Sedimentological, Taphonomic, and Climatic Aspects of Eocene Swamp Deposits (Willwood Formation, Bighorn Basin, Wyoming)

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Alluvially deposited carbonaceous beds are a major source of leaf compression fossils, but the accumulation of both the sediments and the fossils is poorly understood. We have studied the sediments and fossil plant assemblages of three laterally extensive carbonaceous beds from the Eocene of the Bighorn Basin, Wyoming. Field, laboratory, and microscopic observations show that carbonaceous beds are highly heterogeneous, varying in the amount and type of organic matter, preservation of primary sedimentary structures, and the prevalence of paleosol features. Small-scale lateral and vertical variations indicate a mosaic of conditions in the wet floodplain backswamp. The variables controlling this mosaic were primarily rate of clastic input and degree of substrate flooding which, in turn, influenced the amount of pedogenesis.

Fossil plant assemblages occur in sediments least modified by pedogenesis, and the best preservation is in the distal, fine-grained portions of overbank flood deposits. There is no evidence for transport of plant material from channel to backswamp. Fossil leaves in a single quarry sample (i.e., those collected from within 10–20 cm stratigraphically and from an area of 4–10 m^2) probably accumulated during one or several closely-spaced flood events. Quarry samples are thus spatially and temporally constrained "snapshots" of the vegetation that grew on the alluvial floodplain. Each carbonaceous bed is estimated to represent roughly 2,000 years, based on the length of fluvial avulsion cycles. Multiple quarry samples from the same carbonaceous bed can thus be considered coeval for most evolutionary or paleoclimatic analyses.

In the central Bighorn Basin, laterally extensive carbonaceous beds are abundant in the lower and upper Willwood Formation, but essentially absent in the middle. We attribute this to tectonic and climatic change, especially to higher rates of sediment accumulation that caused greater dispersal of organic material within sediments and climatic drying that allowed degradation of organic material before it could be preserved.

INTRODUCTION

Laterally extensive carbonaceous beds are common in lower Cenozoic sequences of the western interior of North America, and are a major source of leaf compression fossils that provide data on floral composition, paleoclimate, and paleoecology. It has been argued that deposition of fossils in these distal alluvial settings is largely autochthonous, providing spatially and temporally constrained "snapshots" of the vegetation that originally grew on the alluvial floodplain (Wing et al., 1995; Wing and DiMichele, 1995). If this is true, then such fossil assemblages are similar to one another and to ecological samples of living vegetation from small areas. However, little is really understood about the formation of these backswamp deposits and the processes that preserve plant fossils within them. In particular, the deposition and preservation of plant fossils on a floodplain surface that supports vegetation presents an apparent paradox, as the preservation potential for leaves in soils is very low (Retallack, 1984). We need to understand the genesis of carbonaceous deposits in order to explain how plant fossils are preserved and not degraded in this setting.

The purpose of this paper is to better define the sedimentary setting and environmental conditions required for the formation of leaf compression assemblages by studying laterally extensive carbonaceous beds from the lower Eocene of the Bighorn Basin, Wyoming. We also evaluate the amount of time represented in a carbonaceous bed, the probability of spatial mixing of fossil leaves within such a deposit, and the type of vegetation represented by its fossil plant assemblages. In addition, we explore the implications of carbonaceous bed deposition for the tectonic and climatic evolution of the Bighorn Basin in the early Eocene.

GEOLOGICAL SETTING

The Bighorn Basin of northwest Wyoming (Fig. 1) is bounded by several structural areas that were elevated during the middle and latter parts of the Laramide orogeny (Bown, 1980). The major uplift of the northern Bighorn, Pryor, and Beartooth Mountains took place during the mid-late Paleocene (Bown, 1980; Beck et al., 1988). During the late Paleocene, uplift around the southern margin of the basin formed the southern Bighorn and Owl Creek Mountains, which were subsequently thrust southward in the mid-early Eocene.

Throughout the Paleocene and early Eocene, the Bighorn Basin accumulated predominantly fluvial sandstones, mudstones, lignites, and freshwater carbonates (Gingerich, 1983). Up to 3,000 m of drab (gray-brown colored) and lignitic sediments deposited during the Paleocene are assigned to the Fort Union Formation (Bown, 1980). Variegated, oxidized mudstones showing clear pedogenic features were deposited in the latest Paleocene and early Eocene, and are assigned to the Willwood Formation (Fig. 2; Bown and Kraus, 1981). Climates warmed

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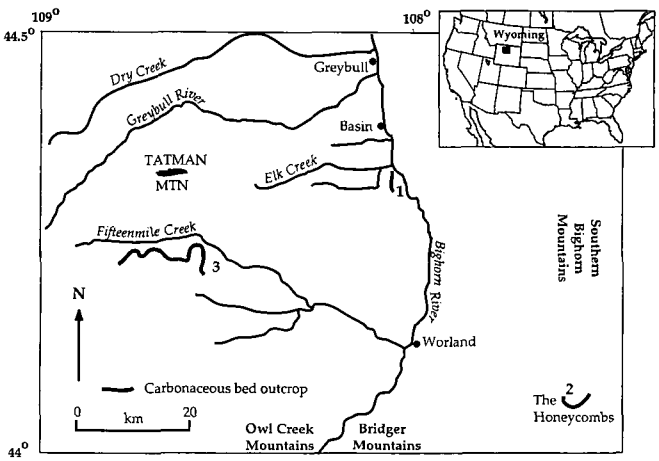

FIGURE 1—Location of Bighorn Basin and the three carbonaceous beds studied. 1: South Fork of Elk Creek carbonaceous bed, 2: Honeycombs carbonaceous bed, 3: Fifteenmile Creek carbonaceous bed.

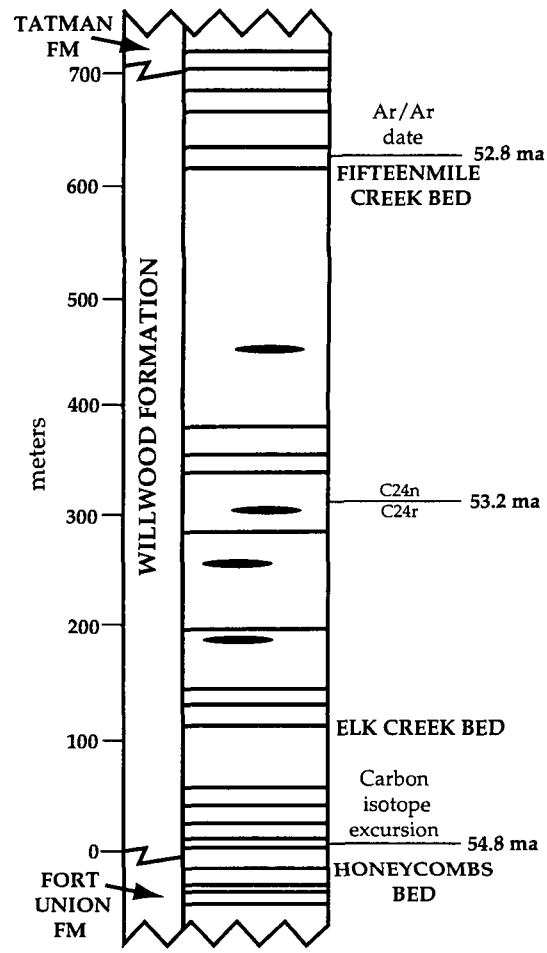

FIGURE 2—Generalized section of the upper Fort Union Formation and Willwood Formation in the Bighorn Basin showing stratigraphic positions of the carbonaceous shales studied with respect to isotopic stratigraphy, lithostratigraphy and magnetostratigraphy. Black horizontal lines in the column represent tabular carbonaceous shale units as described in this paper. Black lenses represent lenticular carbonaceous shale units as described by Wing (1984). Lithostratigraphy from Schankler (1980) and Wing (1984), Chron 24R/Chron 24N boundary from Tauxe et al. (1994), position of carbon isotope anomaly from Koch et al. (1995), radiometric date from Wing et al. (1991).

globally and locally during the late Paleocene, then appear to have cooled in the Bighorn Basin during the earliest Eocene, followed by renewed warming in the middle-late early Eocene, with a thermal maximum around 52–53 Ma (Wing et al., 1991; Wing, in press). The presence of mature, oxidized paleosols in the early Eocene reflects a warm temperate to subtropical climate with seasonally alternating wet and dry conditions (Bown and Kraus, 1981); warm winters are also indicated by physiognomic and floristic analyses of plant fossils found in the Willwood Formation (Wing and Greenwood, 1993).

Study Sites

We have studied three laterally extensive carbonaceous beds from the southern part of the Bighorn Basin (Figs. 3a–c). Beds in all areas are nearly horizontal (less than 3° dip). These beds are located at:

(1) The Honeycombs (KSDV 94 2.0): This bed is located within 20 km of the southeastern margin of the Bighorn Basin. Approximately 200 m of Willwood Formation and 100 m of the underlying Fort Union Formation are preserved in this area. The carbonaceous unit is stratigraphically about 10 m below the Wa0 red bed, which contains basal Wasatchian faunas (Wing, in press), and is exposed for about 3 km in an east-west direction and 1.5 km north-south. At its eastern end, the bed becomes coarser grained and eventually disappears into the subsurface. At the western end of the outcrop, the unit is cut out by channel sandstones and is preserved only as a series of lenses.

