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A paleoclimatic and paleoatmospheric record from peatlands accumulating during the Cretaceous-Paleogene boundary event, Western Interior Basin, Canada

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ABSTRACT

Petrological and $\delta^{13}C$ analyses were undertaken on contiguous specimens of coal and intercalated minor organic-rich clastic sediments collected from coal seams spanning the Cretaceous-Paleogene boundary in the Alberta and Saskatchewan portions of the Western Interior Basin. The generally high smectite content of the coal suggests that the original mires were largely small, disconnected, and rheotrophic, readily receiving abundant waterborne detrital clastic material of largely volcanic origin. Nevertheless, using the distinctive claystone that marks the Cretaceous-Paleogene boundary as a regional datum, it is possible to correlate cycles in the vitrinite and inertinite composition of the coals over >500 km. Estimates of peat accumulation rates suggest that the cycles in vitrinite and inertinite composition represent regional, cyclic fluctuations in wildfire and oxidation of the peatlands and overlying canopy at a frequency of hundreds to thousands of years. The likely causes of these fluctuations were cyclic, regional-scale changes in temperature. The Cretaceous-Paleogene boundary event occurred early during a phase of gradually increasing temperature and/or decreasing rainfall, but peak wildfire and desiccation of peat occurred up to 14,000 yr later than the Cretaceous-Paleogene boundary, and the mires did not experience significant water stress in the immediate aftermath of the extinction event. A persistent, 1.5‰–3.0‰ negative $\delta^{13}C$ excursion occurs across the Cretaceous-Paleogene boundary, but it cannot be readily separated from four, further negative excursions later in the earliest Danian. The negative carbon isotope excursion linked to the Cretaceous-Paleogene boundary began a few hundred years before the event itself, and recovery occurred within 21 k.y., and possibly in as little as just a few thousand years, consistent with recently calibrated shallow-marine $\delta^{13}C$ records. Hence, the atmospheric and surface ocean carbon pools were coupled at this time. The absence of evidence for catastrophic change in the climatic regime at the time of the Cretaceous-Paleogene extinction in these mires supports the notion that the negative shift in atmospheric $\delta^{13}C$ was brought about by changes in the $\delta^{13}C$ composition of the surface ocean. This is consistent with the greater magnitude of extinction experienced by marine fauna relative to the terrestrial realm.

INTRODUCTION

The Cretaceous-Paleogene boundary extinction is considered the second most severe in Earth history in terms of its ecological impact, and the fifth most severe in terms of familial diversity loss (McGhee et al., 2004), and it affected both terrestrial and marine ecosystems (e.g., MacLeod et al., 1997; Nichols and Johnson, 2008; Archibald et al., 2010; Mitchell et al., 2012). Numerous hypotheses have been advanced as to the cause of the mass extinction, including the impact of a single, large extraterrestrial body (e.g., Alvarez et al., 1980; Smit, 1990; Hildebrand et al., 1991; Schulte et al., 2010), extensive volcanism for ~1 m.y. across the Cretaceous-Paleogene boundary (Courtillot et al., 1986; Duncan and Pyle, 1988; Courtillot and Fluteau, 2010; Gertsch et al., 2011), or the surpassing of biological thresholds brought about by the culmination of multiple nonindividually catastrophic factors (e.g., Archibald, 1996; MacLeod et al., 1997; Keller et al., 2003; Miller et al., 2010; Archibald et al., 2010; Mitchell et al., 2012; Tobin et al., 2014).

A transient 1‰–3‰ negative shift in the carbon isotope composition ($\delta^{13}C$) of planktonic carbonate across the Cretaceous-Paleogene boundary has been extensively documented in marine sections around the world (e.g., Thierstein and Berger, 1978; Hsü and McKenzie, 1985; Zachos et al., 1992; D'Hondt et al., 1998; Hart et al., 2004; Keller et al., 2003; Schulte et al., 2010). The fact that the shift is absent or lesser in magnitude in the contemporaneous benthos has been interpreted as evidence for the global reduction or shutdown of marine surface primary productivity (the so-called "Strangelove Ocean"; Hsü and McKenzie, 1985; Zachos et al., 1992) and/or a decrease in the flux of organic material from the surface to deep sea (D'Hondt et al., 1998; Alegret et al., 2012) at that time, resulting in the homogenization of the normal surface-to-depth positive-to-negative $\delta^{13}C$ gradient. A similar 1‰–3‰ negative shift in the $\delta^{13}C$ has also been identified across the Cretaceous-Paleogene boundary in organic carbon deposited in marine settings (e.g., Gilmour et al., 1987; Woolbach et al., 1990; Meyers and Simoneit, 1990; Hollander et al., 1993; Arinobu et al., 1999; Yamamoto et al., 2010) and fully terrestrial environments (Schimmelmann and DeNiro, 1984; Arens and Jahren, 2000; Beerling et al., 2001; Gardner and Gilmour, 2002; Maruoka et al., 2007; Therrien et al., 2007; Grandpre et al., 2013). Because preserved organic carbon of terrestrial origin records the isotopic composition of the paleoatmosphere (Marino and McElroy, 1991; Arens et al., 2000; Jahren et al., 2008), the similarity of the terrestrial and marine $\delta^{13}C$ records has been used to argue for coupling of the atmospheric and shallow-marine carbon reservoirs through the mass extinction event (Beerling et al., 2001).

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Grandpre et al. (2013), however, highlighted the occurrence of numerous other negative and positive shifts in $\delta^{13}C$ immediately before and after the Cretaceous-Paleogene boundary in their data and that of previous studies spanning the Cretaceous-Paleogene boundary in terrestrial successions in North America (i.e., Arens and Jahren, 2000; Maruoka et al., 2007; Therrien et al., 2007), and some of these excursions can be greater in magnitude than the shift associated with the Cretaceous-Paleogene boundary itself. Additionally, sampling resolution has been typically biased toward an ~1-m-thick zone immediately bracketing the Cretaceous-Paleogene boundary and decreases away from it. Outside this narrow zone, sampling spacing is often greater than the stratigraphic thickness that typically records the transient $\delta^{13}C$ excursion at the Cretaceous-Paleogene boundary, leading to the distinct possibility that other excursions of equal magnitude may have been missed in earlier studies. Altogether, the importance of the negative $\delta^{13}C$ excursion at the Cretaceous-Paleogene boundary in the terrestrial realm is likely to have been overstressed in the geological literature (Grandpre et al., 2013). Furthermore, Therrien et al. (2007) and Grandpre et al. (2013) concluded that it is difficult to correlate trends across multiple sections, and extract meaningful regional or global paleoatmospheric $\delta^{13}C$ time series from the terrestrial record because local, episodic deposition, incision, and reworking are characteristic of alluvial and fluvial sediments (MacLeod and Keller, 1991; Ager, 1993; Collinson, 1996).

This study documents systematic regional changes in the petrological and $\delta^{13}C$ composition of mire sediments (coal seams) at eight Cretaceous-Paleogene boundary sections from the Canadian portion of the Western Interior Basin. Only coal is targeted, because peatlands are considered to represent more continuous and in situ records of accumulation compared to terrestrial clastic-dominated sedimentary successions (Davies et al., 2006; Wadsworth et al., 2010), and the surface that separates the bases of some coals and their underlying sediment have been interpreted as hiatal (e.g., Aitken and Flint, 1996; Jerrett et al., 2011b) and likely to have been subject to postdepositional subaerial exposure, truncation, and diagenetic alteration (Gardner et al., 1988; Aitken and Flint, 1996; Driese and Ober, 2005) and postcompaction penetration by the roots of significantly later plants. The purpose of this study is to demonstrate that a regional record of paleoatmospheric $\delta^{13}C$ can be extracted from the terrestrial sedimentary record and to compare the record of terrestrial $\delta^{13}C$ with that from extensively described time-equivalent marine sections. The petrology of the coals is used to assess the degree to which selective degradation of plant material before and during peat formation could have influenced the $\delta^{13}C$ record, and as a proxy for the record of terrestrial wildfire and climate changes at the time of peat accumulation, by analogy with studies of Holocene peat (Blackford, 2000; Marlon et al., 2012).

GEOLOGICAL SETTING

The Late Jurassic to Eocene Western Interior Basin was an enormous composite foreland basin that, at its climax, spanned an east-west distance of >1000 km and a north-south distance of >5000 km from the Canadian Arctic to the Gulf of Mexico (Williams and Stelck, 1975; DeCelles, 2004). Until late Campanian times, subsidence of the basin was primarily caused by lithospheric flexure and dynamic subsidence adjacent and parallel to a zone of predominantly thin-skinned folding and thrusting, associated with the eastward subduction of the oceanic Farallon plate beneath the North American continent (the Sevier orogeny; Jordan, 1981; Beaumont, 1981; Porter et al., 1982; DeCelles, 2004). Late Jurassic to Late Cretaceous high eustatic sea levels (Vail et al., 1977; Haq et al., 1987) caused inundation of much of the Western Interior Basin by marine waters of the Western Interior Seaway, and connected the Arctic Ocean to the Gulf of Mexico (Kauffman and Caldwell, 1993; Robinson Roberts and Kirschbaum, 1995). From the late Campanian onwards, the Laramide orogeny (sensu Dickinson et al., 1988) sequentially segmented the contiguous foreland basin east of the Sevier deformation front into a succession of smaller structural basins flanked by basement-cored uplifts across a zone that comprises present-day northern New Mexico to southern Montana (Dickinson et al., 1988; DeCelles, 2004). The resulting decrease in accommodation space and increase in sediment supply in the Western Interior Basin, coupled with late Mesozoic eustatic sea-level fall (e.g., Haq et al., 1987), led to the sporadic and then permanent withdrawal of the seaway from the Western Interior Basin, from the Maastrichtian onward (Williams and Stelck, 1975; Tweto, 1980).

From central Montana northward, the foreland basin was not disrupted by basement-cored Laramide-style deformation and remained intact. In this area, preserved sedimentary rocks indicate that at the time of the Cretaceous-Paleogene transition, depositional environments were wholly terrestrial (Fig. 1), although eastward, where late Mesozoic to early Cenozoic rocks may have subsequently been removed (e.g., Izett, 1975; Swinehart et al., 1985), a coeval remnant of the Western Interior Seaway is conjectured (Johnson et al., 2002; Fig. 1). Lithostratigraphic formations that bracket the Cretaceous-Paleogene boundary in this area are the Coalspur Formation (west-central Alberta), Willow Creek Formation (southwestern Alberta), Scollard Formation (east-central Alberta), the Frenchman and Ravenscrag Formations (Saskatchewan), and the Hell Creek and Fort Union formations (Montana and the Dakotas; Fig. 1). The formations form an eastward-thinning wedge (Dawson et al., 1994; Fuentes et al., 2011) of largely high-sinuosity fluvial channel sediments and associated floodplain, lacustrine, paleosol, and mire (coal) facies (Fastovsky and Dott, 1986; Jerzykiewicz and Sweet, 1988; McIver and Basinger, 1993; Eberth and O'Connell, 1995; Murphy et al., 2002; Figs. 1 and 2; Table 1). To the west, nearer the positive topography and high sediment fluxes associated with the active Sevier deformation front, deposition may also have occurred in low-sinuosity channels and associated alluvial plains (Fig. 1; Jerzykiewicz and Sweet, 1988; DeCelles et al., 1987; Eberth and O'Connell, 1995).

