 N., R. 37 E.) in northeastern Montana (Fig. 1, Arens and Jahren, unpublished data). Cyclic variation in $\delta^{13}\text{C}$ value between -23.5% and -25% continues through the lignite at the base of the Fort Union Formation (Fig. 6, KRJ Section 86161, 0 cm to $+90$ cm). This lignite immediately underlies the iridium-bearing K-T boundary clay (KRJ Section 86161, $+95$ cm). $\delta^{13}\text{C}$ values from the 10 cm immediately below the K-T boundary clay (KRJ Section 86161, $+90$ cm to $+80$ cm) range from -23.5% at $+80$ cm to -24.5% at $+90$ cm.

At Pyramid Butte, the iridium anomaly and palynological K-T boundary co-occur at $+95$ cm (Table 1, Fig. 6, Nichols and Johnson, this volume; Johnson et al., 1989). Carbon isotope values in the centimeter below and centimeter above the iridium-bearing boundary clay were -24.5% (Table 1) and -24.4% , respectively. Similar $\delta^{13}\text{C}$ values immediately above and below a K-T iridium anomaly have been observed at the

TABLE 1. CARBON ISOTOPE DATA FROM SAMPLES TAKEN IMMEDIATELY STRATIGRAPHICALLY BELOW IRIIDIUM-ENRICHED CRETACEOUS-TERTIARY BOUNDARY IMPACT CLAYS

Locality	Iridium (ppb)	$\delta^{13}\text{C}$ organic (‰)	Rock type
Hell Creek Road, Montana	0.57	-24.4	Mudstone
East Hauso Flat, Montana	2.22 and 11.7	-24.3	Carbonaceous shale
Nirvana, Montana	2.22 and 11.7	-24.6	Silty mudstone
Pyramid Butte, North Dakota	0.72	-24.5	Lignite

Note: The iridium anomaly at Hell Creek Road, Montana (Baadsgaard et al., 1988) is coincident with the palynological Cretaceous-Tertiary (K-T) boundary (Hotton, 1988). Iridium and palynological K-T boundaries also co-occur at Pyramid Butte, North Dakota (Johnson et al., 1989). The Nirvana and East Hauso Flats sections are ~ 0.5 and 1 km, respectively, east of the section discussed by Alvarez (1983, Iridium Hill, Ir = 2.2 ppb) and that of Smit and Van der Kaars (1984, Herpijunk Promontory, Ir = 11.7 ppb), where both iridium and palynological K-T boundaries have been documented. Strata are continuously exposed over this distance and the impact clay was directly traced laterally from the previous sample sites to the localities discussed here. Rock type descriptions were made in the field.