(2) Elk Creek (KSDV 94 1.0): This bed is located in the east-central part of the Big Horn Basin, in a relatively drab section of the lower Willwood Formation exposed along the south fork of Elk Creek. It is at the 112-m level of the Elk Creek section, and is associated with early Graybullian mammalian faunas (Schankler, 1980; Bown et al., 1994). This bed is approximately 0.7 m.y. younger than the Honeycombs bed, and was designated WCS 7 by Wing (1984). The bed is exposed for about 3 km in a north-south direction and about 1 km east-west. The bed dips below the surface at the southern and western ends of the outcrop area and is lost to erosion by the Elk Creek valley at the north end.

(3) Fifteenmile Creek (KSDV 94 3.0): This bed is located in the central Bighorn Basin along the south side of Fifteenmile Creek, in a drab interval of the upper Willwood Formation. It occurs at the 621-m level of the Elk Creek section, and is associated with early Lostcabinian mammals (Schankler, 1980; Bown et al., 1994). The Fifteenmile Creek bed is 13 m stratigraphically beneath a bentonitic tuff dated at 52.8 Ma (Wing et al., 1991), making it approximately 3 m.y. younger than the Honeycombs bed; it is the same bed as WCS 15 of Wing (1984). The Fifteenmile Creek bed is exposed east-west for a distance of 18 km and north-south for a distance of 3 km. At its westward end, the bed thins, becomes less carbonaceous, and is replaced by mudstones and siltier shales. Eastward it is intermittently cut out by a thick sheet sand, but is laterally continuous with a thin organic mudstone lying on top of a thick,

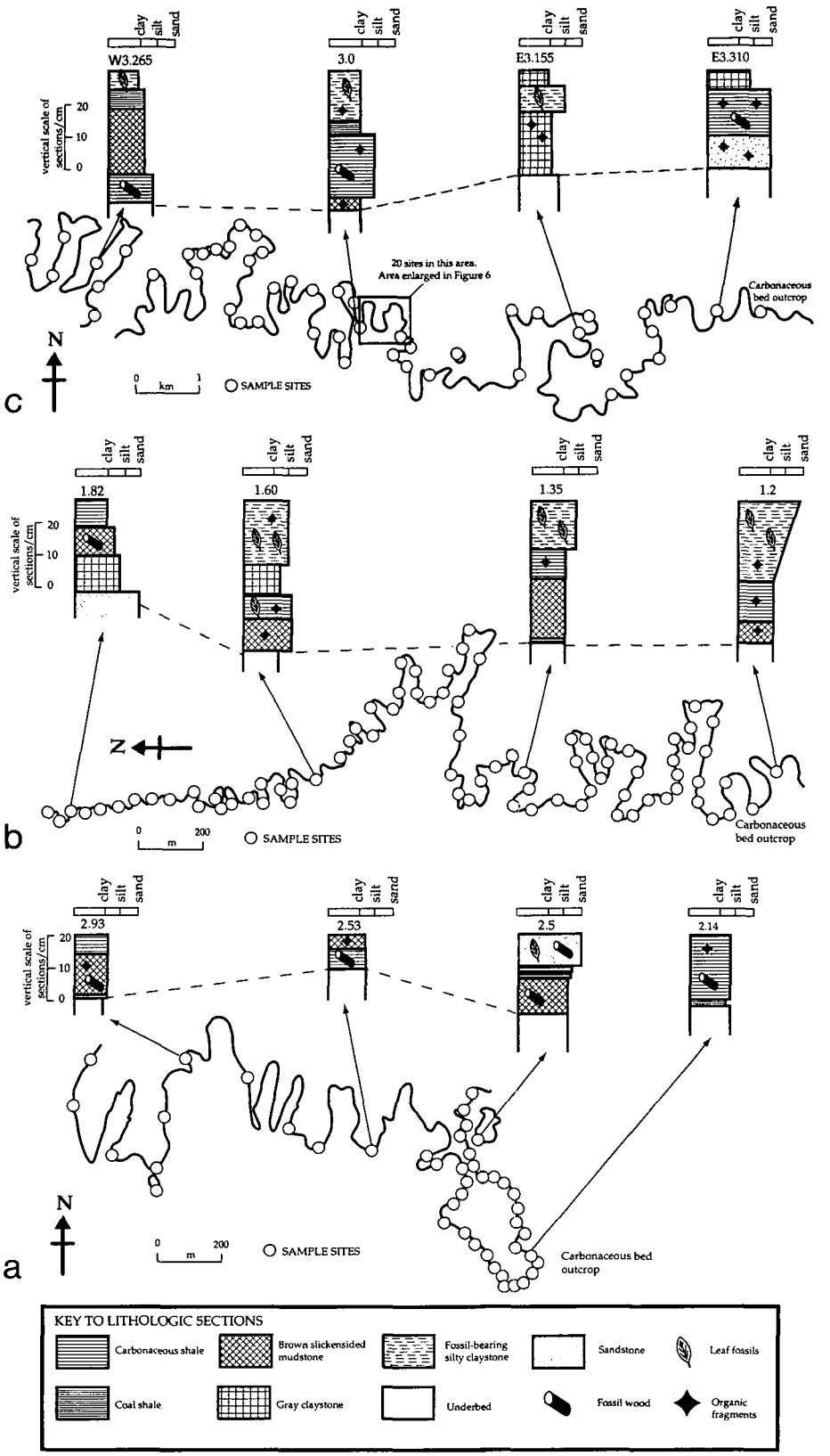

FIGURE 3—Outcrop maps of study area. (a) Honeycombs carbonaceous bed with selected detailed sections shown. (b) South Fork of Elk Creek carbonaceous bed with selected detailed sections shown. (c) Fifteenmile Creek carbonaceous bed with selected detailed sections shown. Area in square is enlarged in Figure 8.

purple, variegated mudstone that may represent the B-horizon of a hydromorphic soil.

Previous Work

Wing (1984) recognized two types of organic-rich units in the Willwood Formation—lenticular and tabular. The lenticular units were inferred to have been deposited in abandoned fluvial channels, and the tabular units were thought to represent deposition on wet, distal floodplains (backswamps). Wing (1984) described the tabular units as being composed of underbeds, carbonaceous shales, interlaminated silt and shale, and overlying, usually down-cutting, sand bodies. He also documented differences in the megafloral content between the carbonaceous shale and the interlaminated lithology. Similar work was carried out in the Fort Union Formation (Hickey, 1980).

Farley (1989) studied the relationship between palynofloral composition and lithological facies in the Willwood Formation. He recognized five facies in what he termed a "carbonaceous unit proper" which included beds overlying and underlying the fine-grained, organic-rich sediments discussed here. Farley also found significant variation in palynofloral composition that correlated with lithological type, although much of the variation was between silt/sand lithologies and clay/silt lithologies. Within the fine-grained portions of the carbonaceous unit, Farley recognized two subfacies: carbonaceous shale, which he attributed to a swamp environment; and brown mudstone, which he referred to a swamp-margin environment.

No previous studies have focused on the fine-grained portions of these tabular carbonaceous beds. The present study emphasizes the fine-grained lithologies not only because they comprise the bulk of the carbonaceous beds, but because they contain the majority of the leaf-fossil assemblages.

METHODS

Field Work

Field sites were selected at regular intervals along the outcrop of each carbonaceous bed. Sites were as close as one hundred meters apart where the outcrop was accessible, exceptionally thick, or of specific interest (for example near the lateral extremes of a deposit), and as far apart as 500 m when access to the deposit was limited or exposure poor. At the Honeycombs, fifty-five sections were described and sampled (Fig. 3a). These were concentrated at the eastern end of the deposit because the bed is inaccessible at its western end where it is exposed on high, steep bluffs. At Elk Creek, eighty-six sections were described and sampled at 100-m intervals (Fig. 3b). Seventy-one sections were dug along the Fifteenmile Creek carbonaceous bed (Fig. 3c). Along this more laterally extensive bed, sections were described and sampled every 500 m except near the middle of the deposit where 100-m sampling was carried out because of exceptional bed thickness, excellent exposure, and abundant fossil localities.

At each site, a section through the carbonaceous bed was cleared and the sedimentology, stratigraphy, and preservation of plant fossils were described in detail. General characteristics of the beds underlying and overlying the deposit were also noted. Lithological samples were collected from each recognizable subunit within a section at every other site along an outcrop as well as from the underlying bed. Eighty of these samples, representative of the three beds and their lithological subunits, were analyzed in the laboratory for grain size, clay mineralogy, weathering product mineralogy, and organic-carbon content. Nineteen were also thin sectioned for microscopic studies.

Laboratory Work

Grain size was determined using a Coulter counter LS-100R laser particle-size analyzer following treatment with hydrogen peroxide to remove organic material. Percentages of sand (>0.06 mm), silt (<0.06>0.02 mm) and clay (<0.02 mm) were calculated (Table 1 presents clay and silt data). Clay mineralogy was analyzed by X-ray diffraction using a Philips Norelco diffractometer with a copper target and nickel filter at settings of 48 Kv and 20 mA and a scanning speed of 1° 2Θ/minute (Moore and Reynolds, 1989). A series of heat and chemical treatments (Davies and Shaw, 1993) were employed to help distinguish the X-ray diffraction peaks of the clay minerals that were identified from tables (Chao, 1969). Weathering product mineralogies were assessed by X-ray powder diffraction using a Scintag diffractometer. Organic carbon content was calculated by low-temperature combustion methods (Wilde et al., 1979). Thin sections were made from well-lithified samples representative of all the lithological types observed, and from a small number of samples from overlying and underlying beds. These were assessed in terms of their micro-structure, birefringence fabrics (Bullock et al., 1985), organic particles, and alteration products.