The base of the Coalspur, Willow Creek, Scollard, Frenchman, and Hell Creek Formations is marked by an unconformity (Kupsch, 1957; Johnson et al., 2002), which occurs in the paleomagnetic polarity subchron C30n (68.2–66.2 Ma; Fig. 3; Gradstein et al., 2012). Up to 60 m of paleotopography is recognized on this surface (Kupsch, 1957), and it is overlain by a late Maastrichtian to Danian succession (Lerbekmo and Sweet, 2008; Peppe et al., 2011) characterized by an upward transition from coarser, more amalgamated fluvial channel sandstones to finer-grained, better preserved floodplain clastics and coal (Fig. 3; Eberth and O'Connell, 1995; Murphy et al., 2002). This transition is characteristic of the late lowstand systems tracts (LST) to transgressive systems tracts (TST) of fluvial sedimentary sequences (Shanley and McCabe, 1994; Ethridge et al., 1998). This succession is unconformably overlain by the latest Danian to Thanetian fluvial Paskapoo Formation and time-equivalent strata (Lerbekmo and Sweet, 2008; Peppe et al., 2011). Thus, in the study area, the Cretaceous-Paleogene boundary occurs within the late LST to TST of an ~3–6-m.y.-duration (third order; cf. Mitchum and Van Wagoner, 1991) stratigraphic sequence (Hamblin, 2004). Within this transgressive context, macrofloral, microfaunal, and stable isotope studies indicate that the climate in this part of the Western Interior Basin at this time varied from warm subtropical, subhumid to semiarid, and cooled by up to 8 °C through the latest Maastrichtian (Wolfe and Upchurch, 1986; Jerzykiewicz and Sweet, 1988; Johnson et al., 2002; Wilf et al., 2003; Tobin et al., 2014).

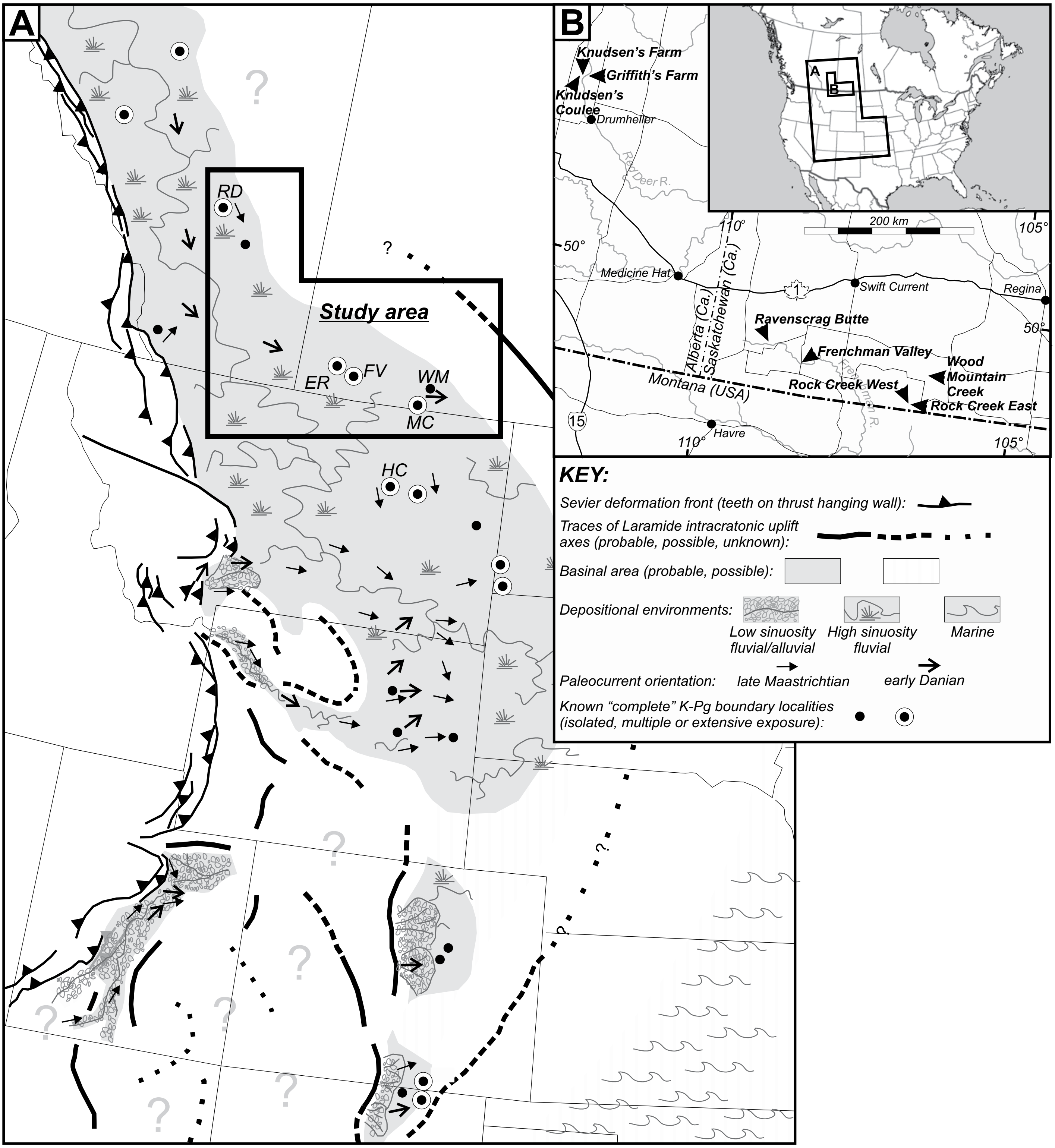

Figure 1. (A). Paleogeographic reconstruction of west-central North America at the time of the Cretaceous-Paleogene (K-Pg) boundary event. The names of locations referred to in the text are given (abbreviations: ER—Eastend/Ravenscrag area; FV—Frenchman Valley; HC—Hell Creek area; MC—Morgan Creek; RD—Red Deer Valley; WM—Wood Mountain Creek). See GSA Data Repository Figure 1 and caption for details (see text footnote 1). (B) Location map of the eight sampling localities of the Nevis/Ferris coal. See Table 2 for precise location coordinates.

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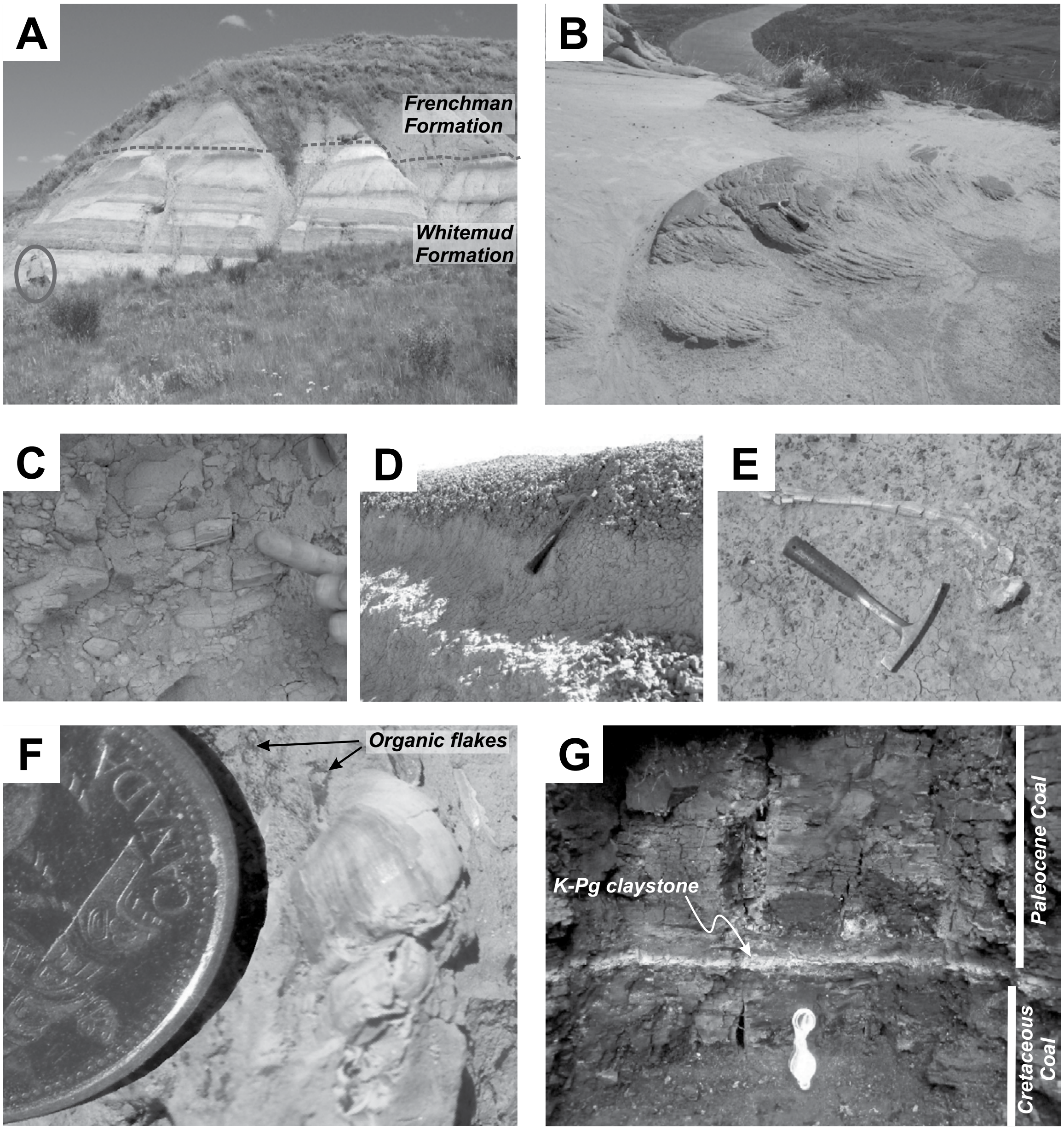

Figure 2. Photographs illustrating facies associations in the stratigraphic interval spanning the Cretaceous-Paleogene (K-Pg) boundary in the study area. (A) LA facies association: paleocurrent-parallel view of large-scale epsilon cross-stratified sandstones and siltstones interpreted as a lateral accretion package (uppermost Whitemud Formation, Wood Mountain Creek). Coauthor circled for scale. (B) SB1 facies association: cross-bedding within channel-fill sandstones (Lower Member of the Scollard Formation, Knudsen's Farm). (C) SB2 facies association: detail of matrix-supported mud-clast conglomerate at the base of a fining-up package (Frenchman Formation, Rock Creek East). (D) OF1 facies association: fining-up package of feldspathic, kaolinitic siltstone to smectitic claystone (Lower Member of the Scollard Formation, Knudsen's Farm). (E) OF1 facies association: disarticulated theropod dinosaur limb in siltstone at the top of a fining-up package (undifferentiated Battle/Whitemud Formation, Rock Creek West). (F) OF2 facies association: freshwater gastropod and disseminated plant material in siltstone (lower Ravenscrag Formation, Ravenscrag Butte). (G) M facies association: ~70-cm-thick coal seam containing a claystone layer interpreted as a reworked air-fall deposit (the Cretaceous-Paleogene boundary claystone; Ferris coal, Frenchman-Ravenscrag Formation, Wood Mountain Creek).

