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2 Alluvial Response to the Paleocene-Eocene Thermal Maximum Climatic Event, Polecat Bench,
3 Wyoming (U.S.A.)

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ABSTRACT

The stratigraphic interval spanning the Paleocene—Eocene Thermal Maximum in the northern Bighorn Basin, Wyoming shows changes in the alluvial record that can be tied to a high resolution climate record. The complexity and stratigraphic spacing of paleosols change through the study section. Comparison to the climate record, reconstructed from paleosols, indicates the changes correspond to Paleocene—Eocene Thermal Maximum climatic fluctuations. In particular, the middle of the section contains thick, welded paleosols and thin avulsion deposits that link to times of well-drained floodplains and lower mean annual precipitation. Stratigraphic intervals below and above the main part of the Paleocene—Eocene Thermal Maximum interval correspond to times of less well drained floodplains and higher mean annual precipitation. These strata contain thinner paleosols and thick avulsion deposits.

Differences in paleosol complexity and spacing suggest that sediment flux to the depositional site varied in response to precipitation fluctuations associated with the Paleocene—Eocene Thermal Maximum. Welded paleosols and thin avulsion deposits indicate reduced floodplain accretion during deposition of the middle of the Paleocene—Eocene Thermal Maximum. Intervals with widely spaced paleosols indicate more rapid accretion. We hypothesize that drier episodes associated with warming caused reduced vegetation in source areas and promoted erosion and increased sediment yield. Because precipitation was reduced, much of that sediment was stored in upstream reaches of the fluvial system rather than moving to the depositional basin. Welded paleosols formed because of diminished sediment supply to the basin. With a return to wetter conditions during the recovery phase of the Paleocene—Eocene Thermal Maximum, upstream water flux increased, stored sediment moved to the basin, and vertically spaced, thinner paleosols developed. The results demonstrate how vertical sections of

alluvial paleosols can provide information on how climate fluctuated through time and how the fluvial system responded to climate change.

Keywords: paleosol, PETM, alluvial architecture, paleoclimate, avulsion

1. Introduction

The generally warm climate that characterized the Early Cenozoic was punctuated by a dramatic and short lived (~200 kyr) spike in global temperatures at the Paleocene/Eocene boundary, ~56 Ma ago (e.g., Kennett and Stott, 1991; Zachos et al., 1993; Westerhold et al., 2009; Cui et al. 2011). During this event, termed the Paleocene—Eocene Thermal Maximum (PETM), global temperatures rose between 5 and 9°C (e.g., Kennett and Stott, 1991; Zachos et al., 2001; McInerney and Wing, 2011). The PETM was coeval with a large (~3-5‰) negative carbon isotope excursion (CIE) that can be detected in both marine and continental strata (e.g., Koch et al., 1992; Zachos et al., 1993, 2005; McInerney and Wing, 2011). Astronomically estimated time spans for the PETM have ranged from 150 kyr (Norris and Röhl, 1999) to 210–220 kyr (Röhl et al., 2000), whereas Murphy et al. (2010) estimated a total duration of ~217 kyr by using extraterrestrial ³He flux rates to adjust for variability in deep marine sedimentation rates.

Changes in the paleohydrological cycle have been attributed to PETM warming with various studies interpreting drier conditions for some mid-latitude areas (e.g., Wing, et al., 2005; Bowen and Bowen, 2008; Handley, et al., 2012) and increased humidity at high latitudes (e.g., Pagani et al., 2006). Several authors have also attributed changes in fluvial systems and their stratigraphic architecture to changes in PETM hydrology. For example, Schmitz and Pujalte (2007) argued that a conglomeratic interval located just above the onset of the PETM in the