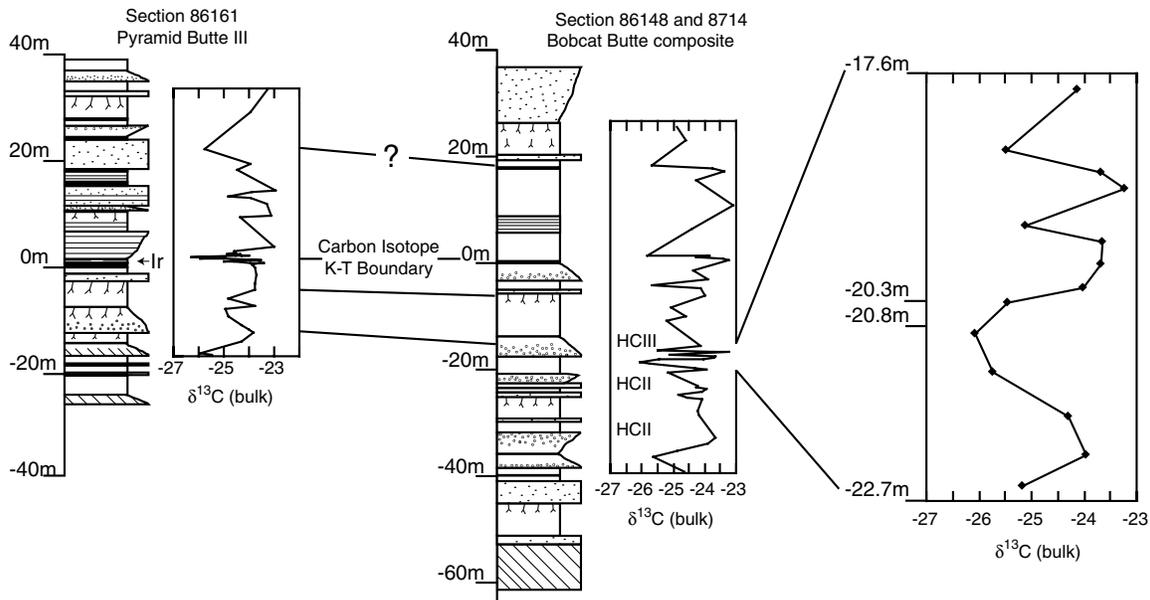


Figure 5. Carbon isotope stratigraphy for the Pyramid Butte and Bobcat Butte sections. Pyramid Butte (KRJ Section 86161), Slope County, North Dakota, Pretty Butte 7.5' Quadrangle, Hell Creek and Fort Union Formations. Stratigraphic position of the iridium anomaly in this section is noted by "Ir". The section was measured on 23 July 1987 by K.R. Johnson and R. Myers; sediment samples for carbon isotope analysis were taken on 23 and 25 July 1998 by N.C.A., H. Pogrel and A. Thompson. Top of the section 650 m S., 1000 m E. of northwest corner sec. 15, T. 134 N., R. 106 W. (UTM 5140888 m N., 578673 m E., Zone 13); base of the section 800 m S., 1020 m E. of the northwest corner sec. 15, T. 134 N., R. 106 W. (UTM 5140744 m N., 578697 m E., Zone 13). Graphical logs correspond to KRJ Section 86161. Bobcat Butte section (KRJ Section 86148), Slope County, North Dakota, Marmarth 7.5' Quadrangle, Hell Creek and Fort Union Formations. The section was measured on 17 July 1987 by K.R. Johnson and R. Myers; sediment samples for carbon isotope analysis were taken on 24, 25, and 27 July 1998 by N.C.A., H. Pogrel and A. Thompson. Top of the section 1610 m E. of the northwest corner sec. 28, T. 133 N., R. 105 W. (UTM 5127261 m N., 586495 m E., Zone 13). Base of the section 1520 m S., 550 m E. of northwest corner sec. 28, T. 133 N., R. 105 W. (UTM 512717 m N., 586287 m E., Zone 13). HCII and HCII megafloral zones are discussed by Johnson and Hickey (1990). Graphical logs correspond to KRJ Section 86148. The carbon isotope K-T boundary and several other chemostratigraphic tie point are noted with lines connecting the sections.

Hell Creek Road locality (Fig. 1, Garfield County, Montana, Maloney Hill 7.5' Quadrangle, sec. 1, T. 20 N., R. 37 E.; Lerbekmo, 1985; Swisher et al., 1993) and the nearby East Hauso Flat and Nirvana localities (Fig. 1, Garfield County, Montana, Hell Hollow 7.5' Quadrangle, sec. 36, T. 21 N., R. 35 E.; Smit and Van der Kaars, 1984). We interpret $\delta^{13}\text{C}$ values of -24.4‰ immediately above the iridium-bearing clay as the carbon isotope signature of vegetation destroyed by impact-related disturbance. Because these plants fixed their carbon under the latest Cretaceous atmosphere, their tissue retains the latest Cretaceous atmospheric signature. However, samples taken at +104 cm (10 cm above the impact clay) have $\delta^{13}\text{C}$ values decreasing to -25.9‰ and -26.3‰ at +124 cm (Fig. 6). The absolute value of the K-T carbon isotope shift at Pyramid Butte was similar to those observed at other iridium-bearing K-T boundary sections in Montana, which range from -25.9‰ at the Hell Creek Road locality to -26.8‰ at East Hauso Flat. The similarity of the carbon isotope absolute magnitudes and the stratigraphic relationship with the K-T boundary iridium

anomaly suggest that we are observing the same geochemical signal in all three localities.

At Pyramid Butte, $\delta^{13}\text{C}$ values recovered to -24.4‰ at +150 cm (55 cm above the K-T impact clay). In the Hell Creek Road section in northeastern Montana, where we have fine time-stratigraphic control, $\delta^{13}\text{C}$ values recovered from a K-T shift within 22 cm above the impact clay—at least 30 cm below the 65.00 ± 0.05 million year (Swisher et al., 1993) volcanic ash preserved in this section. Therefore, carbon isotope recovery can be radiometrically bracketed between 65.16 ± 0.04 Ma (the iridium-bearing impact bed) and 65.00 ± 0.05 Ma volcanic ash (Swisher et al., 1993)—a maximum interval of 250 k.y. Thus the carbon isotope shift and recovery observed in these terrestrial sections was not only similar in magnitude, but also in duration to that observed in the "early recovery" phase of the surface ocean signature (D'Hondt et al., 1998; Hsü et al., 1982; Keller and Lindinger, 1989). If one accepts the correlation between the carbon isotope shifts observed at Pyramid Butte, North Dakota, and Hell Creek Road, Montana, a similar