TABLE 1. SUMMARY SEDIMENTOLOGY OF FACIES ASSOCIATIONS IN THE STRATIGRAPHIC INTERVAL SPANNING THE CRETACEOUS-PALEOGENE BOUNDARY IN THE STUDY AREA

Facies association	Lithology and sedimentary structures	Geometry at outcrop and facies relationships	Fossil occurrence	Process
LA. High-sinuosity fluvial distributary	(Climbing) ripple cross-laminated, tabular or trough cross-bedded, massive or plane-bedded, fine to medium sand, occasionally containing mudstone or siderite pebbles. Organized into fining-up cosets up to 1 m thick (sometimes capped by thin coal stingers), dipping at low angle perpendicular to paleoflow and downlapping onto a well-developed (third-order; sensu Miall, 1985) erosional surface.	Tabular. Coset packages are up to 6 m thick, and laterally extensive at outcrop (>100 m). Sharp, scoured lower contact with OF1 and OF2. Sharp, planar lower contact with M. Gradational upper contact with OF1.	Organic flakes. Rare ex situ tree logs.	Migration of ripples and dunes in the lower-flow regime and transport of sediment in the upper-flow regime under unidirectional currents within channels. Lateral accretion of bar forms.
SB1. Low-sinuosity fluvial distributary	(Climbing) ripple cross-laminated, tabular or trough cross-bedded, massive or plane-bedded, fine to medium sand, occasionally containing mudstone or siderite pebbles. Overlying well-developed (third-order; sensu Miall, 1985) erosional surface, but lacking overall grain-size trend, except near the tops, where they fine upward.	Tabular. Successions up to 7 m, and laterally extensive at outcrop (>100 m). Sharp, scoured lower contact with OF1 and OF2. Gradational upper contact with OF1.	Organic flakes. Rare ex situ tree logs.	Migration of ripples and dunes in the lower-flow regime and transport of sediment in the upper-flow regime under unidirectional currents within channels. Fining-upward at the tops records sediment aggradation, reduction in water depth, and channel abandonment.
SB2. Crevasse splay channel	(Climbing) ripple cross-laminated, tabular or trough cross-bedded, massive or plane-bedded, fine to medium sand, occasionally containing mudstone or siderite pebbles. Overlying well-developed (third-order; sensu Miall, 1985) erosional surface, and displaying a well-developed fining-upward succession.	Tabular to lensoid. Successions up to 2 m, pinching out (<100 m) or laterally extensive at outcrop (>100 m). Sharp, scoured lower contact with OF1 and OF2. Sharp, planar lower contact with M. Gradational upper contact with OF1.	Rare to abundant, whole to fragmented, articulated to disarticulated terrestrial reptiles and dinosaurs. In situ tree stumps and plant fragments.	Migration of ripples and dunes in the lower-flow regime and transport of sediment in the upper-flow regime under unidirectional currents within channels. Fining-upward records waning flows.
OF1. Temporary, or shallow floodplain ponds	Ripple cross-laminated, wavy or parallel-laminated siltstone and laminated or massive claystone. Smectitic siltstone and claystones do not display recognizable sedimentary structures. Organized into well-developed fining-up packages. Interbedded with rare fine-to-coarse tuff beds up to 0.1 m thick.	Sheet. Successions up to 2 m, but typically <1 m. Laterally extensive at outcrop (>100 m). Gradational lower contact with LA, SB1, and SB2. Sharp, planar lower contact with OF2. Sharp, scoured upper contact with LA, SB1, and SB2. Gradational upper contact with OF2.	Rare to abundant, whole to fragmented, articulated to disarticulated terrestrial reptiles and dinosaurs. In situ tree stumps and plant fragments. Sideritized and rarely carbonaceous rootlet horizons up to 2 m deep.	Transformation from migration of ripples in the lower-flow regime to suspension-fallout in standing water records waning flow. Roots indicative of periodic subaerial exposure and hiatus in sedimentation.
OF2. Semipermanent or deeper floodplain lakes and ponds	Laminated to massive dark-gray to black claystone and mid- to pale-gray siltstone. Organized into coarsening-up beds up to 1 m thick, but in one instance, 4 m thick. Beds often capped by sideritized beds.	Sheet. Successions up to 5 m. Laterally extensive at outcrop (>100 m). Sharp, planar lower contact with M. Gradational lower contact with OF1. Sharp, planar upper contact with OF1 and M. Sharp, scoured upper contact with LA.	In situ freshwater gastropods. Abundant disseminated organic flakes. Horizons of compressed plants.	Coarsening-up succession records settling-from-suspension from sediment plume surges in calm standing water. Organic-rich nature of deposits indicative of anoxia and/or rapid sedimentation. Siderite horizons and rootlets indicative of sporadic subaerial exposure.
M. Mire	Massive, jointed or laminated coal up to 1.1 m thick with minor dark-gray/black carbonaceous mudstone or siltstone intervals up to 0.3 m thick.	Tabular. Laterally extensive at outcrop (>100 m). Possibly regionally developed. Sharp, planar lower contact with OF1 and OF2. Sharp, planar upper contact with LA, SB2, OF1, and OF2.	Horizons of compressed plant fragments.	In situ peat accumulation.

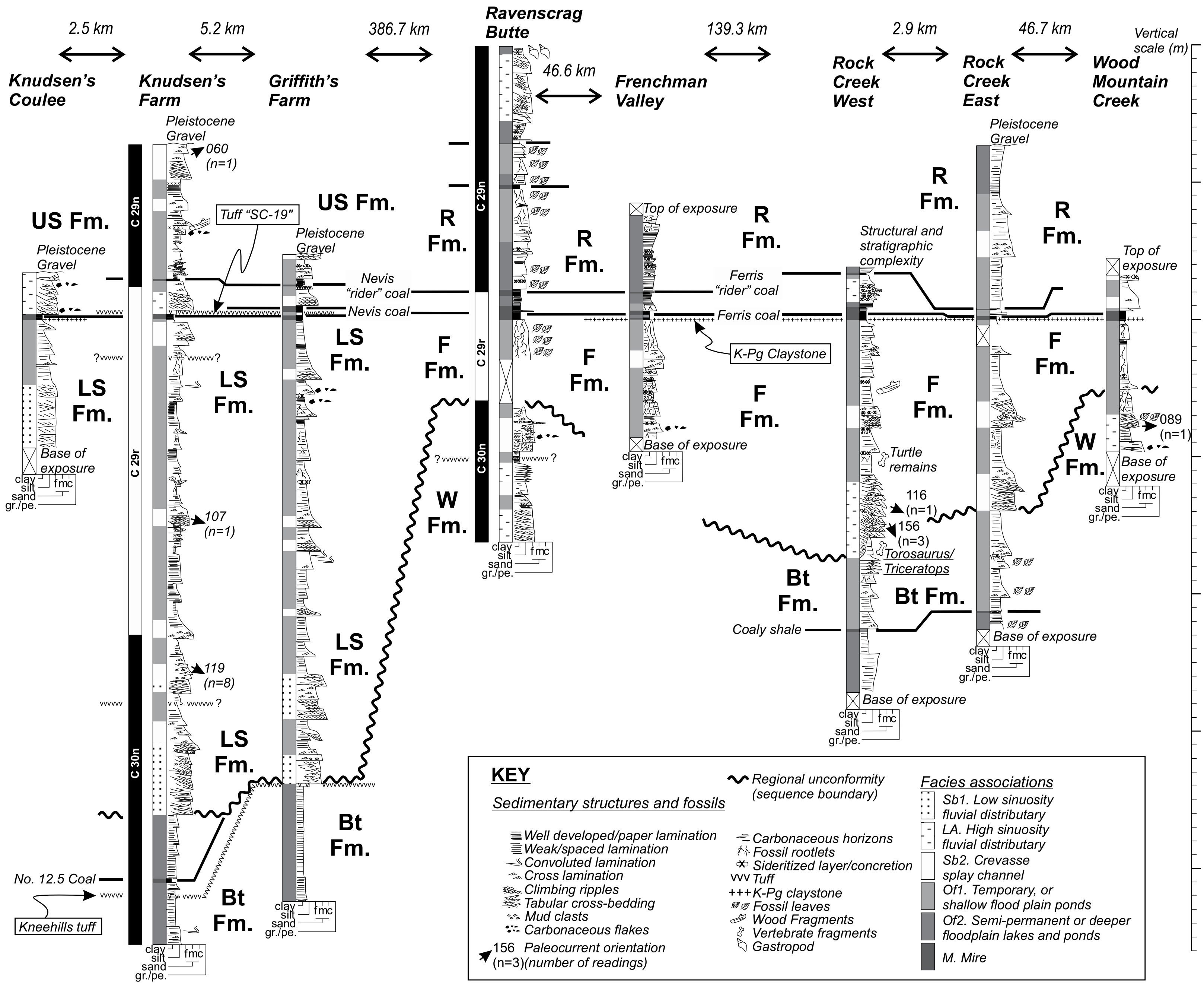

Figure 3. Stratigraphic and sedimentological correlation between all eight measured sections (K-Pg—Cretaceous-Paleogene boundary). Note variable distances between localities. See text for details of age dates. Abbreviations: Bt Fm.—Battle Formation; F Fm.—Frenchman Formation; LS Fm.—Lower Member of the Scollard Formation; R Fm.—Ravenscrag Formation; US Fm.—Upper Member of the Scollard Formation; W Fm.—Whitemud Formation; fmc—fine, medium and coarse.

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"Complete" Cretaceous-Paleogene boundary successions, which record active accumulation and subsequent nonerosion of sediment during the Cretaceous-Paleogene "event," are marked by the occurrence of a composite light brown-to-buff or pink claystone <5 cm thick, containing shock-metamorphosed minerals and/or an Ir anomaly and/or a "spike" in fern pollen abundance (Sweet et al., 1999; Nichols, 2007). Throughout the northern Western Interior Basin, the Cretaceous-Paleogene claystone usually underlies, but also overlies or occurs as a clastic parting within, a coal (Fig. 3G; Bohor et al., 1984; Smit and Van der Kaars, 1984; Tschudy et al., 1984; Nichols et al., 1986; Lerbekmo et al., 1987; Johnson et al., 1989; Sweet et al., 1990, 1999; Sweet and Braman, 1992, 2001; Hotton, 2002; Nichols and Johnson, 2002). The coal is known variously as the Mynheer (central Alberta), Nevis (southern Alberta), Ferris (southern Saskatchewan) or "Z" (Montana and North Dakota) coal, and it is the focus of this study.