RESULTS

Field Observations

Lithological variation was observed in all three beds. Each carbonaceous bed is composed of a complex arrangement of interfingering subunits of sediment that are 4–50 cm thick and extend laterally for tens to hundreds of meters and pinch in and out along outcrop (Figs. 3a–c). The vertical boundaries of these subunits tend to be distinct, whereas the lateral boundaries are generally gradational. The lithologies of these subunits include carbonaceous shale (sensu stricto), coal shale, brown slickensided mudstone, well-bedded silty claystone with fossils, gray clay, and laminated clay, as well as many intermediates. The subunits can be distinguished by transitions in grain size (dominantly silt to dominantly clay), structure (massive to well-bedded), color (light gray to brown-black), amount and type of organic matter (fragments, leaf mats, dispersed leaves, woody debris, coal), and amount and type of weathering products (natrojarosite, gypsum, and rarely goethite).

Some grain-size fluctuation can be observed along the outcrop of each carbonaceous bed. Coarser-grained sediments are derived from higher-energy depositional events, which may also reflect distance from the main channel. The beds also vary along outcrop in the number and arrangement of subunits and total bed thickness.

TABLE 1—*Summary of field and laboratory descriptions of typical lithological sub-units within carbonaceous beds.*

	Carbonaceous shale	Coal shale	Brown mudstone	Silty claystone with fossils	Gray claystone	Laminated claystone
Color	brown	brown-black	dark brown	light gray	light gray	light gray
Thickness	4–34 cm	6–25 cm	5–35 cm	12–42 cm	10–20 cm	6–15 cm
Structure	well bedded	fissile	massive, s/s	well bedded-fissile	massive	well bedded
Organics	leaf mats/ coal lenses	mct/dct/ con/eq	fragments/ fossil wood	dct/mct/con/eq/ woody axes/fragments	fragments/ fossil wood	dct/mct/con/eq
Upper contact	3, 4, 5	3, 4, 5	1, 4, 5	weathered at top	all	1, 3, 4, 5
Lower contact	3, 5, 7	1, 3, 5, 7	1, 5, 7	1, 3, 5	1, 3, 4, 6, 7	1, 2, 3, 5, 7
Weathering minerals	gypsum, natrojarosite, goethite	gypsum, natrojarosite	gypsum, natrojarosite	gypsum, natrojarosite, rare goethite	rare goethite, natrojarosite	goethite, natrojarosite
% clay + silt	67–100	74–100	78–100	88–100	no analysis	29–86
% clay	12–68	17–60	32–73	23–44	not assessed	36–86
% T.O.C.	5–75	43–81	5.4–22	5.8–8.5	5.0	6.3–9.9
Microstructure	homogeneous-laminated	not assessed	homogeneous	laminated	homogeneous	fining-up cycles laminated
B-fabric	uni-strial	no analysis	uni-strial/stip	mosaic	mosaic/grano	none
Other features	organics oriented	assoc. with carb shale		rare coarsening-up, syn-sedimentary deformation	assoc. with silty claystone	syn-sedimentary deformation

Key to characters: Structure: s/s = slickensided. Organic material: mct = monocotyledonous angiosperm fossils, dct = dicotyledonous angiosperm fossils, con = conifer, eq = *Equisetum*. Upper and lower contacts: 1 = with carbonaceous shale, 2 = with coal shale, 3 = with brown mudstone, 4 = with fossiliferous silty claystone, 5 = with gray claystone, 6 = with laminated claystone, 7 = with underlying bed. B-fabric: b-fabric = birefringence fabric, stip = stippled, grano = grano-strial.

Beds are not generally thickest in one area, but instead show a number of thicker pockets along their outcrops. These pockets are commonly associated with the greatest number and most complex arrangement of subunits. In addition to intrabed variation, there is also considerable difference among beds. The carbonaceous bed at Fifteenmile Creek shows the greatest thickness and most complex stacking arrangement of subunits (Fig. 3c). This bed is also the most extensive of the three. In its east-central outcrop area, it can be as much as 2.5 m thick, but varies greatly on either side of this. It is coarsest at its eastern end and fines towards the west, but shows much variability. The Honeycombs carbonaceous bed (Fig. 3a) is thin (in places only 20 cm) with few subunits present in any one section. It did not yield many well-preserved leaf fossils, except at its coarser, more easterly end. The Elk Creek carbonaceous bed (Fig. 3b) is generally about 0.5–0.75 m thick and shows an overall trend from coarser grained in the north to fine grained in the south. It yielded many well-preserved fossil-leaf assemblages in two different lithological types. In all three study areas, the bed underlying the carbonaceous unit was very similar: poorly-lithified, massive, bluish-gray, and varying in grain size along strike from clay to mixed clay, silt, and sand. Beds overlying the carbonaceous units varied from massive silty claystones to small sheet and wedge-shaped sandstones.

Laboratory Observations

Table 1 summarizes all field, laboratory, and microscopic results. In general, carbonaceous beds are fine grained, varying from 12–86% clay and 67–100% clay plus silt (Table 1). Clay minerals tended to flocculate during analysis, even after the removal of all organic particles, so these results may be an underestimate of the finer-grained fraction. The total organic carbon (TOC) of the samples varied from 5–81%. However, most had a TOC content of less than 20% (Table 1). Clay mineral analysis revealed that all samples had the same mineralogy of kaolinite, illite, and smectite. Proportions of these minerals were not calculated owing to the poor reproducibility of clay mineral proportionality on the equipment used and the low accuracy of methods available to calculate this proportionality (Pierce and Siegel, 1969).

Microscopic Observations

Thin sections revealed differences between samples from different types of subunits. Microstructure varied from completely homogeneous to laminated. Birefringence fabrics were present in fine-grained samples, where they ranged from randomly oriented (mostly mosaic type), to linear (mostly uni-strial type). Organic material varied in abundance, shape, and form, from rare, large, amorphous masses to many small fragments. Orientation of organic particles was also variable, and only within well-bedded sediments and fossil-bearing silty claystone did they define laminae. Weathering products were generally amorphous and ranged from completely absent to pervasive.

Typical Lithologies

Although variation in the lithological features in carbonaceous beds is more or less continuous, in Table 1 we list the characteristics of typical subunits in order to illustrate the kinds of features that occur together. These descriptions can be thought of as central tendencies in the range of lithological variation. These subunits are carbonaceous shale (sensu stricto), coal shale, brown mudstone [these three lithologies were referred to as the swamp facies by Farley (1989)], fossil-bearing silty claystone, gray claystone, and laminated claystone. All of these subunits are

common apart from the laminated claystones, which occur as rare, distinct lenses that are 1 to 250 m in lateral extent.

ENVIRONMENTS OF DEPOSITION

Statistical Analysis of Samples

The lithological variation can also be recognized in an ordination analysis of samples based on their grain size, sedimentary structures, organic content, color, fossil content, weathering products, and birefringence fabric. Correspondence analysis was carried out using nineteen characteristics of 55 samples chosen to represent a range of lithologies seen within the carbonaceous beds (Table 2). A plot of axis I against axis II sample scores shows fairly continuous variation from brown mudstone samples with high axis I scores, through carbonaceous shale and coal shale samples near the center of the ordination, to more clastic claystone or siltstone samples containing well-preserved leaf fossils that have very negative scores on axis one (Fig. 4a).

The corresponding plot of sample characteristics reflects the presence of two gradients (Fig. 4b). Higher axis I scores correlate with greater degrees of pedogenesis as measured by slickensides and mosaic birefringence, whereas negative axis I scores correlate with greater preservation of primary stratification and linear types of birefringence. Axis II separates more organic-rich samples from less organic-rich samples. Samples with lower amounts of total organic carbon can be either highly affected by pedogenesis, presumably owing their low amounts of organic carbon to biodegradation, or very little affected by pedogenesis, presumably having little organic carbon because they were deposited by an influx of water carrying mostly clastic sediment.

Samples from sedimentary subunits that contain well-preserved leaf fossils group with other samples that are relatively low in organic carbon and have well-preserved sedimentary microstructure. Thus, high-quality preservation of leaf assemblages, even within the overall context of a wet and organic floodplain, is strongly associated with clastic depositional events and the absence of pedogenesis. Apparently the Eh and pH conditions on these wet floodplains were moderate enough to allow partial recycling of litter, except when sedimentary events temporarily overwhelmed the system.

Inferred Environments

The different types of sedimentary subunits in each tabular organic bed formed in response to a combination of minor variations in floodplain topography, edaphic conditions, and the influx of sediment associated with overbank flooding. Carbonaceous shale, *sensu stricto,* is interpreted to have formed in swampy areas where the rate of clastic sedimentation was slow enough to allow mats of leaf material to build up. Mats were partially preserved because anaerobic, mildly acidic conditions inhibited plant rooting and animal activity on and within the substrate. The oxygen-reduced state allowed formation of pyrite (now oxidized to natrojarosite), a situation that is commonly encountered today in swampy regions where peat is accumulating (Jakobsen, 1988). The coal shale formed under similar conditions but in areas where sediment influx was minimal. The environments in which carbonaceous shales, *sensu stricto,* and coal shales formed may have been colonized by plants that had adaptations to survive waterlogging such as laterally extensive rooting systems (Jenik, 1976). These roots did not penetrate far into the substrate and, therefore, provided a source of organic material to the substrate with little destruction of original sedimentary structures by bioturbation. Brown mudstones formed in more oxic environments that permitted plant colonization, more deeply penetrating root systems, and degradation of plant material. Slickensides, especially those associated with smectite, probably indicate soil formation. Soil horizonation is not observed, which indicates that this was a very immature paleosol, equivalent to an entisol. Brown mudstones formed in areas above or close to the water table that received sediment at rates low enough to allow complete homogenization prior to burial. Some brown mudstones may represent partially oxidized and root-bioturbated carbonaceous shale.