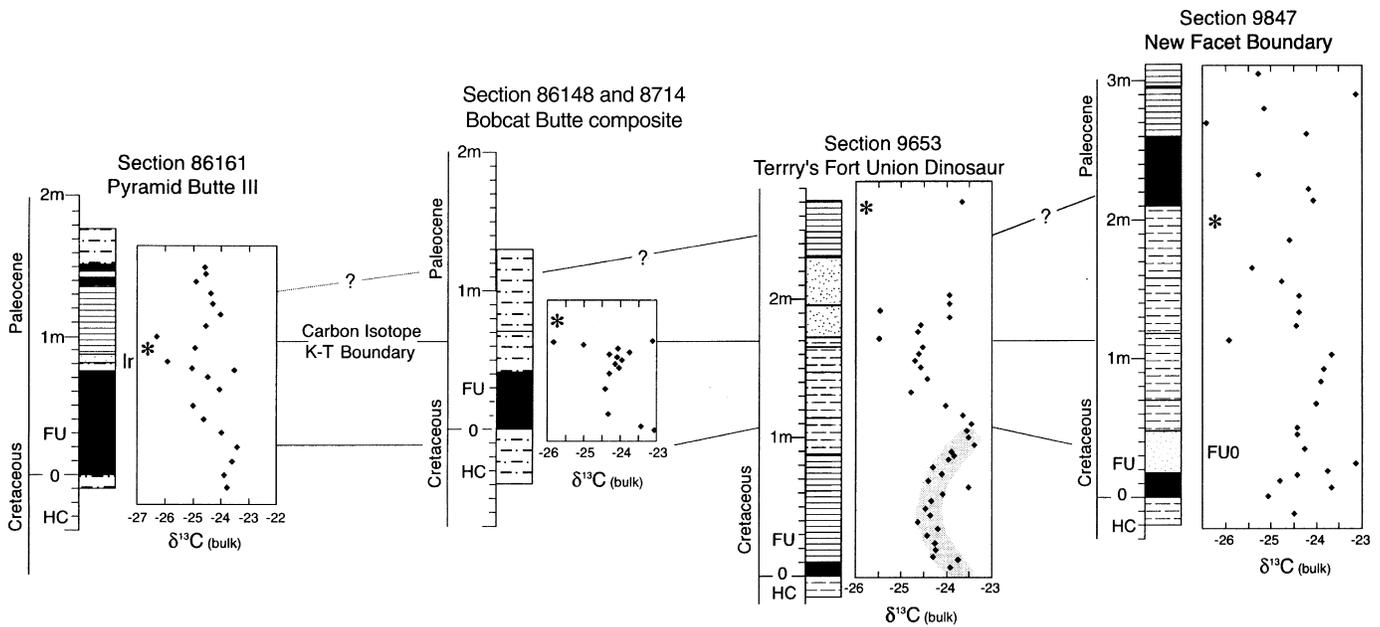


Figure 6. Detailed carbon isotope stratigraphy for the interval surrounding the Hell Creek–Fort Union formational contact and K–T boundary for the four sections analyzed in this study. The carbon isotope K–T boundary and several possible chemostratigraphic tie points are noted with lines connecting the sections. The paleontological K–T boundary is denoted by asterisk (*). Stratigraphic position of the iridium anomaly in the Pyramid Butte (locality information same as Fig. 5) section is noted by “Ir”. Bobcat Butte locality information same as Figure 5. Terry’s Fort Union Dinosaur section (KRJ Section 9653), Slope County, North Dakota, Badland Draw 7.5’ Quadrangle, Hell Creek and Fort Union Formations. The section was measured on 16 October 1996 by K.R. Johnson, D. and K. Nichols, D. Pearson and S. Bowman; sediment samples for carbon isotope analysis were taken on 26 July by N.C.A., K.R. Johnson, T. Farnham, H. Pogrel and A. Thompson. Section located SE ¼ NW ¼ NW ¼ sec. 33, T. 135 N., R. 106 W. (UTM 5146220 m N., 756179 m E., Zone 13). Graphical log corresponds to KRJ Section 9653. Gray shading denotes the range of carbon isotope variation (0.6‰) observed in detailed sampling through this portion of the section. New Facet Butte Boundary section (KRJ Section 9847), Slope County, North Dakota, Pretty Butte 7.5’ Quadrangle, Hell Creek and Fort Union Formations. The section was measured on 25 July 1998 by K.R. Johnson and T. Farnham; sediment samples for carbon isotope analysis were taken on 25 July 1998 by N.C.A. and A. Thompson. Section located sec. 34, T. 135 N., R. 106 W. (UTM 5145400 m N., 578100 m E., Zone 13). Graphical log corresponds to KRJ Section 9847.

time scale can be inferred for terrestrial sedimentation at Pyramid Butte.

Bobcat Butte section

The long section of uppermost Cretaceous sediments at Bobcat Butte permits a direct tie between the carbon isotope stratigraphy in the Hell Creek Formation and megafloreal zone transitions recognized by Johnson and Hickey (1990). Figure 5 shows bulk organic sedimentary $\delta^{13}\text{C}$ values of -25.5‰ at -4020 (KRJ Section 86148), with a trend toward higher values upsection. Between -2267 and -1756 cm (KRJ Section 86148), our data showed an interval of fluctuation in carbon isotope signature. This fluctuation (Fig. 5) is characterized by a positive shift at -2090 cm (-25.2‰ to -23.9‰), followed by a negative shift at -2050 cm (to -26.0‰), a positive shift at -2000 cm (to -24.3‰), a negative shift at -1950 cm (to -25.5‰), a positive shift at -1900 cm (to -23.7‰), a negative shift at -1800 cm (to -25.1‰), and a positive shift at -1750 cm (to -23.2‰). These shifts range in magnitude from 1.2‰ to 2.1‰ , exceeding the variation that might be expected

because of vital effects alone. We observed no indications of unconformities or depositional hiatuses (e.g., erosive contacts, well-developed paleosols, sedimentary lags, concretion horizons, etc.) to suggest extensive missing time. We also do not recognize a significant facies change in this interval; rocks in this interval were buff to light brown siltstone or sandy siltstone. Therefore, we interpret this interval as a geologically brief period of fluctuation in the carbon isotope composition of atmospheric CO_2 . We have not observed this pattern in other sections sampled to date.