In the study area (Fig. 1), burial of the Cretaceous-Paleogene boundary horizon by 0.5–4 km of younger sediments, from east-to-west (Beaumont, 1981; Bustin, 1991; Cameron, 1991), resulted in a first-order variation in present-day coal rank from lignite to subbituminous in the same direction (Bustin, 1991; Cameron, 1991; Smith et al., 1994). Uplift and erosion from the mid Eocene resulted in removal of much of this overburden, and rocks spanning the Cretaceous-Paleogene boundary are now exposed along the walls of canyons incised into the uplifted plateau and on the eroded flanks of nunataks that resisted the advance of Pleistocene glaciation.

METHODS

Field Sampling

Eight exposures of the Ferris-Nevis coal in south-central Alberta and southwestern Saskatchewan were selected for analysis in this study (Fig. 1; Table 2). All the localities have been previously described, and at six of these (Frenchman Valley, Knudsen's Coulee, Knudsen's Farm, Rock Creek East, Rock Creek West, and Wood Mountain Creek), a conformable Cretaceous-Paleogene boundary has been documented by the occurrence of the boundary claystone associated with an Ir anomaly (Table 2). However, at the time of field work for this study (July 2011), previous sample collection had resulted in total removal of the formerly described boundary claystone at the Knudsen's Farm locality. Bentonitic carbonaceous mudrock of volcanic origin is also interbedded with the coal, locally providing additional

TABLE 2. SAMPLING LOCALITY DETAILS AND PREVIOUS STUDIES

Location information			Previous work								Additional remarks
Locality name	Coordinates	Description	Detailed site sedimentological description	K-Pg claystone present?	Shocked minerals present?	Iridium anomaly present?	Palynological study	Fern spore spike present?	Coal petrology	Carbon isotope analyses	
Frenchman Valley	49°20′56″N 108°25′05″W	Partly tallus-covered exposure of the upper Frenchman and lower Ravenscrag Formations, along an ~500-m-long road cut NE of Highway 37 in the valley of the Frenchman River, between the settlements of Climax and Shaunavon.	No	Yes[1]	?	Yes[1,2]	Yes[1,3]	Yes[1]	No	No	The Cretaceous-Paleogene (K-Pg) claystone layer is discontinuous at outcrop scale.
Griffith's Farm	51°54′47″N 112°57′51″W	Sampling sites occur in subcontinuous exposure of the upper Horseshoe Canyon Formation to lowermost Upper Member of the Scollard Formation in badlands along the Red Deer River and side canyons, between Morrin Bridge and McKenzie Bridge. Exposure is commonly disrupted by recent landsliding.	No	No[4]	?	Yes[4]	No	?	No	No	Known as "Scollard Canyon, locality 2" in Sweet and Braman (2001).
Knudsen's Coulee	51°54′27″N 113°02′57″W		Yes[5,6]	Yes[5]	?	Yes[2]	Yes[3,5]	Yes[5]	No	Yes[7]	The K-Pg claystone layer no longer occurs as a result of oversampling.
Knudsen's Farm	51°53′21″N 113°01′42″W		No	Yes[8]	Yes[9,10,11,12]	Yes[13,1,8]	Yes[1,5]	Yes[5]	No	Yes[7]	
Ravenscrag Butte	49°30′11″N 109°01′03″W	Disused quarry occurring in subcontinuous, often tallus-covered exposure of the Eastend to lower Ravenscrag Formations in badlands along the valley of the Frenchman River, between the settlements of Ravenscrag and Eastend.	Yes[14]	No[4]	?	Yes[4]	No	?	No	No	Type section for the Ravenscrag Formation.
Rock Creek East	49°01′27″N 106°32′03″W	Sampling sites occur in subcontinuous exposure of the Whitemud to lower Ravenscrag Formations in badlands along the valley of Rock (or, erroneously, Morgan) Creek, east block of Grasslands National Park.	Yes[5]	Yes[5]	?	Yes[4]	Yes[5,3]	Yes[5]	Yes[15]	No	Also known as Zahursky's Point.
Rock Creek West	49°02′20″N 106°34′00″W		No	Yes[16]	Yes[16,17]	Yes[16,2]	Yes[16,5,3]	Yes[16,5,3]	Yes[18]	No	Also known as the Valley of a Thousand Devils.
Wood Mountain Creek	49°25′20″N 106°19′50″W	Partly overgrown exposure of the upper Whitemud to lower Ravenscrag Formations in a disused quarry west of Wood Mountain Creek.	No	Yes[5]	?	Yes[4]	Yes[5,3]	Yes[3]	Yes[15,18]	No	

Note: References: 1—Lerbekmo et al. (1987); 2—Lerbekmo et al. (1996); 3— Sweet et al. (1999); 4—Sweet and Braman (2001); 5—Sweet and Braman (1992); 6—Eberth and O'Connell (1995); 7—Therrien et al. (2007); 8—Hildebrand and Boynton (1988); 9—Bohor and Izett (1986); 10—Bohor et al. (1987a); 11—Bohor et al. (1987b); 12—Grieve and Alexopoulos (1988); 13—Lerbekmo and Coulter (1984); 14—McIver and Basinger (1993); 15—Belcher et al. (2003); 16—Nichols et al. (1986); 17—Kamo and Krogh (1995); 18—Sweet and Cameron (1991).

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correlation tie lines (e.g., tuff “SC-19” sensu Eberth and Deino, 2005).

At each locality, centimeter-scale sedimentary logging of the encasing strata was carried out (Fig. 3), recording the full range of grain sizes, sedimentary structures, and body and trace fossils observed. It was necessary to excavate the coal surface by up to 1 m before sampling in order to limit the effects of weathering on subsequent analyses. A coal lithotype log, using the classification scheme of Diessel (1992), was produced to ensure that important lithological surfaces were identified prior to sampling. In total, 184 contiguous samples of coal and associated clastic sediment were recovered, representing the whole coal seam thickness of nine coals at the eight localities (one coal at each locality, but two coals at Griffith’s Farm; Figs. 4 and 5). Whenever possible, the specimens were removed intact, and their younging direction was recorded, such that their internal stratigraphy could be preserved. The average stratigraphic thickness of each specimen was ~3 cm.

Petrological Analysis

The land plants that accumulated in mires in the Cretaceous and early Paleocene all utilized the C3 photosynthetic pathway (Osborne and Beerling, 2006), in which atmospheric CO_2 is taken up and fractionated in a quantifiable way during the fixation of carbon into biomass (Farquhar et al., 1989). Thus, whole plant average isotopic composition of the plants growing in Cretaceous-Paleogene mires recorded changes in the isotopic composition of the atmo-

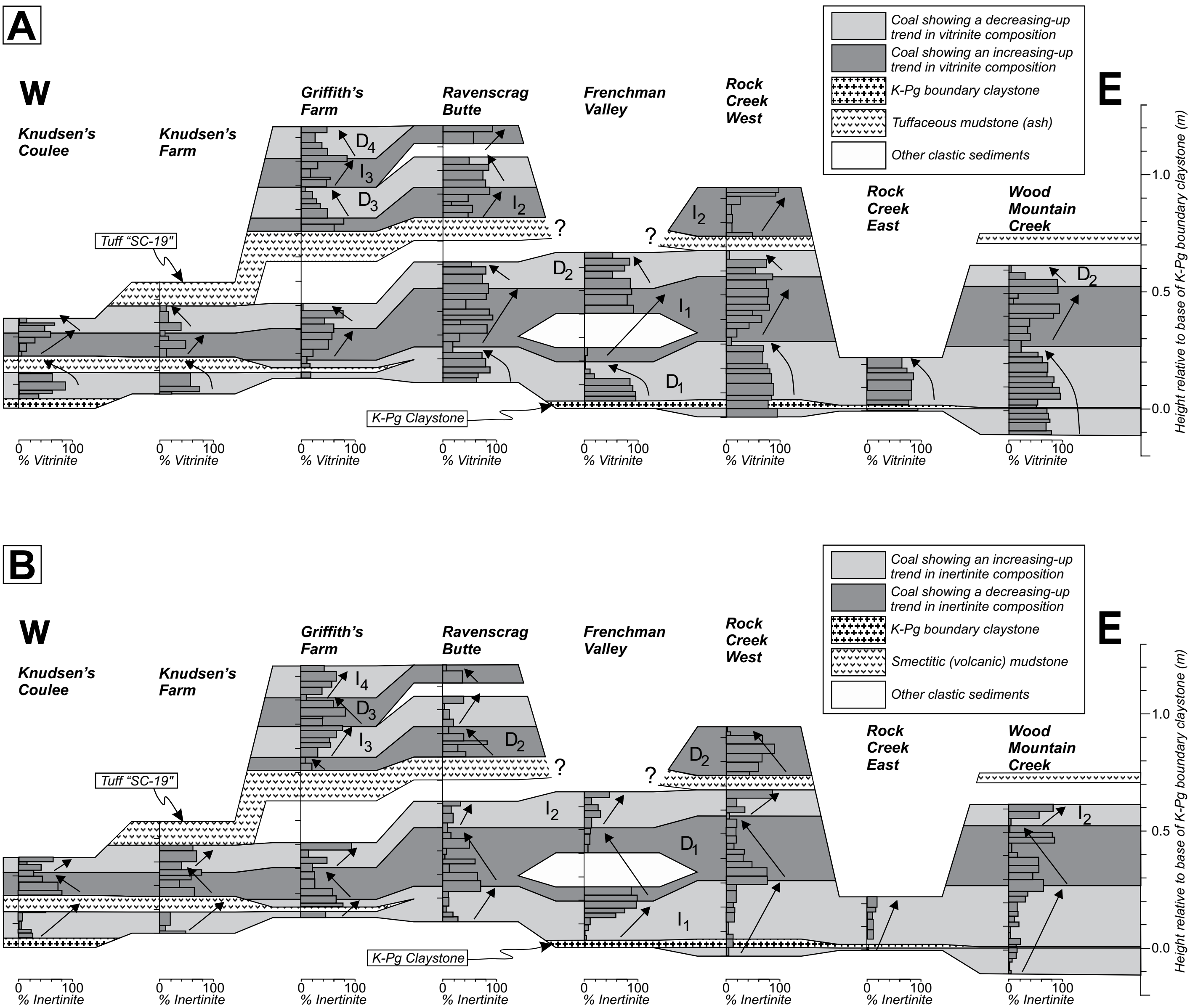

Figure 4 (*on this and following page*). Schematic correlation panel showing the petrological properties of the Nevis/Ferris coal at all eight sampled localities. Datum is the base of the Cretaceous-Paleogene (K-Pg) claystone layer. Correlation lines are for interpretational purposes only and do not represent the interpreted stratigraphic architecture of the Nevis/Ferris coal. The vertical bar width represents the width over which the contiguous sample was collected. (A) Vitrinite content (mineral free). (B) Inertinite content (mineral free).