The silty claystones that contain fossils formed during influxes of fine-grained sediment as distal overbank-flood deposits during periods of high discharge in the main channel. Bedding and preservation of leaves is due to rapid sediment deposition as well as the prevailing weak acidity and anoxia of the backswamp. Lateral gradations from silty claystone with fossils to gray clay have been observed. This implies that the gray clay lithology formed when fossil-bearing silty claystones were colonized and homogenized by early successional plants such as *Equisetum,* a common fossil in these sediments.

Laminated claystone lenses formed in small floodplain ponds. Their fining upward sequences record multiple flood events when fine clastic material settled from suspension in standing water. They may have formed in areas of the backswamp that were connected to active channels by a permanent inlet that supplied oxygen and sediment-saturated water via periodic influxes (Pelzer et al., 1992). Lenses may also have formed in depressions on the floodplain. They are uncommon, which means that either this was an uncommon environment or that lenses were often bioturbated, resulting in the formation of gray claystones. These lenses are distinct from the "Type I" lenticular beds described by Wing (1984) in that they are thinner, not underlain by sandstone, not unconformable with the underlying sediments, and are found within tabular carbonaceous units.

MODEL FOR THE DEPOSITION OF TABULAR CARBONACEOUS BEDS

Although each carbonaceous bed may have formed under slightly different circumstances, similarities in the lithologies and lithological relationships among all of the beds have been used to develop a single model that can explain most of their features. The environmental conditions responsible for the deposition of carbonaceous beds are described below and illustrated in Figure 5.

Prior to carbonaceous bed deposition, the distal floodplain was wet but still experienced some fluctuation in water table. Pedogenic modification of this surface led to the development of a distinctive blue-gray "underbed" with

TABLE 2—Samples by characteristics matrix used in the correspondence analysis.

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
1-44D	0.25	0.52	0.07	2	1	1	0	1	1	1	0	0	1	1	0	0	1	0	1
1-64C	0.29	0.47	0.07	3	1	1	0	0	1	1	0	0	1	1	1	0	1	0	1
1-84B	0.37	0.63	0.10	3	1	1	0	1	0	0	0	0	0	0	0	1	1	0	1
1-85iB	0.29	0.57	0.08	2	0	0	0	1	0	1	0	1	0	1	0	1	1	0	1
2-viB	0.68	0.32	0.08	2	1	1	0	0	0	0	0	1	1	0	1	0	1	0	1
2-16B	0.43	0.41	0.13	1	1	1	0	0	0	1	0	1	0	0	0	0	1	0	1
2-28D	0.34	0.56	0.11	3	1	1	0	0	1	1	1	1	1	1	1	0	1	0	1
2-53B	0.49	0.51	0.12	3	1	1	0	0	0	0	1	1	1	0	1	0	1	0	1
2-118B	0.37	0.30	0.20	4	1	1	0	0	0	1	1	0	1	0	1	0	1	0	1
W3-225E	0.18	0.60	0.13	3	1	1	0	0	0	1	1	0	1	0	1	1	1	0	1
W3-355B	0.22	0.73	0.77	4	1	1	0	0	0	0	1	1	1	1	0	0	1	0	1
W3-90F	0.22	0.46	0.56	3	1	1	0	0	0	1	1	0	1	1	0	0	1	0	1
W3-20D	0.31	0.59	0.50	3	1	1	0	0	0	1	1	0	1	1	0	0	1	0	1
E3-125B	0.12	0.72	0.76	4	1	1	0	0	0	0	1	0	1	1	1	1	1	0	1
2-113B	0.60	0.40	0.43	1	1	1	0	0	0	0	1	1	1	1	0	1	1	0	1
2-108B	0.40	0.56	0.45	4	1	1	0	0	0	0	1	0	1	0	1	0	1	0	1
E3-290C	0.18	0.77	0.70	4	1	1	0	0	0	0	1	1	0	1	0	1	1	0	1
E3-90D	0.18	0.68	0.53	4	1	1	0	0	0	1	1	0	1	1	0	0	1	0	1
E3-18B	0.18	0.80	0.81	4	1	1	0	0	0	0	1	0	1	1	0	0	1	0	1
W3-108B/C	0.28	0.67	0.46	4	1	1	0	0	0	0	1	0	0	1	0	0	1	0	1
W3-150F	0.17	0.73	0.58	4	1	1	0	0	0	0	1	0	1	1	0	0	1	0	1
W3-330D	0.37	0.63	0.53	4	1	1	0	0	0	0	1	0	1	1	0	0	1	0	1
1-2C	0.45	0.52	0.06	2	1	1	0	0	0	1	0	0	1	1	0	0	0	0	0
1-20B	0.38	0.59	0.09	3	0	0	1	0	0	0	1	0	1	1	0	0	0	0	0
1-82C	0.43	0.47	0.05	3	0	0	1	0	0	0	0	1	0	1	0	0	0	0	0
1-38B	0.50	0.48	0.09	2	0	0	1	0	0	1	0	0	0	1	0	0	0	0	0
1-60B	0.40	0.56	0.07	3	0	0	1	0	0	1	0	0	1	1	0	0	0	0	0
2-viiB	0.47	0.37	0.09	1	0	0	1	0	0	1	0	1	1	0	1	0	0	0	0
2-22B	0.55	0.45	0.11	3	0	0	1	0	0	0	0	0	1	0	1	0	0	0	0
2-63C	0.64	0.36	0.07	2	0	0	1	0	0	1	1	1	1	0	0	1	0	0	0
2-90C	0.56	0.22	0.07	2	0	0	1	0	0	0	1	1	0	0	0	0	0	0	0
2-108C	0.62	0.30	0.07	2	0	0	1	0	0	0	1	1	1	0	0	0	0	0	0
W3-330C	0.73	0.27	0.21	1	0	0	1	0	0	1	0	1	1	0	1	0	0	0	0
W3-150C	0.48	0.50	0.14	2	0	0	1	0	0	0	1	1	1	0	1	1	0	0	0
W3-20C	0.48	0.50	0.06	1	0	0	1	0	0	1	0	0	0	0	1	0	0	0	0
E3-6B	0.39	0.55	0.07	2	1	1	0	0	0	1	0	1	1	1	0	0	0	0	0
E3-18C	0.41	0.55	0.22	3	0	0	0	0	0	1	1	1	0	1	0	0	0	0	0
E3-110D	0.32	0.68	0.15	3	1	0	0	0	0	0	1	1	0	1	0	1	0	0	0
E3-290B	0.32	0.55	0.07	2	0	0	0	0	0	1	0	0	0	0	1	1	0	0	0
1-2D	0.44	0.56	0.07	1	1	1	0	1	1	0	0	0	0	1	1	0	0	0	0
1-30C	0.23	0.71	0.06	1	1	0	0	1	1	1	0	0	1	1	1	0	0	0	0
1-66D	0.38	0.62	0.06	1	1	1	0	1	0	0	0	0	1	1	1	0	0	0	0
1-80B	0.36	0.55	0.07	2	1	1	0	1	1	0	0	0	1	1	0	0	0	0	0
E3-2C	0.27	0.73	0.09	2	1	1	0	1	0	1	0	1	0	0	0	0	0	0	0
W3-110G	0.27	0.61	0.08	2	1	0	0	1	1	1	1	0	1	0	0	0	0	0	0
W3-355C	0.47	0.53	0.08	2	1	0	0	1	0	0	1	0	0	1	0	0	0	0	0
1-2B	0.29	0.52	0.08	2	0	0	0	1	0	0	0	0	1	0	0	0	1	1	0
1-24B	0.36	0.46	0.07	1	0	0	0	1	0	0	0	0	1	0	0	0	1	1	0
1-48C	0.32	0.55	0.08	1	1	1	0	1	1	1	0	0	1	1	0	0	1	1	0
1-74D	0.37	0.60	0.07	1	1	1	0	0	0	1	0	0	1	0	0	0	1	1	0
2-22C	0.64	0.36	0.10	1	0	0	0	0	0	1	1	0	0	0	1	0	1	1	0
2-68B	0.56	0.36	0.06	1	0	0	0	0	0	0	0	1	0	0	1	0	1	1	0
E3-290B	0.08	0.86	0.08	2	1	0	0	0	0	1	1	0	1	1	0	0	1	1	0
E3-8C	0.38	0.56	0.07	1	1	0	0	1	0	0	0	1	1	0	0	0	1	1	0
W3-10F	0.35	0.61	0.08	1	1	0	0	1	1	1	0	1	1	0	0	0	1	1	0

Key to characters: 1 = % clay, 2 = % silt, 3 = % organic carbon, 4 = color, 5 = bedding, 6 = fissility, 7 = slickensides, 8 = fossils of trees, 9 = fossils of herbs, 10 = organic fragments, 11 = coal, 12 = wood fragments, 13 = lower boundary, 14 = natrojarosite, 15 = goethite, 16 = gypsum, 17 = thin section bedding, 18 = fining up cycles, 19 = linear birefringence. For all binary characters: 1 = present and 0 = absent. For lower boundary: 1 = sharp, 0 = gradational. For color: 1 = light gray, 2 = gray-brown, 3 = brown, 4 = black.

characteristics of a gley soil (Davies-Vollum, 1996). This underbed was coarser-grained proximal to the main channel, finer-grained in distal areas, and had little relief.