The interval of carbon isotope fluctuation in the uppermost Cretaceous at Bobcat Butte (-1756 to -2267 cm, KRJ Section 86148) could be driven by a variety of factors. Climate change or regression-related releases of methane hydrate from marine sedimentary reservoirs could drive a rapid decrease in atmospheric $\delta^{13}\text{C}$ values, followed by photosynthetic compensation. This mechanism explained the 2.5‰ variation reported in $\delta^{13}\text{C}$ data from the latest Paleocene (Dickens et al., 1995). Alternatively, an input of volcanic CO_2 could drive $\delta^{13}\text{C}$ values higher (Taylor, 1986), with biomass oxidation returning the atmosphere toward lower values. Climatic variation could also

drive carbon isotope fluctuation in the atmosphere by changing the size of the terrestrial biomass/soil carbon reservoir. For example, a shift of up to 1.5‰ has been reported between Pleistocene glacial and interglacial cycles (Marino et al., 1992). At this point, we lack additional lines of evidence needed to distinguish between these hypotheses. However, we note that the transition between HCII and HCIII floral zones (Johnson and Hickey, 1990) occurred at the carbon isotope instability interval (Fig. 5). Therefore, we hypothesize that factors in the physical environment may have promoted latest Cretaceous vegetation change. To test this hypothesis, we suggest a closer look at the stratigraphic interval between floral zones at other sites within the region, and a hunt for additional lines of evidence designed to falsify the various mechanisms of carbon isotope variation proposed above. What seems clear at this point is that upper Hell Creek time was not characterized by environmental stability, at least in this portion of the Western Interior region.

At -580 cm, we note a shift in $\delta^{13}\text{C}$ value from -24.1‰ to -25.7‰ to -23.9‰ . This shift is similar in magnitude to the K-T boundary shift, but occurs in dinosaur-bearing beds, which also contain Maastrichtian palynomorphs and megaflores (Nichols and Johnson, this volume). We also observed this latest Cretaceous shift in terrestrial sections at the Bone Hollow locality in Montana and in some high-resolution marine sections (Keller and Lindinger, 1989). This feature is not apparent at Pyramid Butte.

No iridium anomaly has been recognized in the Bobcat Butte section. However, a -2.1‰ carbon isotope shift (latest Cretaceous $\delta^{13}\text{C}$ value of -23.7‰ , overlain by a boundary shift $\delta^{13}\text{C}$ value of -25.8‰) at $+62$ cm (KRJ Section 86148) suggested that the carbon isotope K-T boundary can be recognized at the top of the carbonaceous shale overlying the basal lignite of the Fort Union Formation at this locality. As at Pyramid Butte, the palynological K-T boundary also overlies the basal Fort Union lignite bed ($+75$ cm, Table 1, see also Nichols and Johnson, this volume). Extensive weathering and poor exposure prohibited detailed sampling into the Fort Union Formation at this locality.

Terry's Fort Union Dinosaur section

This approximately 3 m section preserved a partially articulated frill, horn, and associated ribs of a ceratopsian dinosaur in a 78-cm-thick silty mudstone (Fig. 6, $+80$ cm, see also Pearson et al., this volume), which overlies the basal lignite bed of the Fort Union Formation. Although it would not be unusual for the palynological K-T boundary to occur in the Fort Union Formation (e.g., Pyramid Butte and Bobcat Butte Sections, see also Nichols and Johnson, this volume), the position of the chronostratigraphic boundary relative to the dinosaur remains is critical to discussions of the timing of dinosaur extirpation in the region.

Detailed sampling through the section revealed apparently cyclic variation between -24.5‰ and -23.5‰ (Fig. 6). This

variation was similar to that observed in the uppermost Hell Creek Formation at Pyramid Butte (KRJ Section 86161, -365 cm to -1235 cm), but at Terry's Fort Union Dinosaur, the pattern is condensed over a much shorter section (0 cm to $+165$ cm). Sampling density in the latter section also allows us to note a range of variation (in gray, Fig. 6) around the secular trend. This variation is $\sim 0.6\text{‰}$, which is slightly less than the range of variation associated with plant vital effects predicted both by the actualistic data set (Arens et al., 2000) and the modern ecosystem data sets discussed above. This observation suggests that although vital effects and taphonomic alteration produce some variation in bulk carbon isotopic measurements, variation $>1\text{‰}$ probably represents change in the isotope composition of the source CO_2 used to build ancient plant tissue. Therefore, shifts, trends, and cycles greater than this magnitude can be interpreted chemostratigraphically, as suggested by Williams et al. (1988).

We observed low $\delta^{13}\text{C}$ values, suggestive of the K-T carbon isotope shift, at $+170$ cm and $+190$ cm (KRJ 9653). These values (both -25.5‰) were higher than that observed at the Pyramid Butte K-T boundary (-26.3‰), but similar to that observed at Bobcat Butte (-25.8‰). Therefore, we place the carbon isotope K-T boundary ~ 1 m above the dinosaur remains. The palynological K-T boundary occurs at approximately $+260$ cm, ~ 180 cm above the dinosaur fossil.