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sphere at the time at which it was fixed (Lloyd and Farquhar, 1994), with a minor, quantifiable (0.08‰) error due to fluctuations in ecological conditions (including light, nutrient, and water availability and salinity), and variations in plant physiology (Ahrens et al., 2000). However, within C3 plants, $\delta^{13}C$ values range between –23‰ and –24‰ (e.g., Marino and McElroy, 1991; Arens et al., 2000), and the formation of peat involves the selective degradation of the least-resistant plant tissues, followed by minor reorganization of the bipolymers that remain (Hatcher and Clifford, 1997). It is therefore possible that changes in the isotopic composition of the preserved plant material in coal could represent changes in the type and degree of chemical alteration experienced by plant matter after death, before and during the lifetime of the accumulating mire (Benner et al., 1987; Bechtel et al., 2007). During the accumulation of peat, the selective and variable degradation of contributing plant material results in a predictable suite of macroscopic to microscopic organic materials with a distinctive chemistry and physical structure (Frenzel, 1983; Diessel, 1992; Scott, 2002). During the diagenetic conversion of peat to coal, the distinct chemistry and morphology of these grains are maintained, except at high ranks of coal (i.e., anthracite); these distinct grains can be recognized optically and are known as macerals (Stopes, 1935; Diessel, 1992). It is therefore possible to use coal petrology to quantify the type and degree of chemical alteration experienced by the components of peat before conversion to coal, and to assess the potential impact on the resulting $\delta^{13}C$ record.

Hence, of the 184 collected samples, 179 were cured whole in epoxy resin, cut perpendicular to depositional layering, and polished in accor-

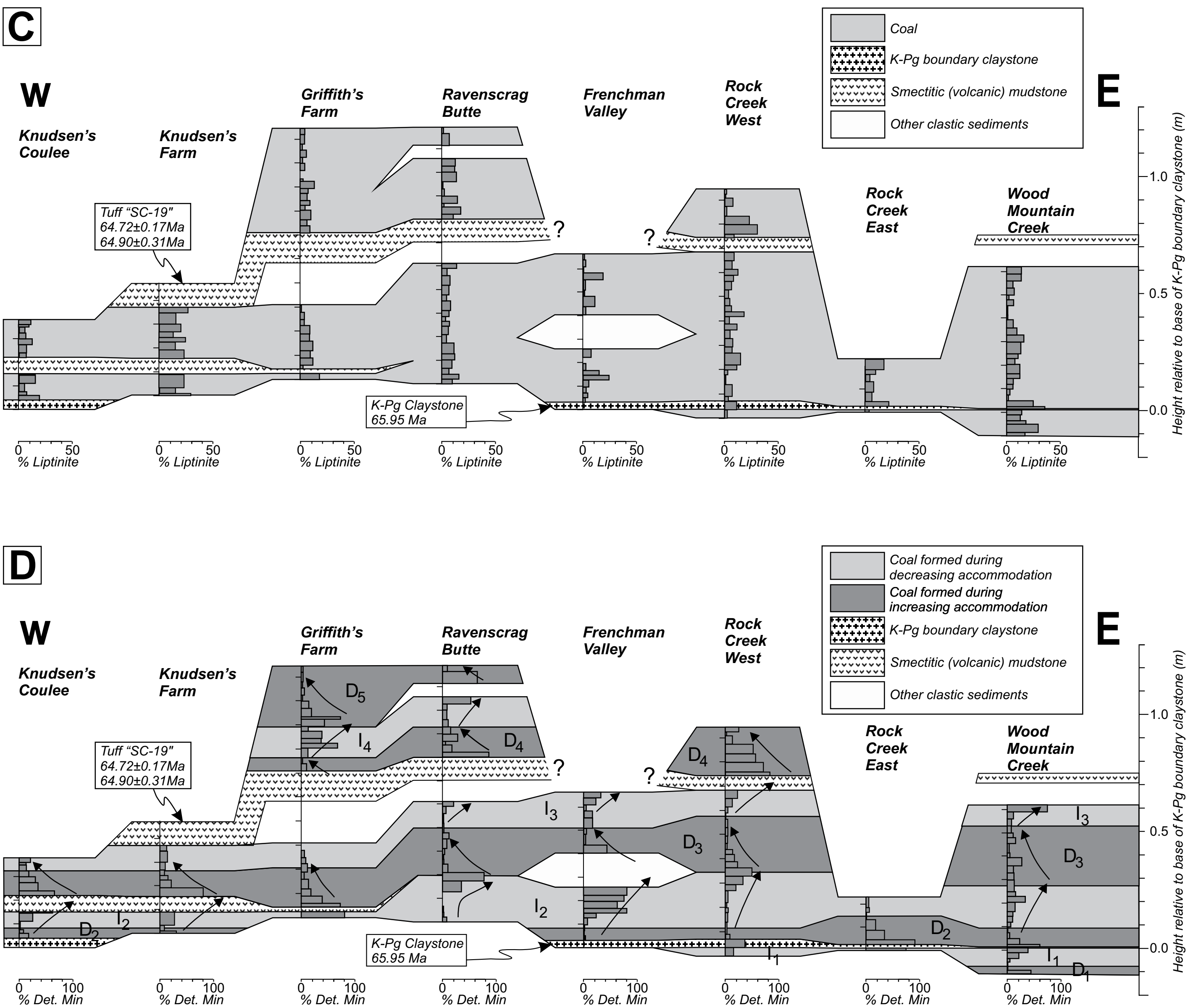

Figure 4 (*continued*). (C) Liptinite content (mineral free). (D) Total mineral content. Abbreviations: D1—decreasing-up trend 1; I1—increasing-up trend 1; D2—decreasing-up trend 2; I2—increasing-up trend 2; D3—decreasing-up trend 3; I3—increasing-up trend 3; D4—decreasing-up trend 4; I4—increasing-up trend 4; D5—decreasing-up trend 5.

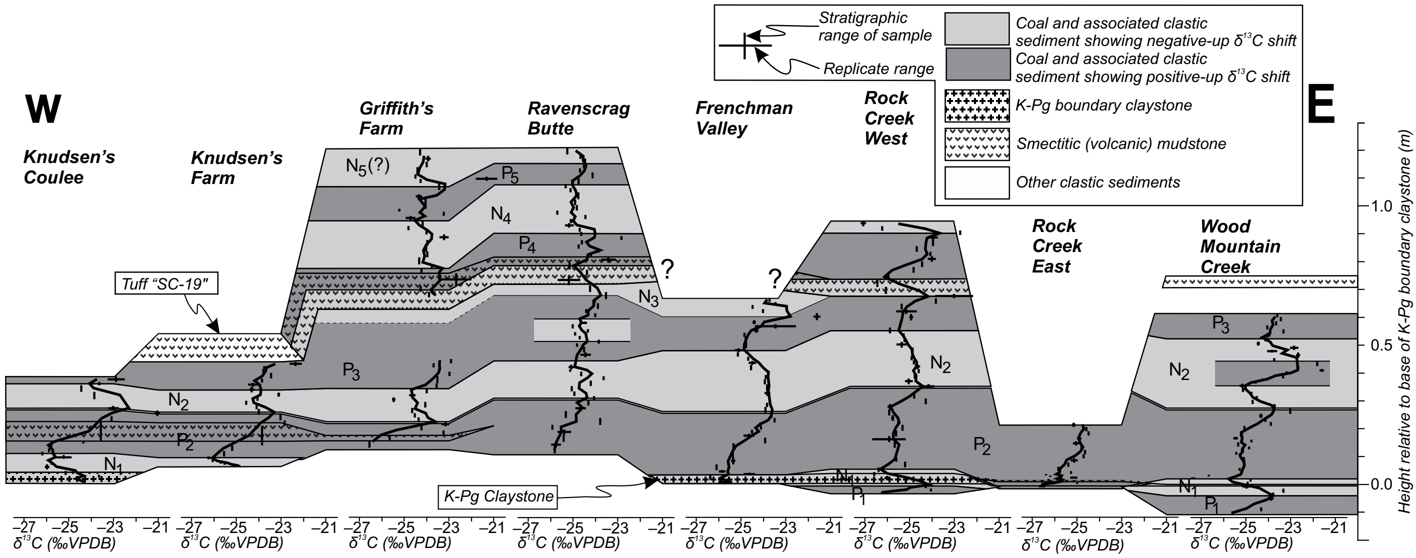

Figure 5. Correlation panel of δ^{13}C values of all 319 subsamples of coal and intervening clastic sediment associated with the Nevis/Ferris coal at all eight sampled localities. Vertical bars indicate the stratigraphic range of the samples, and horizontal bars indicate the δ^{13}C range where multiple readings from the same sample were taken. A moving mean curve of every three samples is superimposed upon the data. Datum is the base of the Cretaceous-Paleogene (K-Pg) claystone layer. Correlation lines are for interpretational purposes only and do not represent the interpreted stratigraphic architecture of the Nevis/Ferris coal. VPDB—Vienna Peedee belemnite.

dance with standard methods for oil-immersion incident light microscopy (e.g., Australian Standard AS 2061–1989, 1989). The five remaining samples were too brittle to be collected intact and were therefore crushed to a maximum grain size of 2 mm, and a representative grain mount was produced instead. The maceral and mineral composition of 166 samples was determined by counting 300 points per sample using a manual point counter with a stepping distance of 0.5 mm, in accordance with standard guidelines (Australian Standard AS 2856.2–1998 1998), except the macerals semifusinite and fusinite were defined as >0.02 mm. The petrological composition of the remaining 18 coaly mudrock samples could not be determined due to their inherently high smectite content, which prohibited adequate polishing for incident light microscopy. A summary of the origin and significance of the different maceral and mineral components of peat is shown in Table 3.

TABLE 3. ORIGINS OF THE MACERAL AND MINERAL COMPONENTS OF COAL

Group	Maceral	Origin	Significance
Vitrinite	Telovitrinite	Humified stem, root, bark, and leaf tissue, which has survived intact and displays remnants of cellular structure	High vitrinite content, especially the structured telovitrinite, indicates rapid preservation under anaerobic conditions
	Detrovitrinite	Stem, root, bark, and leaf tissue deposited as fine-grained attritus prior to humification	
Liptinite	Sporinite	Resins, fats, waxes, and oils	Increased content of the mechanically resistant liptinites indicates loss of vascular biomass associated with poor preservation conditions
	Cutinite	Cuticles of needles, shoots, stalks, leaves, roots, and stems	
	Resinite	Resins, fats, waxes, and oils	
Inertinite	Fusinite	Partially combusted (pyrolyzed) vascular plant material that has survived intact and shows remnants of cellular structure	High fusinite and semifusinite content is indicative of rapid preservation via pyrolysis or partial oxidation at elevated temperatures; high inertodetrinite and the unstructured macrinite and micrinite content indicate poor preservation under relatively aerobic conditions
	Semifusinite	Weakly humified vascular plant material that has also undergone partial microbial oxidation, dehydration, and/or combustion, has survived intact, and shows remnants of cellular structure	
	Inertodetrinite	Fragmented semifusinite and fusinite	
	Macrinite	(Partially) microbially oxidized plant material that has lost visible physical structure through gelification, or (partially) microbially oxidized lipids	
	Micrinite	Product of disproportionation reactions, or any other inertinite detritus	
Inorganic minerals	Detrital minerals	Allochthonous clastic sedimentation	High detrital mineral content is indicative of (periods of) standing water, but may be concentrated by biomass loss associated with poor preservation conditions; syngenetic pyrite is indicative of the bacterial reduction of sulfates in the presence of iron in anaerobic conditions
	Syngenetic pyrite	Framboidal and concentric forms of pyrite that crystallized syngenetically with peat	

Note: Compiled from Berner (1970); Lyons et al. (1986); Cohen et al. (1987); Teichmüller (1989); Diessel (1992, 2010); Scott and Jones (1994); Guo and Bustin (1998); Petersen et al. (1998); Taylor et al. (1998); Diessel et al. (2000); Sýkorová et al. (2005); Scott and Glasspool (2007).