Carbonaceous bed deposition commenced when the local water table rose as a result of subsidence, climatic change, or both. The increase in water table reduced oxygen levels in the substrate, leading to the preservation of organic matter. Carbonaceous shales (sensu stricto) and coal shales formed under these dysoxic and mildly acidic conditions (pH 5–6 is indicated by the formation of pyrite

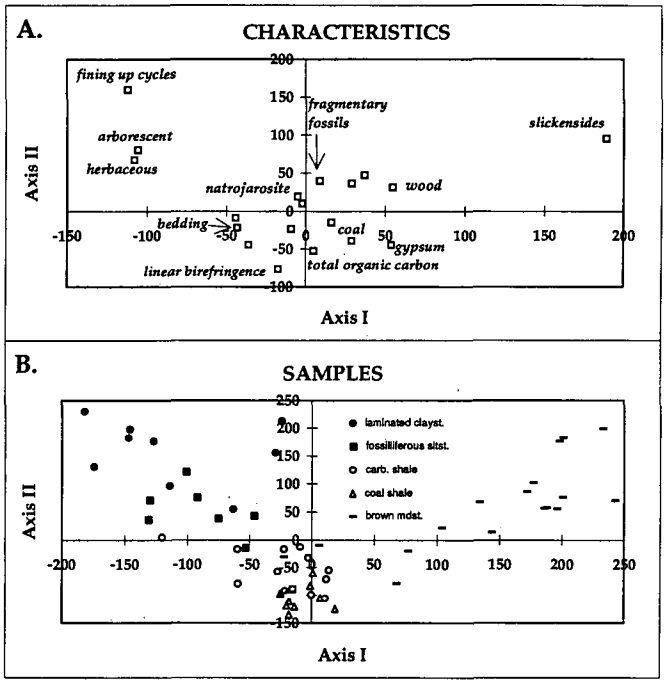

FIGURE 4—Correspondence analysis of 55 lithological samples chosen to represent lithological variation in the carbonaceous beds. (A) Plot showing the relationships of lithologic characteristics used to describe the samples plotted. (B) Plot showing the positions of the samples. The key in (B) identifies the lithologies to which the samples were assigned in the field.

in the sediments, although the pyrite has subsequently been weathered to natrojarosite). Plants adapted to waterlogged substrates were able to colonize these areas. The slightly higher and drier areas of the distal floodplain were more vegetated, or at least more deeply rooted, and were bioturbated to form brown mudstones. Palynofloral studies of the brown mudstones and carbonaceous shales show that the vegetation growing on these two types of substrate was slightly different in taxonomic composition (Farley, 1989, 1990). As carbonaceous sediment accumulated to near the level of the water table, additional areas were colonized and bioturbated. Leaves from the vegetation that grew on these immature, waterlogged soils accumulated within the nearby carbonaceous and coal shale-forming environments or were degraded in the soil.

The accumulation of organic-rich sediment was periodically interrupted by the distal portions of high-discharge flood events that brought fine-grained sediment to the lowest parts of the floodplain. These clastic influxes preserved leaves local to the backswamp environment in less organic-rich sediment and formed fossil-bearing silty claystones. Some exposed flood deposits were colonized by early successional plants and bioturbated to form gray claystones. Small ponds also developed in the lowest areas and were filled by periodic influxes of sediment-saturated water that facilitated preservation of leaves and formed laminated claystones.

At any one time, the distal floodplain would have been a mosaic of environments and vegetation types: fresh splay surfaces either lacking vegetation or covered by early successional plants; wetter areas of swamp vegetation growing on immature soils; even wetter patches of deep swamp vegetation in which carbonaceous shale was forming; and small areas of open water. Individual patches in this mosaic, now represented by sedimentary subunits in the carbonaceous bed, were on the order of tens to hundreds of meters across.

During the later phases of carbonaceous bed deposition, clastic sedimentation in distal areas became more common as the main channel approached avulsion. Major floods submerged the floodplain, depositing fine-grained material in the backswamp as fossil-bearing silty claystones. These fine-grained sediments incorporated locally derived leaves similar to those found in carbonaceous shale.

As the avulsion of the channel was achieved, sandstone bodies with channel features were deposited directly above the carbonaceous bed in some areas. In other areas, interlaminated siltstones and sandstones were formed, which preserved leaves of riparian plants (Wing, 1984). The influx of coarser-grained sediments loaded and squeezed the saturated carbonaceous sediment beneath them, causing its injection into beds above, and deformation of structure and bedding. In other areas, drab-colored mudstones were deposited above the carbonaceous unit, reflecting floodplain substrates that were oxygen depleted, but too biologically active for preservation of large amounts of organic material.

Kraus and Aslan (1993) proposed that avulsion cycles within the Willwood Formation were induced by climatic cycles associated with Earth's precession, and that wetter climates occurred just prior to channel switching towards the end of each avulsive cycle. We suggest that development of carbonaceous beds occurred towards the end of a fluvial cycle, just prior to avulsion, when distal areas of the alluvial floodplain were stable in a fashion similar to that described for the Saskatchewan River (Smith et al., 1989). A period of increased precipitation prior to channel switching could have caused the water-logging that induced the onset of swampy conditions and deposition of each carbonaceous bed.

Carbonaceous beds of the Willwood Formation show relationships to other fluvial sediments that are similar to those seen in floodplain swamps of the Saskatchewan and other smaller rivers in Alberta, Canada (Smith et al., 1989; Smith and Smith, 1980), and the Magdalena River, Colombia (Smith, 1986).

TEMPORAL AND SPATIAL DIMENSIONS OF PLANT FOSSIL ASSEMBLAGES

Time Represented

Kraus and Aslan (1993) estimated that a time span of 20,000 years separated vertically successive avulsion deposits in the Eocene of the Bighorn Basin. If we are correct that carbonaceous deposits were deposited in a brief period just prior to a major avulsive event, it is probable that each existed for a small portion of the 20,000 year period, perhaps 1,000–3,000 years. Each sedimentary subunit containing a fossil plant assemblage would have been deposited in only a fraction of this time, probably just one or a few sedimentation events occurring over a season to a few years. Data from fossil sites along a carbonaceous bed

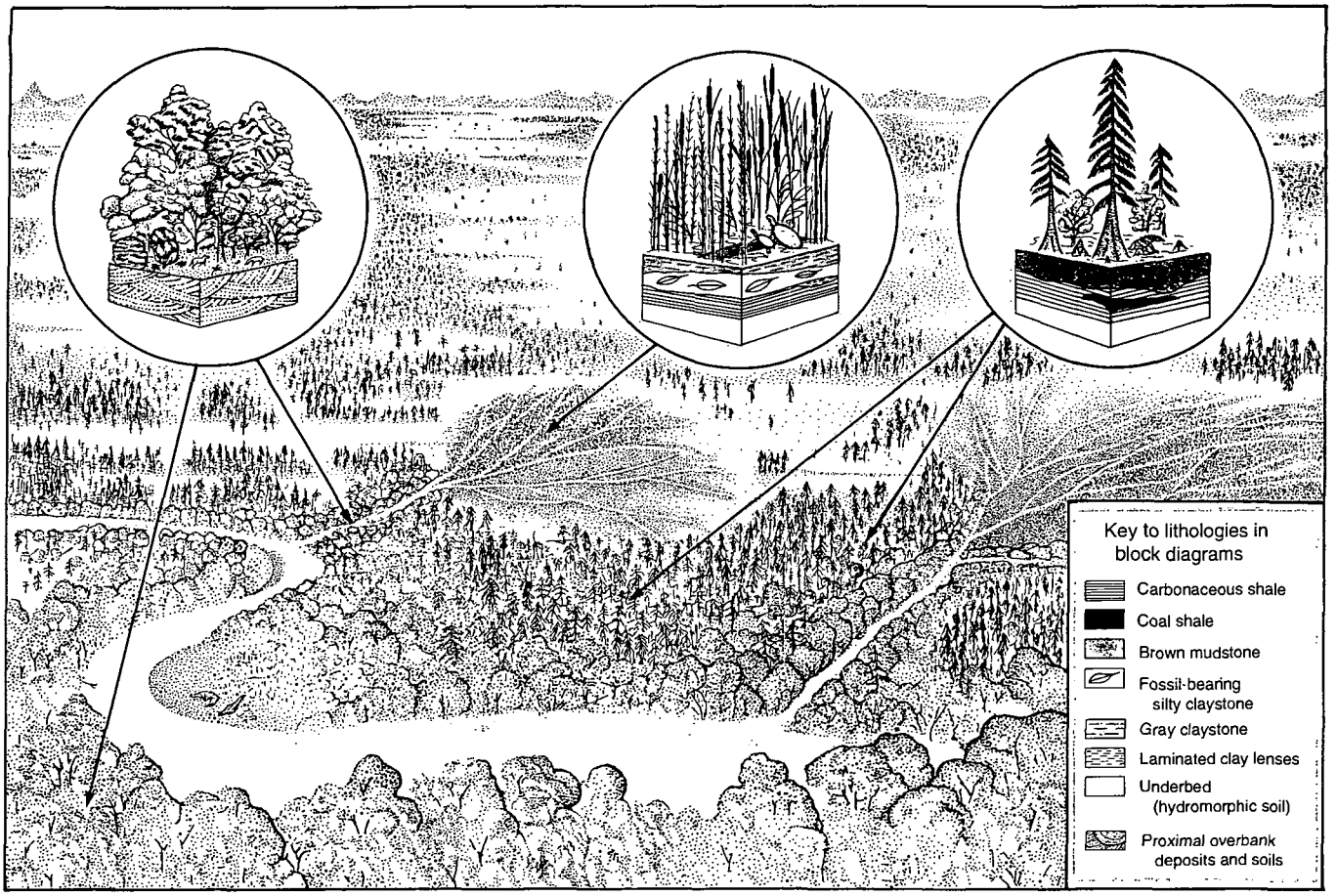

FIGURE 5—Reconstruction of the Eocene floodplain landscape showing the mosaic of backswamp environments and vegetational types. Inserts depict vegetation types prevalent on areas indicated by arrows, their scale is given by animals shown in inserts.