Using the Pyramid Butte section (Fig. 5) as an interpretive template, we correlate $\delta^{13}\text{C}$ values of -24.0‰ and -24.4‰ at $+70$ cm at Pyramid Butte (KRJ Section 86161) with values of -24.4‰ to -24.6‰ + 130 cm to $+150$ cm at Terry's Fort Union Dinosaur (KRJ Section 9653). We also propose to correlate higher $\delta^{13}\text{C}$ values (-23.4‰ to -23.6‰) observed at 0 cm to $+10$ cm at Pyramid Butte (KRJ Section 86161) with similar values (-23.5‰ to -23.9‰) at $+100$ cm at Terry's Fort Union Dinosaur (KRJ Section 9653). The trend toward lower values between $+70$ cm and $+90$ cm at Terry's Fort Union Dinosaur is best correlated with a similar trend between -360 cm and 0 cm at Pyramid Butte (KRJ Section 86161). Low $\delta^{13}\text{C}$ values at -425 cm (KRJ Section 86161) at Pyramid Butte (-23.8‰) and $+62$ cm (KRJ Section 9653) at Terry's Fort Union Dinosaur (-23.5‰) may also be correlated. Therefore, the base of the Terry's Fort Union Dinosaur section is approximately correlated with -745 cm at Pyramid Butte (KRJ Section 86161). Approximately 9 m of sediment at Pyramid Butte are condensed into ~ 70 cm at Terry's Fort Union Dinosaur. This interpretation highlights the differential sedimentation rates common to terrestrial deposition. This example also illustrates another power of the chemostratigraphic curve to provide isochronous links between sections within a region.

New Facet Boundary section

Reliable chemostratigraphic correlation requires that several features of the carbon isotope curves be matched among sections. The New Facet Boundary section (Fig. 6) was difficult

to interpret because it lacked features characteristic of other curves where the K-T boundary could be diagnosed based on the presence of iridium and shocked mineral. At +110 cm, we recorded a carbon isotope value of -25.9‰ , which overlaid values of -23.7‰ and -23.9‰ (Fig. 6). Another low value (-26.4‰) was recorded at +270 cm. This value concluded a series of values that decrease up section through the stratigraphic interval +210 cm to +270 cm. This interval of decreased carbon isotope values is underlain by values no higher than -24.1‰ at $\sim +210$ cm to +220 cm. Palynological data placed the K-T boundary at +199 cm (see also Nichols and Johnson, this volume).

Two interpretations of these data are possible. First, the lower, -25.9‰ excursion may be interpreted as the K-T boundary carbon isotope anomaly. In this interpretation, carbon isotope values stratigraphically below the excursion are somewhat higher than those typically reported at iridium-bearing K-T boundary sites in the region (Table 1). This interpretation placed the palynological transition 90 cm above the carbon isotope anomaly.

Alternatively, the low value at +270 m might be interpreted as the K-T boundary carbon isotope excursion. We observe a similar pattern of steady decrease in carbon isotope values to a low K-T boundary value in the iridium-bearing Pyramid Butte section. However, at Pyramid Butte, the decrease occurs over a shorter stratigraphic interval (approx. 40 cm) compared to the New Facet Boundary section (approx. 70 cm). This could be explained by higher sedimentation rates at the New Facet Boundary site. Carbon isotope values underlying the +270 cm low are similar to those from iridium-bearing reference sections (Table 1). Correlating the +270 cm low with the K-T boundary carbon isotope excursion placed the palynological K-T boundary approximately 70 cm below the carbon isotope minimum attributed to the K-T boundary. This is similar to the pattern at Pyramid Butte, where the palynological K-T boundary occurs 10 cm below the minimum carbon isotope value (Table 2). Once again, higher sedimentation rate must be invoked to explain this difference. Under this interpretation, the -25.9‰ value at +110 cm is without parallel in any previ-

ously reported K-T boundary-crossing section. A carbon isotope value of -25.7‰ was recorded at -580 cm in the Bobcat Butte section. However, correlation of the -25.9‰ excursion at +110 cm at New Facet boundary with the -580 cm excursion at Bobcat Butte seems improbably. Such an interpretation would require significantly lower rates of sedimentation or a significant sedimentary hiatus at New Facet Butte. There is no evidence of significant disconformity in the New Facet Boundary section and low sedimentation rates are at odds with the palynological data. Furthermore, such a correlation would require much greater time transgression of the Hell Creek–Fort Union contact than has been reported previously in this region.

Despite our inability to develop a more definitive correlation for the New Facet Boundary section, both the carbon isotope and palynological data suggest that the FU0 flora (see also Johnson, this volume) at +30 cm (KRJ Section 9847) is of Cretaceous, rather than Paleocene, age. Difficulty in developing a carbon isotope correlation for New Facet Butte may also suggest that the section may be incomplete, as discussed below.