$\delta^{13}C$ Analysis

The 184 samples were further subdivided into 319 subsamples for $\delta^{13}C$ analysis. Whole-rock samples were oven dried (30 °C, 24 h), crushed using a granite pestle and mortar, and decarbonated using excess hydrochloric acid (10% v/v) until any visible reaction had ceased. The samples were then repeatedly washed with deionized water until a neutral solution was obtained, and samples were oven dried (30 °C, 24 h). Carbon isotope analyses were conducted using the Plymouth University Isoprime isotope ratio mass spectrometer connected to an Isoprime Microcube elemental analyzer. Carbon isotope ratios are expressed using the internationally accepted per mil (‰) standard notation relative to the Vienna Peedee belemnite (VPDB) standard. Instrument calibration was achieved using three international standards, USGS 40 (l-glutamic acid, $\delta^{13}C$ = –26.389‰), USGS 24 (graphite, $\delta^{13}C$ = –16.049‰), and IAEA CH-7 (polyethylene, $\delta^{13}C$ = –32.151‰). The standard deviation on replicates in run analyses of the USGS 40 standard was ±0.12‰.

RESULTS

The results of all petrological and $\delta^{13}C$ analyses performed on the 184 collected samples and 319 subsamples are provided in GSA Data Repository Table S1.[1]

Petrological Analysis

The average composition of all 166 petrological analyses from the eight sampling localities is 46.2% vitrinite, 8.2% liptinite, 23.6% inertinite, 21.9% detrital minerals, and less than 0.1% syngenetic pyrite. Total vitrinite composition varies from 24.7% at Knudsen's Farm to 55.5% at Ravenscrag Butte; total liptinite composition varies from 4.0% at Frenchman Valley to 15.6% at Knudsen's Farm; total inertinite composition varies from 8.4% at Rock Creek East to 37.1% at Knudsen's Farm; and total detrital mineral content varies from 17.9% at Ravenscrag Butte to 32.9% at Rock Creek East. Individual measurements of maceral and mineral groups vary from 0.7% to 93.0% for vitrinite, 0% to 31.7% for liptinite, 0% to 94.0% for inertinite, and 0% to 90.3% for mineral material (GSA Data Repository Table S1 [see footnote 1]). Although individual measurements in this study fall within the usual range expected for coals (Diessel, 1992), the large range of values, particularly between individual samples, but also between the average composition of the eight sections, could be suggestive of significant variation in the source of organic matter and degree of organic matter degradation before and during peat accumulation, which could have altered the inherent atmospheric $\delta^{13}C$ record.

Vertical profiles of the vitrinite, inertinite, liptinite, and detrital mineral content for each of the eight sampled sections are shown in Figures 4A–4D. Maceral percentages were calculated on a mineral-free basis. The Cretaceous-Paleogene boundary claystone was used as a datum, and two other tuffs beds were used as independent correlation tie points. By subdividing the coal into time-equivalent depositional units that broadly display either increasing-up (I) or decreasing-up (D) trends in vitrinite, inertinite, and detrital mineral composition, regional sinusoidal trends in these indices could be determined. Up to three complete cycles in vitrinite and inertinite composition are recognizable, starting with a decreasing-upward trend in vitrinite, and an increasing-upward trend in inertinite at the base of the coal beneath and through the Cretaceous-Paleogene boundary itself. Up to four cycles are recognized in detrital mineral composition, starting with a decreasing-up trend at the base of the coal (Figs. 4 and 6). For the vitrinite, inertinite, and detrital mineral contents, the cycles are most convincingly correlated between the Cretaceous-Paleogene claystone and the upper SC-19 tuff bed, where data from all eight sampled sections can be filtered visually to eliminate noise and more minor, high-frequency cycles developed in some, but not all, the sections. Liptinite composition is more difficult to correlate between the sampled sections. This may be an analytical artifact: When coals contain large volumes of detrital minerals, they can be difficult to distinguish from liptinites in the absence of a fluorescent reflected light source.

$\delta^{13}C$ Analysis

The average $\delta^{13}C$ of all coal samples is –24.6‰, varying from –25.4‰ at Rock Creek East to –24.1‰ at Griffith's Farm, with individual measurements varying from –27.3‰ to –21.0‰ (GSA Data Repository Table S1 [see footnote 1]), consistent with the composition of C3 plants (O'Leary, 1988). Regression of $\delta^{13}C$ measurements against the maceral and mineral composition of all samples shows that there is no correlation between the maceral composition and the $\delta^{13}C$ measurement of individual samples (GSA Data Repository Table S2 [see footnote 1]). Hence, the style of chemical alteration experienced by plant material during conversion to peat does not appear to have impacted the $\delta^{13}C$ composition of the preserved organic material in coal. Thus, the measured $\delta^{13}C$ values from the coal samples are representative of the composition of the vegetation that contributed to the original peat, which in turn, reflect paleoatmospheric CO_2 values.

The $\delta^{13}C$ values of all 319 subsamples of coal and intercalated clastic sediment associated with the Nevis/Ferris coal at all eight sampled sections are shown in Figures 5 and 6. By subdividing the coal into time-equivalent depositional units that display positive (P) and negative (N) $\delta^{13}C$ trends, regionally developed trends are evident. The correlation of $\delta^{13}C$ trends is consistent with that of the petrological trends, in that the $\delta^{13}C$ correlation lines are broadly parallel to and do not cross the petrological correlation lines (cf. Fig. 4). From the base of the coal at Wood Mountain Creek to the top at Griffith's Farm, at least five positive-negative excursion cycles are readily determined. Two complete cycles occur between the Cretaceous-Paleogene claystone and the SC-19 tuff bed. Determining a "background" $\delta^{13}C$ value against which the magnitude of carbon isotope excursions can be measured is difficult, due to the inherent sinuosity displayed though the sections. However, following Grandpre et al. (2013), a two-sample Kolmogorov-Smirnov (K-S; Hammer and Harper, 2006) test was used to determine the statistical significance of the observed Cretaceous-Paleogene carbon isotope shifts in the context of observed isotopic variations immediately above and below the Cretaceous-Paleogene boundary at Rock Creek West and Wood Mountain Creek. The boundary sample size (n = 3) for each section consists of the $\delta^{13}C$ values of the proposed Cretaceous-Paleogene negative shift and n = 17 for the comparison sample for each section. The results of the K-S tests for both sections indicate that a difference could be demonstrated (at the 0.05 level of significance) between the boundary excursion and variations above and below the extinctions. However, when the entire data set for each section is included as the comparison sample, a statistical difference cannot be seen. This undoubtedly relates to the observation that at Rock Creek West and Wood Mountain Creek, the Cretaceous-Paleogene $\delta^{13}C$ minimum is followed by up to four more positive-negative excursions (see following).

DISCUSSION

Petrology of the Cretaceous-Paleogene Coals

The petrology of coal seams has been widely used to reconstruct the paleoenvironmental conditions in ancient mires and the surrounding

[1]GSA Data Repository item 2015170, details of how Figure 1 was compiled, and all petrographic and $\delta^{13}C$ analyses, is available at http://www.geosociety.org/pubs/ft2015.htm or by request to editing@geosociety.org.

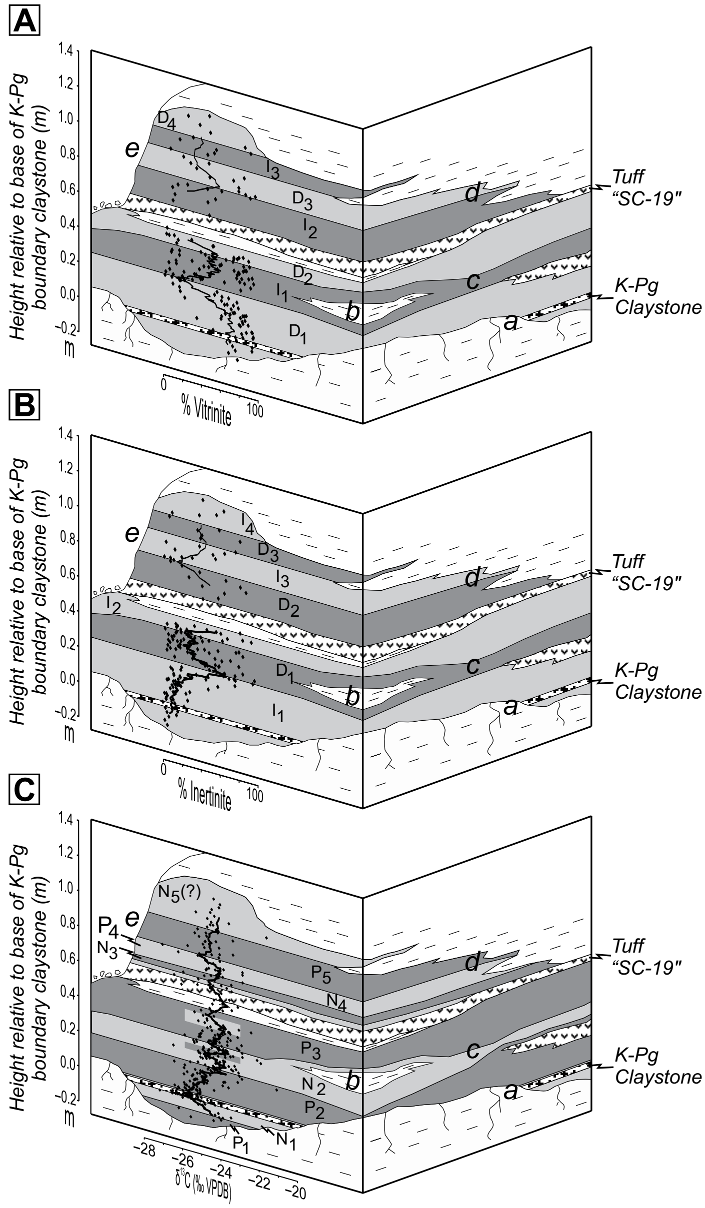

Figure 6. All (A) vitrinite, (B) inertinite, and (C) δ^{13}C data points measured in this study (K-Pg—Cretaceous-Paleogene boundary). The median stratigraphic range of each data point is indicated, as well as a moving average line of every 10 points. Correlated units discussed in the text are indicated. (A) The stratigraphic positions of vitrinite data have been stretched or collapsed to fit the thickness of unit D_1 at Wood Mountain Creek, and units I_1 to D_4 at Griffith's Farm (cf. Fig. 4A). (B) The stratigraphic positions of inertinite data have been stretched or collapsed to fit the thickness of unit I_1 at Wood Mountain Creek, and units D_1 to I_4 at Griffith's Farm (cf. Fig. 4B). (C) The stratigraphic positions of δ^{13}C data have been stretched or collapsed to fit the thickness of units P_1 to P_2 at Wood Mountain Creek, and units N_2 to N_5 at Griffith's Farm (cf. Fig. 5). VPDB—Vienna Peedee belemnite. The data are superimposed on a summary of the range of stratal and stratigraphic relationships observed in this study that influence the preservation of the petrological and δ^{13}C record of coal. (a) Loss of the lower part of the record due to onlap against inherited topography. (b) Expansion of the record due to rapid sedimentation associated with avulsion into the mire and/or deposition of volcaniclastics. (c) Contraction of the record due to differential compaction and/or reduced rates of peat accumulation. (d) Loss and/or expansion of the upper part of the record through progressive interfingering with overlying clastic sediments. (e) Loss of the upper part of the record through removal by channel incision. See Figures 4 and 5 for key to other symbols.