can be aggregated (analytically time-averaged in the sense of Behrensmeyer and Hook, 1992) for studies that do not need to resolve events in an ecological time frame. Individual sites potentially could be used to study floral changes related to local ecological processes.

Spatial mixing

Different types of sediment within the fine-grained portion of a carbonaceous bed vary with respect to the states and amounts of preservation of plant fossil material but, in general, contain assemblages of similar floristic composition. In contrast, those from overlying, interbedded silt and sandstone have more riparian species (Wing, 1984). Taxa that are common in channel or proximal crevasse-splay sediments are extremely rare in the fine-grained carbonaceous sediments, indicating minimal incorporation of plant material from riverside areas into backswamp environments, even during flood events. Fossil leaf assemblages in carbonaceous beds reflect backswamp vegetation regardless of whether they occur in carbonaceous shale (in the strict sense) or more clastic-rich sediment.

Transport of identifiable plant remains between different floodplain environments appears to be very rare, yet there are commonly strong differences in floral composition even among closely-spaced quarry sites that sample the same fine-grained sedimentary subunit. Eleven quantitative floral samples were taken from the Fifteenmile Creek carbonaceous bed to document variation in floral composition over a small area (Fig. 6). All of these floras were derived from the same microstratigraphic interval within the bed (150–175 cm above the underbed), all come from similar lithologies, and the sites are within 250 m of each other. All of the floral samples are probably derived from the same sedimentary event or closely-spaced series of sedimentary events. Together they contain 3,612 leaves assigned to 43 species.

Despite these samples being nearly or actually coeval and derived from a very small area, there is substantial variation in floral composition (Table 3). Are the differences in sample composition greater than would be expected from a series of samples drawn randomly from a homogeneous leaf bed? To assess this possibility we calculated the standard deviation of the mean number of leaves of each species per sample assuming binomial error. The standard deviation (SD) around the mean is given by:

$$SD = 1/(p+q) * \sqrt{\{pq(p+q-N)/(N(p+q-1))\}}$$

where p = the number of leaves in the combined samples assigned to a given species, q = the number of leaves in the combined samples not assigned to that species, and, N = the number of leaves in the sample for which the confi-

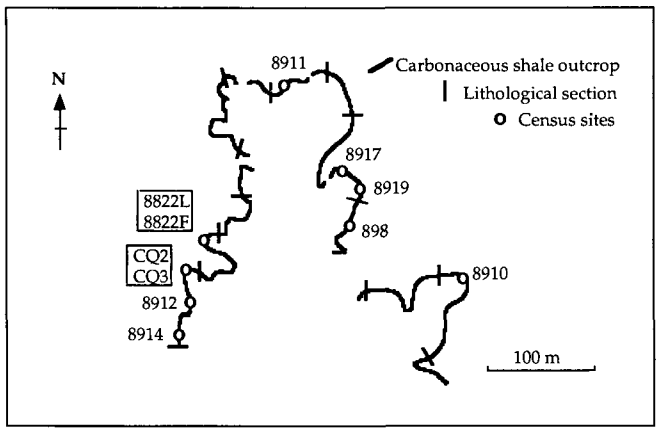

FIGURE 6—Map of the outcrop of a portion of the Fifteenmile Creek carbonaceous layer showing the positions of collecting sites. This corresponds to the boxed area in Figure 3c. Note that all floral samples were collected within about 200 meters of one another.

dence intervals are being calculated (Wilf, in press). Note that, because standard deviations are calculated separately for each species in each sample, a value that is more than four standard deviations from the mean in a large sample (e.g., 24 leaves of *Dombeya novi-mundi* at locality 8910) may not be so significant in a small sample (e.g., 24 leaves of *Dombeya novi-mundi* at locality CQ2).

Table 3 shows that each of the eleven sites has at least one species that differs from its mean abundance by more than four standard deviations (values underlined in Table 3). Furthermore, all of the species that are not rare have more variation in their abundance among samples than can be explained by simple sampling error. Particularly striking are abundant and ubiquitous taxa such as *Glyptostrobus* and *Platycarya,* which are absent at one site each, and two taxa that are present at only one site but are nevertheless abundant where they occur (*Equisetum* and sp8911-4). Because we have excluded temporal and depositional effects by confining the sampling area and lithology, and have demonstrated that sampling effects alone are not large enough to account for the differences in sample composition, the most likely remaining hypothesis is that the differences in sample composition reflect small scale variation in the composition of the original wet floodplain-leaf litter. Variation in litter composition probably reflects the proximity of different sites to individual trees of varying species. This is consistent with the idea (derived from analysis of the sediments alone) that the floras are essentially autochthonous, and with the observation of similar heterogeneity in untransported leaf litter under living forests (Burnham, 1994).

If the samples analyzed above are typical of leaf assemblages from laterally extensive carbonaceous beds, and our observations suggest they are, then fossils from such environments are samples of vegetation that are quite analogous to one another and to forest-litter accumulations. Hence leaf-compression assemblages from carbonaceous beds can be used to track changes in site diversity and small-scale vegetational heterogeneity across geological time.

CARBONACEOUS BEDS AND BASIN EVOLUTION

Tabular carbonaceous beds require stable, virtually year-round wet environments to form. If there is any substantial dry season, organic material that accumulated during the wet season is degraded and a drab colored, hydromorphic soil forms (Retallack, 1990). High water tables can be achieved through subsidence, increased precipitation, or decreased seasonality in precipitation. Important locally are substrate type and floodplain topography, which can cause small-scale variation in edaphic conditions.

Tabular carbonaceous beds compose 20% of stratal thickness within the bottom and top 100 m of the Willwood Formation (0–100-m/650–750-m intervals) but are rarer in strata between these levels (Fig. 2; Wing, 1984). The prevalence of carbonaceous beds covaries with sediment-accumulation rates in the central Bighorn Basin. Rates of sediment accumulation for the Bighorn Basin have been calculated by relating measured sections to the geomagnetic reversal time scale through combined isotopic, magnetostratigraphic, and biostratigraphic data (Tauxe et al., 1994; Koch et al., 1995; Wing, in press). Paleosol maturity has also been used to identify periods of faster or slower accumulation (Bown and Kraus, 1993; Kraus and Bown, 1993). All methods imply low sediment accumulation in the earliest Eocene (350 m in approximately 1.7 m.y.), followed by a brief period of much higher rates in the mid-early Eocene (300 m in approximately 0.5 m.y.), followed by a return to lower rates in the last half of the early Eocene (350 m in approximately 2.5 million years). It has been suggested that the sediment-accumulation curve for the Willwood Formation reflects the infilling of space created by earlier thrusting along the Oregon basin fault and subsidence due to the redistribution of tectonic load during Eocene erosion and deposition (Kraus and Bown, 1993). In the southern part of the basin sediment-accumulation rates have been specifically related to southward thrusting of the Owl Creek Mountains in the early Eocene to their maximum structural elevation in the late early Eocene (Bown and Kraus, 1993).

Organic material in carbonaceous beds is preserved because of the physico-chemical conditions of the distal backswamp environment, which limit biodegradation, and low rates of clastic accumulation, which allow organic material to be a major component of the bed. Increase in clastic influx alone could be a factor in the absence of tabular carbonaceous beds in the middle part of the Willwood Formation. However, not only are tabular carbonaceous beds absent from the 350–600-m interval of the Willwood Formation, but plant fossil material is confined to isolated abandoned channel deposits. Increased sediment-accumulation rates alone probably do not explain the rarity of preserved organic material in this part of the section. On the contrary, higher sedimentation rates should protect leaf fossils from biodegradation, so plant fossils should be preserved as compressions in some floodplain sediment types.

The absence of plant fossils in floodplain sediments of the middle Willwood Formation reflects the degradation and reworking of organic matter before it could be buried and fossilized. There is ample evidence for temperature increase during the early Eocene (Wing et al., 1991; Wolfe,

TABLE 3—Sites by species matrix of eleven localities shown in Figure 6. Cell values are the number of leaves. Underlined values are more than four standard deviations from the mean. Detailed explanation is in the text.