FEATURES OF THE CARBON ISOTOPE CURVES THAT CAN BE USED AS CHEMOSTRATIGRAPHIC TIE POINTS

Reliable chemostratigraphic correlation seldom hinges on single features. Greater confidence can be placed in correlations that match several tie points among curves. In Figures 5 and 6, we indicate several such points of correlation. In these sections, one of the most relevant tie points is the K-T boundary carbon isotope shift. As discussed above, this marker has been found in association with the K-T iridium anomaly in the global stratotype section at El Kef, Tunisia (Keller and Lindinger, 1989), and at other marine (Arinobu et al., 1999; Arthur et al., 1979; D'Hondt et al., 1998; Shackleton and Hall, 1984; Zachos et al., 1992) and terrestrial (Holmes and Bohor, 1994; Lomax et al., 1998; Schimmelmann and DeNiro, 1984) sections. Furthermore, this carbon isotope shift is backed by a widely accepted mechanism that explains the link between terrestrial and marine data for the interval (Arinobu et al., 1999; Arthur et al., 1985; Caldeira et al., 1990; D'Hondt et al., 1998; Ivany and Salawitch, 1993; Kasting et al., 1986; Kump, 1991; Stott and Kennett, 1989, 1990; Tappan, 1968; Williams et al., 1988; Wolbach et al., 1985; Zachos and Arthur, 1986; Zachos et al., 1989). Because the negative carbon isotope shift at the K-T boundary is not unique in the Hell Creek Formation, we have used other data, particularly biostratigraphy, to help distinguish the K-T negative shift from others of similar magnitude within the section (e.g., -580 cm at Bobcat Butte, Fig. 5, KRJ Section 86148). Williams et al. (1988) and Gale et al. (1993) recommend this combined approach, when possible.

A second potential chemostratigraphic tie point between the Pyramid Butte and Bobcat Butte sections (Fig. 5) is the shift to lower, then higher, isotopic values recognized in the HCIII floral zone (Johnson, this volume; Johnson and Hickey, 1990).

TABLE 2. COMPARISON OF THE STRATIGRAPHIC POSITION OF THE PALYNOLOGICAL AND CARBON ISOTOPE MARKERS FOR THE CRETACEOUS-TERTIARY (K-T) BOUNDARY AND THE STRATIGRAPHIC DIFFERENCE BETWEEN THE TWO

Locality	Carbon isotope K-T boundary (cm)	Palynological K-T boundary (cm)	Difference (cm)
Pyramid Butte	+105	+95	-10
Bobcat Butte	+62	+75	+13
Terry's Fort Union Dinosaur	+170	+261	+91
New Facet Boundary	+110	+184	+74

Note: Stratigraphic position given relative to the Hell Creek–Fort Union formational contact. See Nichols and Johnson (this volume) for further discussion of the palynologically based K-T boundary at these localities. The iridium anomaly in the Pyramid Butte section occurs at +95 cm.

In the Pyramid Butte section (KRJ Section 86161) this interval occurs between -400 cm and -900 cm, and between -950 cm and -1350 cm at Bobcat Butte (KRJ Section 86148). In this case, the pattern, absolute magnitude, and biostratigraphic position of these features suggest their correlation.

A third potentially useful chemostratigraphic feature is the interval of carbon isotope fluctuation identified between -1750 cm and -2270 cm in the Bobcat Butte section. This interval straddles the boundary between HCII and HCIII floral zones (Johnson, this volume; Johnson and Hickey, 1990) and is unusual in the Hell Creek Formation. We have not yet recognized this feature in any other terrestrial sections in North Dakota and Montana. However, zones of carbon isotopic instability have been reported in the uppermost Cretaceous marine record (Shackleton and Hall, 1984), suggesting that this is indeed a real feature of the carbon biogeochemical record.

A fourth chemostratigraphic feature that may provide correlative power is the trend in earliest Paleocene toward lower carbon isotopic values, which has been reported in a number of marine sections (D'Hondt et al., 1998; Shackleton and Hall, 1984). This trend is seen generally in both the Pyramid Butte (KRJ Section 86161) and Bobcat Butte (KRJ Section 86148) sections (Fig. 5). However, in both of these sections, poor exposure and extensive weathering did not permit adequate sampling to develop a detailed picture of these features in the sections sampled to date.

Together, these features of the carbon isotope curves provide several points of correlation for terrestrial sections in the region. Although much attention has been given to the K-T boundary negative carbon isotopic shift because of its importance for a variety of paleobiological questions, this feature cannot be interpreted alone. The K-T boundary shift is best interpreted in the overall context of secular variation in the carbon isotope signal in long and apparently complete sections such as that of Bobcat Butte and as part of a correlation scheme that includes many different lines of independent evidence.

COMPARISON OF K-T BOUNDARY MARKERS

A number of lines of evidence have been used to diagnose the K-T boundary. An iridium- and shocked mineral-bearing boundary clay is the preferred, but perhaps least common marker of the K-T boundary in terrestrial rocks. In the absence of an iridium anomaly, biostratigraphy is most commonly used. A carbon isotope chemostratigraphic approach can provide additional information that will aid the interpretation of terrestrial sections. We will briefly discuss the stratigraphic relationships between these different lines of evidence.

In the high-stratigraphic-resolution marine section at El Kef, Tunisia, the negative carbon isotope shift begins in the sample that also records the lowest horizon of elevated iridium abundance (Keller et al., 1995) and coincides with an increase in organic carbon deposition (Keller et al., 1995; Keller and Lindinger, 1989). At El Kef, the carbon isotope shift begins