environment (e.g., Cohen et al., 1987; Spears, 1987; Diessel, 1992; Scott et al., 2000; Belcher et al., 2003; Davies et al., 2006; Collinson et al., 2007; Hudspith et al., 2012). The three maceral groups recognized petrologically in coals are vitrinites, inertinites, and liptinites (Table 3). Vitrinites are largely derived from vascular plant material (cell walls) that has undergone humification in the absence of oxygen. Liptinites are the products of H-rich plant organs such as spores, cuticles, waxes, and resins. The origin of inertinites is the same as vitrinites and liptinites, but they have usually undergone charring (pyrolysis) prior to anaerobic humification (Table 3; Scott and Glasspool, 2007). Because individual macerals within the maceral groups are classified according to their size and morphology, numerous studies (e.g., Scott et al., 2000; Belcher et al., 2003; Collinson et al., 2007; Hudspith et al., 2012) have utilized the distribution of different-sized inertinite macerals through coal to reconstruct distant (windblown), canopy, surface, and ground fire events, the frequency of which, by analogy with studies of Holocene charcoal records, may be related to climate change on geologically short time scales (Power et al., 2008; Marlon et al., 2012).

Published petrological studies of Cretaceous-Paleogene boundary coals (i.e., Sweet and Cameron, 1991; Belcher et al., 2003) do not tabulate coal composition, but on a mineral-free basis, the bulk composition of the Cretaceous-Paleogene boundary coals in this study is enriched in inertinite and liptinite relative to slightly younger Paleocene coals from the Frenchman Formation in the same study area (Beaton et al., 1991; Potter et al., 1991; Frank and Bend, 2004) as well as relative to Mesozoic to Cenozoic coals and Quaternary peats more generally (Diessel, 2010; Glasspool and Scott, 2010). This suggests that the frequency of wildfires in the interval bracketing the Cretaceous-Paleogene boundary was relatively high, superficially supporting the hypothesis that bolide impact at the Cretaceous-Paleogene boundary would have ignited regional-to-global-scale wildfires (Melosh et al., 1990; Woolbach et al., 1990; Morgan et al., 2013). In detail, however, within the sampled Cretaceous-Paleogene coal, total inertinite (charred material) and vitrinite (noncharred material) fluctuate cyclically, and up to three full cycles can be readily determined from the base to the top of the coals. In the interval immediately bracketing the Cretaceous-Paleogene boundary, total inertinite composition is low (<10%), and vitrinite composition is high (>50%; Figs. 4 and 6). Inertinite composition gradually increases to a peak of >60% at the top of unit I_1, and vitrinite composition gradually falls to <15% at the top of unit I_1, some 20–30 cm above the Cretaceous-Paleogene boundary claystone. This result is precisely in accordance with the observations of Belcher et al. (2003), who proposed, on the basis of petrological evidence from North American coals, that there was no evidence from the peat record to support the Cretaceous-Paleogene global wildfire hypothesis (cf. Morgan et al., 2013). These cycles in the inertinite and vitrinite composition of the coals can be correlated over distances of over 500 km (Figs. 4 and 6), strongly indicative of some regional-scale control on the stratigraphic distributions of these indices of wildfire. Moreover, the quantities of microscopic inertinite, indicative of distant, windblown soot, soot washed-in by rivers, or fires in the vegetation canopy, and macroscopic inertinite, indicative of surface or ground fires in the peat itself, increase and decrease in tandem (Figs. 4 and 6). This suggests a regional, most likely climatic, control on the occurrence of wildfires, both in the canopy of growing vegetation and within the peat itself. At individual locations, wildfire occurrence is typically promoted by increased temperature and limited rainfall, although an ignition mechanism such as lightning, wind as a spreading mechanism, and fuel are also required (e.g., McKenzie et al., 2004; Daniau et al., 2012).

Not all inertinite is universally believed to have been formed by charring, and some is considered the result of microbial and fungal oxidation after burial (O'Keefe et al., 2013; Table 3). Such oxidation of plant material is promoted by long residency of plant material in the upper 0.5 m of the peat profile (the acrotelm; sensu Ingram, 1982), in response to a relatively slowly rising water table to generate accommodation for the permanent accumulation of peat (Clymo, 1984; Diessel, 1992). Whether the product of charring or biogenically mediated oxidation, the regional-scale, cyclic nature of inertinite and vitrinite distribution within the Cretaceous-Paleogene coals in western Canada is suggestive of regional-scale cyclic fluctuations in temperature and/or rainfall during the lifetime of the mires. There is no suggestion from the petrological data of an increase in wildfire occurrence, or any significant change in the climatic regime at or immediately following the Cretaceous-Paleogene boundary event, which is surprising given the scale of local and global ecological devastation at this time.

The mineral content of coals has also been used in paleoenvironmental studies (e.g., Cohen et al., 1987; Spears, 1987; Davies et al., 2006). Detrital mineral matter is transported into peat largely via fluvial or marine inundation. Coal containing low quantities of detrital minerals (typically <10% by volume) has therefore been interpreted as the product of peat accumulation in raised, ombrotrophic mires, where peat accumulation utilizes excess rainfall-derived moisture to dome above the regional water table and level of regional flooding (Staub, 1991; Diessel et al., 2000). Coals with significant detrital mineral volumes have been interpreted as representing peat that accumulated in low-lying, rheotrophic mires, which typically in-fill topographic hollows and are readily subjected to flood events. However, oxidative biomass loss from peat can concentrate inorganic mineral material into discrete layers (McCabe 1984; Jerrett et al., 2011a). The high mean detrital mineral content of the Cretaceous-Paleogene coals is indicative of significant primary clastic input into the mires (Petersen and Andsbjerg, 1996; Staub, 2002), even between the events that formed the clastic partings that are a feature of the Cretaceous-Paleogene boundary coals (Fig. 4D). This supports the notion that the mires were largely rheotrophic and were limited in spatial extent (the sampled coals represent a series of coeval, disconnected mires), close to a source for the clastic material. The concentration of detrital minerals may also have been enhanced by oxidative degradation and organic biomass loss, consistent with the relatively high inertinite and liptinite content of the coal (McCabe, 1984; Diessel, 1992; Jerrett et al., 2011c). The four complete cycles in the volume of detrital mineral contained within the Cretaceous-Paleogene coals (Fig. 4D) may represent cyclic, simultaneous transitions from rheotrophy to ombrotrohpy and back again within the Cretaceous-Paleogene mires. Neither modern nor ancient mires are usually interpreted as wholly rheotrophic or ombrotrophic in their structure (e.g., Greb et al., 2002), and the contemporaneous switch from rheotrophy to ombrotrophy in numerous Holocene northwest European mires (termed the "fen-bog-transition") is well documented as a response to regional increases in effective rainfall (Walker, 1970; Blackford and Chambers, 1991). Hence, regional fluctuations in detrital mineral content of the Cretaceous-Paleogene coals could reflect changes from rheotrophy to ombrotrophy and vice versa, in turn reflecting basin-scale changes in rainfall patterns. Although Cretaceous-Paleogene mires have, in part, been interpreted as ombrotrophic in origin (Sweet and Cameron, 1991), there is little evidence to further support the hypothesis that cycles in detrital mineral content reflect transitions from rheotrophy to ombrotrophy. Cycles in detrital mineral content are not in phase with the cyclicity in vitrinite and inertinite composition of the coal. Additionally, comparison of the detrital mineral composition of the Cretaceous-Paleogene coals with previous palynological studies (Table 2) does not show any correlation with monotaxic or stunted flora indicative of singularly ombrotrophic conditions, such as *Sphagnum*. The absence of clastic material in portions of the coal is not necessarily evidence for ombrotrophy in mires, since studies of modern rheotrophic mire systems indicate that flocculation of clays by low-pH standing mire waters and baffling by vegetation can significantly inhibit the transport of clastic material by currents into the mire center (Staub and Cohen, 1979). Given that much of the clastic material is smectitic, and likely of volcanic origin, the regional-scale trends in detrital mineral content most likely reflect several basin-scale pulses of airborne delivery and/or fluvial redistribution of volcanic material ultimately derived from the adjacent Sevier orogeny: a similar process envisaged for the deposition of the Cretaceous-Paleogene boundary claystone (Hildebrand, 1993; Nichols et al., 1992) and volcanic sediments preserved in coals elsewhere (e.g., Greb et al., 1999).

$\delta^{13}C$ Record of the Cretaceous-Paleogene Coals

The sampling at Knudsen's Coulee and Knudsen's Farm replicates the work of Therrien et al. (2007), and the $\delta^{13}C$ profiles at those localities largely corroborate both in terms of their shape and magnitude of excursions. The results of this study are also coarsely consistent with previous studies from terrestrial sediments elsewhere in the western interior of America that show $\delta^{13}C$ maxima immediately at, or some centimeters below, the Cretaceous-Paleogene claystone, and a negative excursion of 1.5‰–2.5‰ culminating in $\delta^{13}C$ minima above the Cretaceous-Paleogene claystone (Schimmelmann and DeNiro, 1984; Arens and Jahren, 2000; Beerling et al., 2001; Arens and Jahren, 2002; Gardner and Gilmour, 2002; Maruoka et al., 2007; Therrien et al., 2007; Grandpre et al., 2013). In this study, the position of the $\delta^{13}C$ minimum occurs within the Cretaceous-Paleogene claystone at Frenchman Valley, to 5–10 cm above the Cretaceous-Paleogene claystone at Knudsen's Coulee (Fig. 5). In those studies where sampling resolution is comparable to this study, the $\delta^{13}C$ minimum typically also occurs in the first 10 cm above the Cretaceous-Paleogene claystone (Beerling et al., 2001; Maruoka et al., 2007; Grandpre et al., 2013). In the sampled localities of this study, the $\delta^{13}C$ minimum above the Cretaceous-Paleogene claystone is followed by up to four more positive-negative excursions of lesser magnitude (P_2–N_2 to P_5–N_5; Fig. 5), which are not recognized or described in previous published studies. The magnitude of each excursion is variable between sampling localities, and the Wood Mountain Creek locality notably contains an additional positive-negative excursion within unit N_2 (Fig. 5).