	Localities											
Taxa	8914	8912	CQ2	CQ3	8822f	8822l	8911	8917	8919	898	8910	$\bar{X}$
Aleurites glandulosa	0	0	0	0	0	3	0	0	0	0	0	0.3
Allantoidiopsis erosa	0	0	0	0	1	0	0	0	2	0	0	0.3
Alnus sp.	7	15	3	1	118	127	33	2	0	28	7	31.0
Asplenium eoligniticum	0	0	0	0	0	1	0	0	0	0	0	0.1
Cnemidaria magna	3	0	7	0	0	0	1	0	0	0	33	4.0
dicot III	5	30	14	4	22	5	96	64	15	12	135	36.5
dicot X	11	6	0	0	5	7	3	4	0	26	1	5.7
dicot XI	22	6	3	1	1	13	0	3	0	0	0	4.5
dicot XII	0	0	1	0	0	0	0	0	0	3	0	0.4
dicot XXV	7	4	0	1	0	11	4	0	0	0	0	2.5
dicot XXVII	0	0	0	0	15	7	0	0	0	0	0	2.0
dicot XXVIII	1	3	0	0	0	0	0	0	0	1	0	0.5
dicot XXX	0	1	0	0	0	0	0	0	0	0	0	0.1
dicot XXXI	10	0	0	0	0	0	0	14	18	0	0	3.8
dicot XXXII	84	0	0	0	0	0	0	0	0	0	0	7.6
dicot XXXIV	0	0	0	0	0	0	2	0	0	0	0	0.2
dicot XXXIX	0	0	0	0	0	1	0	0	0	0	0	0.1
dicot XXXV	0	0	0	0	0	0	0	0	0	11	0	1.0
dicot XXXVI	0	7	0	0	2	0	0	0	1	117	0	11.5
dicot XXXVII	0	0	0	0	0	4	0	0	0	0	0	0.4
dicot XXXVIII	0	0	0	0	0	1	0	0	0	0	0	0.1
Dombeya novi-mundi	19	53	24	20	67	66	46	42	51	89	24	45.5
Equisetum sp.	0	0	0	0	105	0	0	0	0	0	0	9.5
Glyptostrobus europaeus	222	115	115	67	0	108	166	45	40	163	100	103.7
Hymenophyllaceae sp.	1	0	0	0	0	0	0	0	0	0	0	0.1
Lygodium kaulfussi	11	5	1	1	2	0	23	11	9	8	4	6.8
Platycarya castaneopsis	29	112	24	7	41	41	3	51	0	5	102	37.7
Populus wyomingiana	0	0	0	0	0	8	0	0	0	0	0	0.7
cf. *Schoepfia republicensis*	0	1	0	0	0	0	0	0	0	0	0	0.1
"tatman fern"	7	43	0	0	0	0	3	0	0	0	0	4.8
Thelypteris iddingsii	0	0	0	1	0	4	0	8	1	0	0	1.3
sp898-3	0	0	0	0	0	0	0	0	0	1	0	0.1
sp8910-1	0	0	0	0	0	0	0	0	0	0	6	0.5
sp8910-2	0	0	0	0	0	0	0	0	0	0	1	0.1
sp8910-3	0	0	0	0	0	0	0	0	0	0	1	0.1
sp8911-1	0	0	0	0	0	0	3	0	0	0	0	0.3
sp8911-4	0	0	0	0	0	0	25	0	0	0	0	2.3
sp8912-2	15	0	0	0	0	0	0	0	0	0	0	1.4
sp8914-8	1	0	0	0	0	0	0	0	0	0	0	0.1
sp8914-9	1	0	0	0	0	0	0	0	0	0	0	0.1
sp8822-3	0	0	0	0	3	0	0	0	0	0	0	0.3
sp8822-8	0	0	0	0	0	1	0	0	0	0	0	0.1
sp8822-9	0	0	0	0	0	3	0	0	0	0	0	0.3
# leaves	456	401	192	103	382	411	408	244	137	464	414	328
# spp	18	14	9	9	12	18	13	10	8	12	11	12

1989; Bown and Kraus, 1981; Hickey, 1980), and this may also have played a role in higher rates of degradation of organic material. Bighorn Basin climate may also have become drier during the mid-early Eocene. Drying trends in the early Eocene have been observed in other basins of the northern Rocky Mountains. MacGinitie (1969) suggested increasing seasonal dryness during the early Eocene based on his study of plant fossils in the Green River Basin. The Niland tongue of the Wasatch Formation in the Green River Basin also records drying from lacustrine to swampy conditions in the lowest Lostcabinian (Roehler, 1993). By the mid-Lostcabinian in the same area, evaporites were forming as the Wilkins Peak Member of the Green River Formation (Smoot, 1983). In the Uinta and Piceance basins, the large freshwater Lake Uinta became increasingly saline from the earliest Eocene to the late early Eocene, which can also been attributed to drying (Franczyk et al., 1992). Within the Big Horn Basin, Bown and Kraus (1993) observed a decrease in soil hydromorphism, especially in the 350–600-m interval of the Willwood Formation where tabular carbonaceous beds are not preserved. This may, in part, be attributed to drying of climate, although it may also be affected by tectonics and relative subsidence.

Drying in the Bighorn Basin also may have been generated by local tectonics. The uplift of the Owl Creek Mountains to the south and southeast of the Basin during the early Eocene might have caused an orographic rain shadow in the Bighorn Basin (Bown and Kraus, 1981). This is assuming that a significant proportion of precipitation was derived from the southeast, which is not unreasonable in view of the size and location of the Mississippi Em-

bayment. Erosion of uplifted areas would have led to higher sediment-accumulation rates in the basin. When uplift ceased, continued erosion gradually would have reduced the elevation of the mountains. Once equilibrium was reached, erosion would have slowed and sediment-accumulation rates returned to a lower level. Uplifted areas would eventually have been eroded to such an extent that they no longer would have cast a rain shadow. Ultimately, the Tatman Formation, composed of lacustrine and anastomosed stream sediments, filled the central Bighorn Basin when downwarping and subsidence, associated with up-thrusting of the basement, resumed in the middle Eocene (Olander, 1987; Filkins, 1986).

With present information, it is not possible to unambiguously attribute changes in the abundance and type of carbonaceous beds to climatic as opposed to tectonic factors. Quite possibly, both forces were involved and may have interacted with one another to influence the distribution of carbonaceous units.

CONCLUSIONS

The complex and variable arrangement of sedimentary subunits within laterally extensive carbonaceous beds reflects the original mosaic of subenvironments found in distal backswamps of fluvial systems with avulsive phases. Fossil floras preserved in carbonaceous beds are essentially autochthonous and are believed to represent a maximum period of about 2,000 years, with individual assemblages from within a particular sedimentary sub-unit having formed on the order of years or tens of years and individual bedding planes less than this. Such fossil floras are not significantly time-averaged with respect to most evolutionary or climatic processes, although combining quarry samples from a single carbonaceous bed may disguise short-term ecological processes and patterns such as successional change.

Initiation of carbonaceous bed deposition is probably related to climatic or tectonic forces, but termination is controlled by fluvial events associated with channel avulsion. Carbonaceous beds develop towards the end of a fluvial cycle when distal areas are stable and low-lying, and their formation ceases when avulsion onto these areas occurs.

The large-scale stratigraphic distribution of carbonaceous beds in the Bighorn Basin was probably controlled by both tectonic and climatic change. Tabular carbonaceous beds are absent from the middle part of the Willwood Formation because floodplains had few stable, wet, distal areas required for the preservation of organics. Less stable, seasonally dry floodplains were probably the result of a period, lasting 0.5–1 m.y., of seasonally drier climates and higher rates of clastic sediment accumulation that were, in turn, caused by local orographic rain shadows, and possibly by regional or global climatic change.

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REFERENCES

BECK, R.A., VONDRA, C.F., FILKINS, J.E., and OLANDER, J.D., 1988, Syntectonic sedimentation and Laramide basement thrusting, Rocky Mountain foreland: timing and deformation: *in* SCHMIDT, C.J., and PERRY, W.J., eds., Interaction of the Rocky Mountain Foreland and Cordilleran thrust belt: Geological Society of America Memoir, v. 171, p. 465–487.

BEHRENSMEYER, A.K., HOOK, R.W., BADGLEY, C.E., BOY, J.A., CHAPMAN, R.E., DODSON, P., GASTALDO, R.A., GRAHAM, R.W., MARTIN, R.D., OLSEN, R.D., SPICER, R.A., TAGGART, R.E., and WILSON, M.V.H., 1992, Paleoenvironmental contexts and taphonomic modes: *in* BEHRENSMEYER, A.K., DAMUTH, J.D., DIMICHELE, W.A., POTTS, R., SUES, H.D., and WING, S.L., eds., Terrestrial Ecosystems through Time: Evolutionary Paleoecology of Terrestrial Plants and Animals: University of Chicago Press, Chicago, p. 15–136.

BOWN, T.M., 1980, The Willwood Formation (lower Eocene) of the southern Bighorn Basin, Wyoming, and its mammalian fauna: *in* GINGERICH, P.D., ed., Early Cenozoic paleontology and stratigraphy of the Bighorn Basin, Wyoming: University of Michigan Museum of Paleontology, v. 24, p. 127–138.