2 cm below the level of peak iridium enrichments and reaches its lowest values ~ 50 cm above the maximum iridium enrichment. Unusually low $\delta^{13}\text{C}$ values are maintained for ~ 3 m above the iridium-defined boundary (Keller and Lindinger, 1989). Thus, although the marine carbon isotope shift begins coincident with iridium enrichment, the maximum shift does not co-occur with the iridium-enriched boundary clay. This observation is consistent with the mechanism proposed for the negative carbon isotope shift. If the marine shift begins with the cessation of marine productivity in the surface ocean because of postimpact environmental disturbance, it should co-occur with the iridium anomaly. However, we assert above that an additional input of $^{12}\text{CO}_2$ into the postimpact atmosphere is required to account for the observed magnitude of the negative carbon isotope shift. Whether because of terrestrial wildfires (Ivany and Salawitch, 1993; Wolbach et al., 1985) or other sources, a K-T related $^{12}\text{CO}_2$ input would have occurred after deposition of the impact-related fallout and thus the iridium anomaly. This lag explains why the maximum shift in $\delta^{13}\text{C}$ value is observed above the iridium anomaly at El Kef. Since our data from terrestrial sections record only atmospheric variation in carbon isotope composition, it is the postimpact maximum values, not the initial shift, that are recorded in terrestrial organic carbon. This mechanism predicts that the terrestrial K-T carbon isotope shift will always occur above an iridium-bearing boundary clay. We observe the predicted pattern at Pyramid Butte, where the minimum carbon isotope values for the shift occur 10 cm above the iridium-bearing boundary clay. The distance between the K-T boundary impact clay and minimum carbon isotope values will vary from section to section depending on depositional rate at a given site. However, it is important to note that we have correlated the lowest $\delta^{13}\text{C}$ values in Figures 5 and 6. While these lowest values are temporally coincident and thus, suitable for correlation tie points, they are not identical to the moment of impact, as recorded in the iridium-enriched boundary clay. At the Hell Creek Road locality in northeastern Montana, the terrestrial carbon isotope shift and recovery lasted significantly less than 250 k.y. Consequently, the maximum carbon isotope anomaly may still be a suitable marker for the K-T boundary for questions of biotic change when an iridium-bearing boundary clay is not preserved.

Pollen and spores have been extensively used to identify the K-T boundary in terrestrial sediments. Recognition of the palynological K-T boundary relies on the last appearance of some members of the *Wodehouseia spinata* assemblage (Nichols and Sweet, 1993). Tschudy and Tschudy (1986) stated that "The palynological Cretaceous/Tertiary boundary in the Western Interior of North America consistently occurs precisely at the horizon at which a boundary clay layer yielding anomalously high iridium and other elements is found" (p. 667). This coincidence, in fact, allowed Tschudy to successfully guide iridium sampling in the Raton Basin or New Mexico (Orth et al., 1981). However, the extinction of characteristic Cretaceous forms may occur over some stratigraphic distance. At Lance

Creek, the K-T boundary can be bracketed only within several meters using pollen and spores (Leffingwell, 1971), although a K-T iridium anomaly was subsequently identified (Bohor et al., 1987b). At Pyramid Butte, the last appearances of *Wodehouseia spinata* assemblage palynomorphs is distributed over approximately 50 cm bracketing the iridium-defined boundary (Johnson et al., 1989).

The mechanism underlying palynostratigraphic recognition of the K-T boundary is the extirpation or extinction of typical Cretaceous pollen and spore taxa due to impact-related disturbance (Nichols and Fleming, 1990; Smit and Van der Kaars, 1984; Tschudy et al., 1984). Consequently, we would expect the palynostratigraphic K-T boundary to be coincident with evidence of impact. Since the carbon isotope shift associated with the boundary should always occur stratigraphically above the impact clay because of a temporal lag between the impact and its effect on the atmosphere, the carbon isotope shift should similarly occur stratigraphically above the palynostratigraphic K-T boundary. In the four North Dakota sections studied to date, this occurs only at Pyramid Butte, where the palynological pick for the K-T boundary coincides with the iridium anomaly (Fig. 6, Table 2). In each of the other sections studied, the palynological pick for the K-T boundary (Nichols and Johnson, this volume) occurs from about 10 cm to nearly one meter above the minimum $\delta^{13}\text{C}$ value interpreted as the K-T carbon isotopic shift (Fig. 6, Table 2). Based on the few sections studied to date, we are reluctant to interpret this pattern; however, it should be noted that the last occurrence of several *Wodehouseia spinata* assemblage taxa also occurs stratigraphically above the carbon isotopic shift at Pyramid Butte (Johnson et al., 1989).

Beyond locating a stratigraphic horizon such as the K-T boundary or correlating among sections, another important application of carbon isotope stratigraphy in terrestrial sections may be the recognition of incomplete or paraconformable sections. Shackleton and Hall (1984) suggested this application for marine sediments. They proposed that for questions that required complete sections across the K-T boundary, the presence of the characteristic carbon isotope excursion offered the best check for section completeness (Shackleton and Hall, 1984). To illustrate this method, Figure 7 compares a complete hypothetical stratigraphic section with one in which the 2 m surrounding the boundary clay are missing. In both cases, a biostratigraphic transition would be recognized. However, comparison of the carbon isotope curves reveals a different pattern. In the incomplete section, the reversible carbon isotope shift associated with the biostratigraphic transition is missing. In its place, we see a non-reversible shift to larger $\delta^{13}\text{C}$ values. In a real-world example, we concluded above that the latest Cretaceous portion of the New Facet Boundary section may be incomplete, because its carbon isotope curve lacked features observed in other sections in the area. Because the relative completeness of fossil-bearing sections may be particularly important to the interpretation of the first and last occurrences that comprise our data