Grandpre et al. (2013) questioned the use of terrestrial $\delta^{13}C$ curves alone as chemostratigraphic markers, as proposed by Arens and Jahren (2002). The successful correlation of these five positive-negative $\delta^{13}C$ cycles between the sampled sections over a distance of more than 500 km, between contemporaneous mires that were most likely disconnected, demonstrates that it is possible to resolve an intricate time series of atmospheric $\delta^{13}C$ from the terrestrial sedimentary record (Figs. 5 and 6). However, in this case, this is achievable by maintaining a high sampling resolution (~1 cm), and collecting contiguous samples from vertical sections through coal seams alone (and minor intercalated clastic sediments), as suggested by Therrien et al. (2007). It is proposed that such a detailed, regional-scale record of atmospheric $\delta^{13}C$ could not be produced from analysis of disseminated organic material from time-equivalent clastic floodplain deposits, where sediment deposition was more intermittent, and the total time represented by sedimentation may be relatively small, compared to the more continuous in situ accumulation of plant material in mires.

Peat Accumulation Rates and the Regional Record of Climate and Carbon Cycling

By estimating peat accumulation rates, it is possible to speculate upon the frequency of cyclicity in the petrological and stable isotopic record of the Cretaceous-Paleogene coals. The Cretaceous-Paleogene claystone horizon that occurs at five of the sampling localities (Figs. 4 and 5) has been shown through detailed palynological studies to mark the extinction of numerous Cretaceous flora (Sweet et al., 1999). The claystone representing the Cretaceous-Paleogene boundary was originally dated at ca. 65.16 Ma by Swisher et al. (1993), in the Hell Creek Formation (eastern Montana), while the SC-19 tuff at Knudsen's Farm and the proposed correlative smectitic mudstone (Figs. 4 and 5) were dated at 64.72 ± 0.17 Ma (Baadsgaard et al., 1988) and 64.90 ± 0.31 Ma (Eberth and Deino, 2005). More recently, Kuiper et al. (2008) recalculated these Cretaceous-Paleogene boundary ages by calibrating the $^{40}Ar/^{39}Ar$ dating method to astronomical tuning and redated the boundary to 65.84 ± 0.12 Ma and 65.99 ± 0.12 Ma. These new dates suggest the dates of Baadsgaard et al. (1988) and Eberth and Deino (2005) also require recalibration. Indeed, applying typical vertical compaction ratios from surface peat to coal (i.e., 1.2:1 to as much as 30:1; Ryer and Langer 1980) and the earlier radiogenic dates yields peat accumulation rates in the Cretaceous-Paleogene coals of up to 0.08 mm/yr. This rates appears unrealistic because it is orders of magnitude lower than the range of Holocene peat accumulation rates of 1–5 mm/yr compiled by Diessel et al. (2000).

Renne et al. (2013) more recently redated the Cretaceous-Paleogene boundary claystone, and two tuffs within the correlative IrZ and Z coal further south in the basin at Hell Creek, Montana (Fig. 1). These authors determined an age of 66.043 +0.011/–0.043 Ma for the Cretaceous-Paleogene boundary, broadly consistent with the age published by Kuiper et al. (2008) and Gradstein et al. (2012). The tuffs ("Z_2" and "Z_1"; sensu Renne et al., 2013), located 80 and 120 cm above the Cretaceous-Paleogene boundary, yielded ages of 66.019 +0.021/–0.046 Ma and 66.003 +0.033/–0.053 Ma, respectively. Although Renne et al.'s (2013) ages for these tuffs are consistent with stratigraphic order, they are indistinguishable statistically. However, taking into account the full range of peat-to-coal compaction ratios provided by Ryer and Langer (1980), and the maximum difference derived from the radiometric ages of Renne et al. (2013), peat accumulation rates could be up to ~0.4 mm/yr, consistent with the lower measured Holocene peat accumulation rates compiled by Diessel et al. (2000). The ages of Renne et al. (2013) therefore provide the most realistic estimate of peat accumulation rates at the time of the Cretaceous-Paleogene boundary and again suggest that the dates of Baadsgaard et al. (1988) and Eberth and Deino (2005) require revision.

Lateral variability in the thicknesses of units correlated between time lines in Figures 4 and 5 naturally implies that peat accumulation rates varied from locality to locality, and most likely with time. Assuming peat accumulation rates of 0.4 mm/yr (i.e., similar to the Hell Creek area of northeastern Montana) and peat to coal compaction ratios of 1.2:1–30:1, calculations imply that the Cretaceous-Paleogene coal at Ravenscrag Butte, where it is most thickly developed, may represent ~75 k.y. of peat accumulation. Thus, the three complete regionally developed vitrinite and inertinite cycles (Fig. 6) represent regional fluctuations in wildfire occurrence and oxidation in the peat at a centennial to millennial scale, similar to that recorded in the Pleistocene and Holocene (e.g., Marlon et al., 2012). Peak occurrence of wildfire or desiccation, as evidenced by peak inertinite content, occurs at the top of unit I_1 (Figs. 4B and 6B), more than 0.2 m above the Cretaceous-Paleogene boundary, which equates to ~14,000 yr later than the boundary event. The cause of long-term, regional changes in the inertinite composition of peat, if considered to be purely the product of charring (e.g., Scott et al., 2000; Scott and Glasspool, 2007), may be temperature, since studies of global pre-anthropogenic Pleistocene and Holocene charcoal abundance in sediment have been shown to vary with this parameter (Power et al., 2008; Marlon et al., 2012). Hence, the Cretaceous-Paleogene boundary event may have occurred in the early phases of a more than 14,000 k.y. regional warming event (unit I_1; Figs. 4B and 6B). This, and other short-duration warming and cooling events of the same order identified within the Ferris/Nevis coal (units I_1 to I_4; Fig. 4B) would be superimposed upon, or have quickly followed, the longer-term (~100–300 k.y. duration) cooling event that is considered to have characterized the latest Cretaceous (Wilf et al., 2003; Tobin et al., 2014). The interpretation that the inertinite record of the coal represents a proxy for relative temperature should be strongly tempered by the fact that (1) fire frequency is also controlled by rainfall patterns, amongst other parameters (McKenzie et al., 2004; Daniau et al., 2012), and (2) much inertinite may not be the product of charring, but may instead be the product of microbial and fungal activity associated with desiccation in the mire acrotelm (e.g., Diessel et al., 1992; O'Keefe et al., 2013). Hence, the inertinite record of coal may represent a proxy for rainfall in mires accumulating in nonparalic environments (for discussion, see Wadsworth et al., 2002). Whether inertinite is a proxy for temperature or rainfall, it should be noted that the trend of decreasing vitrinite composition and increasing inertinite composition begins before the Cretaceous-Paleogene boundary in the latest Maastrichtian, and other, similar cycles of vitrinite and inertinite content occur later in the Danian.

Similarly, the five positive-negative $\delta^{13}C$ excursions represent rapid cycles in atmospheric carbon budget, also at a centennial to millennial scale. The data from Rock Creek West and Wood Mountain Creek show the incipience of the first negative $\delta^{13}C$ excursion (unit N_1; Fig. 5) ~5 cm below the Cretaceous-Paleogene claystone, which predates the Cretaceous-Paleogene boundary by a maximum of a few thousand years. At the same two localities, organic $\delta^{13}C$ recovers to higher values within 0.3 m of the Cretaceous-Paleogene boundary (unit P_2; Fig. 5), or a maximum of ~21,000 yr after the event. Thus, the $\delta^{13}C$ anomaly at the Cretaceous-Paleogene boundary is consistent in magnitude, timing, and rapidity of onset and recovery with the existing terrestrial records of Renne et al. (2013) and with the recalibrated data of Arens and Jahren (2000), as well as some recent marine carbonate records (e.g., Schulte et al., 2010). The result is markedly similar to the terrestrial organic $\delta^{13}C$ record of Smit (1990), indicating negative excursion and partial recovery over just 3–5 k.y. Marine carbonate $\delta^{13}C$ records do typically show rapid initial negative excursion, but they typically have longer recovery intervals of ~300 k.y. to 3 m.y. (e.g., D'Hondt et al., 1998; Alegret et al., 2012), although D'Hondt et al. (1998) identified an early phase of recovery marked by the return of planktic-to-benthic $\delta^{13}C$ differences to low but relatively stable levels within the first several hundred thousand years after the extinction. The study therefore supports the notion that the surface ocean and atmosphere behaved as coupled reservoirs at this time, similar to other times in the Cretaceous (Gröcke et al., 2005), as opposed to a decoupled system (cf. Fang et al., 2013).

SYNTHESIS AND CONCLUSIONS

The Cretaceous-Paleogene boundary in the Canadian portion of the Western Interior basin occurred during the late lowstand to transgressive systems tract of a third-order (~3–6 m.y.) depositional sequence, and followed a 100–300 k.y. episode of regional cooling in temperatures at the end of the Maastrichtian. At the time of the Cretaceous-Paleogene boundary, peat accumulated in small, disconnected, rheotrophic mires, subject to ready influx of waterborne detrital clastic material of largely volcanic origin. The peatlands and overlying canopy were subject to cyclic fluctuations in wildfire and oxidation at a frequency of hundreds to thousands of years, in accord with observations of Holocene peatlands. The likely causes of these fluctuations were cyclic, regional-scale changes in temperature and/or rainfall. The Cretaceous-Paleogene boundary event occurred during a phase of gradually increasing temperature and/or decreasing rainfall, but peak wildfire and desiccation of peat occurred up to ~14,000 yr later than the Cretaceous-Paleogene boundary, and the mires did not experience significant wildfire activity or water stress in the immediate aftermath of the extinction event.

This study demonstrates that a meaningful, regional record of paleoatmospheric $\delta^{13}C$ can be extracted from the terrestrial sedimentary record. Although a persistent, 1.5‰–3.0‰ $\delta^{13}C$ excursion occurs across the Cretaceous-Paleogene boundary, it cannot be readily separated from four, further excursions later in the earliest Danian. Additionally, the level of microsampling required (at a resolution of less than perhaps 3 cm), the requirement for a demonstrably relatively continuous sedimentation rate, and the existence of other, lesser-magnitude isotopic excursions at the Cretaceous-Paleogene boundary suggest that the stable isotope record alone should be used with care as a chemostratigraphic tool in terrestrial settings (cf. Therrien et al., 2007; Grandpre et al., 2013). The negative carbon isotope excursion linked to the Cretaceous-Paleogene boundary preempts the event by a few hundred years, suggesting that the negative excursion at the Cretaceous-Paleogene boundary may have been overemphasized in terms of its relationship to whichever catastrophe led to mass extinction at this time. Recovery from the immediate negative excursion in $\delta^{13}C$ at the Cretaceous-Paleogene boundary was rapid (less than 21 k.y.), but the atmospheric carbon budget was subject to continued, cyclic fluctuations in isotopic composition at frequencies on the centennial to millennial scale. This confirms that the earliest Danian was also a period of major, cyclic transfer of carbon from one reservoir to another.

The similarity in magnitude of the $\delta^{13}C$ excursions and time scale of recovery between the terrestrial and shallow-marine realm suggest the two behaved as coupled reservoirs at this time. The absence of evidence for catastrophic change in the climatic regime at the time of the Cretaceous-Paleogene extinction, in the mires of the north-central Western Interior Basin, supports the notion that the negative shift in atmospheric $\delta^{13}C$ was driven by changes in the $\delta^{13}C$ composition of the surface ocean. This is consistent with the greater magnitude of extinction experienced by marine fauna relative to the terrestrial realm.

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