BOWN, T.M., and KRAUS, M.J., 1981, Lower Eocene alluvial paleosols (Willwood Formation, Northwest Wyoming, U.S.A.) and their significance for paleoecology, paleoclimatology and basin analysis: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 34, p. 1–30.

BOWN, T.M., and KRAUS, M.J., 1993, Time-stratigraphic reconstruction and integration of paleopedologic, sedimentologic, and biotic Events (Willwood Formation, lower Eocene, northwest Wyoming, U.S.A.: PALAIOS, v. 8, p. 68–80.

BOWN, T.M., ROSE, K.D., SIMONS, E.L., and WING, S.L., 1994, Distribution and stratigraphic correlation of upper Paleocene and lower Eocene fossil mammal and plant localities of the Fort Union, Willwood, and Tatman Formations, southern Bighorn Basin, Wyoming: U.S. Geological Survey Professional Paper, No. 1540, p. 1–269.

BULLOCK, P.N., FEDEROFF, A., STOOPS, G., and TURSINA, T., 1985, Handbook for soil thin section description: International Society for Soil Science, Waine Research Publications, 150 p.

BURNHAM, R.J., 1994, Patterns in tropical leaf litter and implications for angiosperm paleobotany: Review of Palaeobotany and Palynology, v. 81, p. 99–113.

CHAO, G.Y., 1969, 2Θ(Cu) table for common minerals: Geological Paper 69-2, Carleton University, Ottawa, Canada.

DAVIES-VOLLUM, K.S., 1996, Underbeds beneath Tertiary carbonaceous shales of the Bighorn Basin, Wyoming: Geological Society of America, Abstracts with Programs, v. 28 (7), p. A208.

DAVIES, K.S., and SHAW, G., 1993, Fixation of ^{137}Cs by soils and sediments in the Esk Estuary: Science of the Total Environment, v. 132, p. 71–92.

FARLEY, M.B., 1989, Palynological facies fossils in nonmarine environments in the Paleogene of the Big Horn Basin: PALAIOS, v. 4, p. 565–573.

FARLEY, M.B., 1990, Vegetation distribution across the early Eocene depositional landscape from palynological analysis: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 79, p. 11–27.

FILKINS, J.E., 1986, Geochemistry and depositional history of the coals of the Bighorn Basin, Wyoming: M.S. thesis, Earth Sciences Deptartment, Iowa State University, Ames, 132 p.

FRANCZYK, K.J., FOUCH, T.D., JOHNSON, R.C., MOLENAAR, C.M., and COBBAN, W.A., 1992, Cretaceous and Tertiary paleogeographic reconstructions for the Uinta-Piceance Basin study area, Colorado and Utah: U.S. Geological Survey Bulletin, No. 1787, 37 p.

GINGERICH, P.D., 1983, Paleocene-Eocene faunal zones and a preliminary analysis of Laramide structural deformation in the Clark's Fork Basin, Wyoming: Wyoming Geological Association Guidebook of the 34th Annual Field Conference, p. 185–195.

HICKEY, L.J., 1980, Paleocene stratigraphy and flora of the Clark's Fork Basin: *in* GINGERICH, P.D., ed., Early Cenozoic paleontology and stratigraphy of the Bighorn Basin, Wyoming: University of Michigan Museum of Paleontology, v. 24, p. 33–49.

JAKOBSEN, B.H., 1988, Accumulation of pyrite and Fe-rich carbonate and phosphate minerals in a lowland moor area: Journal of Soil Science, v. 39, p. 447–455.

JENIK, J., 1976, Roots and root systems in tropical trees: Morphologic and ecologic aspects: *in* TOMLINSON, P.B., and ZIMMERMAN, M.H., eds., Tropical trees as living ecosystems. Cambridge University Press, Cambridge, p. 323–349.

KOCH, P.L., ZACHOS, J.C., and DETTMAN, D.L., 1995, Stable isotope stratigraphy and paleoclimatology of the Paleogene Bighorn Basin (Wyoming, USA): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 115, p. 61–90.

KRAUS, M.J., and ASLAN, A., 1993, Eocene hydromorphic paleosols: Significance for interpreting ancient floodplain processes: Journal of Sedimentary Petrology, v. 63, p. 453–463.

KRAUS, M.J., and BOWN, T.M., 1993, Short term sediment accumulation rates determined from Eocene paleosols: Geology, v. 21, p. 743–746.

MACGINITIE, H.D., 1969, The Eocene Green River flora of northwestern Colorado and northeastern Utah: University of California Publications in Geological Sciences 83, 140 p.

MIALL, A.D., 1987, Recent developments in the study of fluvial facies models: *in* ETHRIDGE, F.G., FLORES, R.M., and HARVEY, M.D., eds., Recent Developments in Fluvial Sedimentology, Contributions from the Third International Fluvial Sedimentology Conference: Society of Economic Paleontologists and Mineralogists Special Publication 39, p. 1–9.

MOORE, D.M., and REYNOLDS, R.C., 1989, X-ray Diffraction and the Identification and Analysis of Clay Minerals: Oxford University Press, Oxford, 332 p.

OLANDER, J.D., 1987, Stratigraphy and sedimentology of the Eocene Tatman Formation, Bighorn Basin, Wyoming: M.S. thesis, Earth Sciences Dept., Iowa State University, Ames, 133 p.

PELZER, G., RIEGEL, W., and WILDE, V., 1992, Depositional controls on the Lower Cretaceous Wealden coals of Northwest Germany: *in* MCCABE, P.J., and PARRISH, J.T., eds., Controls on the distribution and quality of Cretaceous coals: Geological Society of America Special Paper 267, p. 227–244.

PIERCE, J.W., and SIEGEL, F.R., 1969, Quantification in clay mineral studies of sediments and sedimentary rocks: Journal of Sedimentary Petrology, v. 39, p. 187–193.

RETALLACK, G.J., 1984, Completeness of the rock and fossil record: Some estimates using fossil soils: Paleobiology, v. 10, p. 59–78.

RETALLACK, G.J., 1990, Soils of the past. An introduction to paleopedology: Unwin Hyman, Boston, 520 p.

ROEHLER, H.W., 1993, Eocene climates, depositional environments, and geography, Greater Green River Basin, Wyoming, Utah and Colorado: U.S. Geological Survey Professional Paper 1506-F, 74 p.

SCHANKLER, D. M., 1980, Faunal zonation of the Willwood Formation in the central Bighorn Basin, Wyoming: *in* GINGERICH, P.D., ed., Early Cenozoic paleontology and stratigraphy of the Bighorn Basin, Wyoming: University of Michigan Museum of Paleontology, v. 24, p. 99–114.

SMITH, D.G., 1986, Anastomosing river deposits, sedimentation rates and basin subsidence, Magdalena River, northwestern Colombia, South America: Sedimentary Geology, v. 46, p. 177–196.

SMITH, D.G., and SMITH, N.D., 1980, Sedimentation in anastomosed river systems: examples from alluvial valleys near Banff, Alberta: Journal of Sedimentary Petrology, v. 50, p. 157–164.

SMITH, N.D., CROSS, T.A., DUFFICY, J.P., and CLOUGH, S.R., 1989, Anatomy of an avulsion: Sedimentology, v. 36, p. 1–23.

SMOOT, J.P., 1983, Depositional subenvironments in an arid closed basin, the Wilkins Peak Member of the Green River Formation (Eocene), Wyoming, U.S.A.: Sedimentology, v. 30, p. 801–827.

TAUXE, L., GEE, L., GALLET, Y., PICK, T., and BOWN, T.M., 1994, Magnetostratigraphy of the Willwood Formation, Bighorn Basin, Wyoming: New constraints on the location of the Paleocene/Eocene boundary: Earth and Planetary Science Letters, v. 125, p. 159–172.

WILDE, S.A., COREY, R.B., IYER, J.G., and VOIGHT, G.K., 1979, Soil and plant analysis for tree culture: Oxford and IBH Publishing Co., New Delhi, 224 p.

WILF, P., in press, When are leaves good thermometers? A new case for leaf margin analysis: Paleobiology.

WING, S.L., 1984, Relation of paleovegetation to geometry and cyclicity of some fluvial carbonaceous deposits: Journal of Sedimentary Petrology, v. 54, p. 52–66.

WING, S.L., in press, Late Paleocene—early Eocene floral and climatic change in the Bighorn Basin, Wyoming: *in* BERGGREN, W., AUBRY, M.P., and LUCAS, S., eds., Late Paleocene-Early Eocene Biotic and Climatic Events: Columbia University Press, New York, p.

WING, S.L., and DIMICHELE, W.M., 1995, Conflict between local and global changes in plant diversity through geological time: PALAIOS, v. 10, p. 551–564.

WING, S.L., and GREENWOOD, D.R. 1993, Fossils and fossil climates: the case for equable continental interiors in the Eocene: Philosophical Transactions of the Royal Society of London, series B, v. 341, p. 243–252.

WING, S.L., ALROY, J., and HICKEY, L.J., 1995, Plant and mammal diversity in the Paleocene to early Eocene of the Bighorn Basin: Palaeoclimatology, Palaeogeography, Palaeoecology, v. 115, p. 117–156.

WING, S.L., BOWN, T.M., and OBRADOVICH, J.D., 1991, Early Eocene biotic and climatic change in interior western North America: Geology, v. 19, p. 1189–1192.

WOLFE, J.A., 1989, North American Eocene vegetation and its climatic implications: Transaction of the American Geophysical Union (EOS), v. 70, p. 375.

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