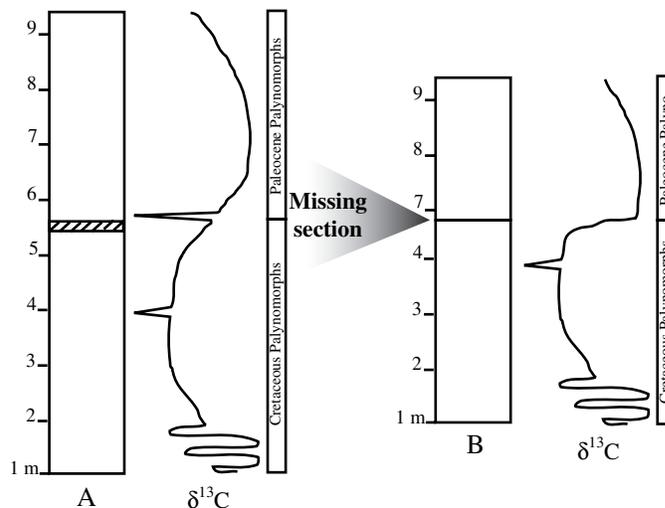


Figure 7. Carbon isotope stratigraphy can facilitate the recognition of incomplete sections. A: Hypothetical complete K-T boundary crossing section showing an iridium anomaly (hashed bar), carbon isotope curve and palynological distributions. B: The same section with the 2-m interval between 4 m and 6 m not represented by sediment. Note that vertical scales correspond in both sections. In both cases, a palynological boundary can be assigned in the correct stratigraphic position. However, differing shapes of the carbon isotope curves betray missing section.

for extinction and biotic change in the fossil record (Dingus, 1984), this application of carbon isotope stratigraphy may be the most significant.

SUMMARY AND CONCLUSIONS

When asking questions about the pattern and timing of extinction and biotic change across the K-T boundary, we are faced with several challenges: (1) diagnosing the chronostratigraphic boundary in the majority of terrestrial sections where an iridium anomaly is not preserved, (2) correlating in detail among fossil-bearing sections, and (3) recognizing incomplete and condensed sections. In the absence of impact-related iridium anomalies, lithostratigraphy, palynostratigraphy, and land mammals have been used extensively to place the K-T boundary. Although this approach works well, any biostratigraphic method will introduce circularity to questions of extinction and biotic change when the same fossils are used both to place extinction horizons and document the extinction itself. Furthermore, both palynomorphs and land mammals have been unable to provide detailed correlations within the Hell Creek and Fort Union Formations on a local scale. Biostratigraphy also commonly fails to recognize incomplete or condensed sections, a particular problem in terrestrial facies.

A -1.5% to -2.5% carbon isotope shift has been recognized above the K-T boundary in many marine sections. The marine $\delta^{13}\text{C}$ shift has been interpreted as a cessation of primary productivity in the surface ocean following the terminal-

Cretaceous impact, coupled with a change in the $\delta^{13}\text{C}$ value of atmospheric CO_2 , which then equilibrated with the surface ocean. Because vascular land plants sample atmospheric CO_2 directly during photosynthesis, their tissue preserves the carbon isotope signature of the atmosphere under which they grew (Arens et al., 2000). Therefore, an atmospheric carbon isotope shift could be used to correlate in terrestrial, as well as marine, sediments.

Using this approach, we have developed a carbon isotope stratigraphy for four fossil-bearing sections from the Marmarth region of Slope County, North Dakota. We use the Pyramid Butte section, in which an iridium anomaly marks the chronostratigraphic K-T boundary, as the basis for regional correlation. At Pyramid Butte, we observed a -2.8% carbon isotope shift beginning in the 5 cm that immediately overlies the iridium-enriched horizon and peaking ~ 30 cm above the iridium anomaly. Carbon isotope shifts also marked the K-T boundary at Bobcat Butte and Terry's Fort Union Dinosaur localities and tentatively, at New Facet Boundary. All of these sections lack iridium-enriched boundary clays.

At Pyramid Butte, we also observed secular variation in the carbon isotope composition of sedimentary organic material in the upper Hell Creek Formation, suggesting cyclic variation in the $\delta^{13}\text{C}$ value of the latest Cretaceous atmosphere. Similar patterns, which we used to correlate between these sections, were observed at both Bobcat Butte and Terry's Fort Union Dinosaur localities. These data suggest that some environmental instability characterized the latest Cretaceous in this region. This conclusion agreed with interpretations based on leaf size and leaf margin analysis that documented changing climate through this interval (Johnson and Hickey, 1990). The long section at Bobcat Butte also allowed us to tie changes in the Hell Creek megafloora directly to an interval of fluctuating carbon isotope values. At Bobcat Butte, the transition from HCII to HCIII leaf floras was marked by ~ 6 m within which $\delta^{13}\text{C}$ values fluctuated over a magnitude greater than might be expected by plant vital effect or taphonomic variation alone. The carbon isotope data document rapid and substantial change in global carbon cycling. Although our data do not permit us to pinpoint a particular mechanism for the observed fluctuation, such environmental disruption may have driven the observed change in vegetation. Further investigation will show whether this pattern stands in other sections and whether similar carbon isotope instability marks the HCI-HCII floristic change as well. It seems clear at this point, however, that the latest Cretaceous was not a time of environmental stability in this portion of the Western Interior region of North America. This result should be factored into any hypothesis of extinction and biotic change.

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