Spanish Pyrenees developed as a result of more seasonal precipitation related to PETM warming. Foreman et al. (2012) concluded that a shift to thick multistory sandstones in the Piceance Basin is linked to an increase in precipitation associated with the PETM. Similarly, the PETM interval in the northern Bighorn Basin is associated with an unusually thick and laterally extensive channel-belt sandstone, termed the boundary sandstone, because of its association with the Paleocene/Eocene boundary (e.g., Gingerich, 2001; Foreman, 2014). Prior to identification of the PETM, Kraus and Middleton (1987) attributed the boundary sandstone to a decrease in the rate of basin subsidence; however, Foreman (2014) has linked the sandstone to climate fluctuations during the PETM. These studies have focused on the influence of PETM climate change on the architecture of coarse-grained fluvial deposits; however, climate fluctuations also influence the architecture of non-channel deposits. Additionally, few studies have demonstrated the importance of climate change on changes in alluvial architecture. The sparseness of these studies, in part, lies in the difficulty of determining climate changes at a sufficient stratigraphic resolution.

The goal of this study is to examine the relationship between the alluvial architecture and climate for an interval that spans from pre-PETM time through the recovery phase of the PETM. The strata are situated at Polecat Bench in the Bighorn Basin, where excellent exposures of PETM strata have been the focus of stratigraphic, sedimentologic, and paleontologic studies (e.g., Gingerich, 2001; Magioncalda, et al., 2004; Kraus and Riggins, 2007; Smith, et al., 2008). We focus on ~85 m thick interval that is dominated by non-channel deposits, including vertically stacked paleosols, which provide a high resolution record of climate. Qualitative estimates of soil moisture and paleo-precipitation come from previous study of the paleosols and ichnofossils (Smith et al., 2008). Quantitative estimates of precipitation are made using the CALMAG

method (Nordt and Driese, 2010). The stratigraphic interval shows changes in grain size and in the character and stratigraphic spacing of the paleosols; comparison to the climate record indicates that the changes are linked to episodes of drying or seasonal drying during PETM time. Our results suggest that sediment flux varied in response to climate fluctuations. Climate influenced both the amount of water available for fluvial processes as well as the kind of vegetation and its spatial distribution.

2. Geological setting

The Paleogene Willwood Formation is a ~780-m-thick alluvial succession deposited in the Bighorn Basin, which is a foreland basin bounded on three sides by major uplifts that were active during latest Cretaceous through Paleogene time (Fig. 1). The Beartooth and Bighorn uplifts, which form the northwestern and eastern margins of the basin, respectively, show little evidence of major movement after the latest Paleocene (e.g., Blackstone, 1986; Omar, et al., 1994). Orogenic timing of the Owl Creek Mountains, which form the southern boundary of the Bighorn Basin, is less certain. Some authors indicate low relief mountains in the area of the Owl Creek Mountains in early Eocene time (Lillegraven and Ostresh, 1988); however, paleocurrent measurements from Paleocene strata indicate northward flow away from the present mountain belt (Seeland, 1998). The volcanogenic Absaroka Mountains postdate Willwood deposition (e.g., Smedes and Prostka, 1972). The Willwood Formation has been divided into three basic facies: i) thick (>10 m), laterally extensive sheet sandstones ascribed to deposition in meandering trunk channels; ii) heterolithic deposits consisting of interbedded sandstone and mudstone with weak pedogenic modification, attributed to channel avulsion; and iii) fine-grained deposits showing moderate to extensive pedogenesis and interpreted as overbank deposits (Kraus and Aslan, 1993; Kraus, 1996). Willwood channel-belt sandstones are usually 10—25 m thick and 1 to 1.5 km

wide (e.g., Kraus, 2002; Jones and Hajek, 2007). Statistical analysis of photomosaics shows that channel-belt sandstones make up less than 10% of the Willwood stratigraphy (A. Noda and P.L. Heller, pers. comm, 2014). Overbank and avulsion deposits dominate the Willwood Formation (e.g., Kraus and Aslan, 1993; Kraus, 1996).

The study was conducted on Polecat Bench in the northern Bighorn Basin, Wyoming (Fig. 1). The study section is an ~85 m thick composite section chosen because it excludes large sandstone bodies and, thus, allows analysis of floodplain stratigraphy and architecture (Fig. 2A). In this part of the Bighorn Basin, the PETM is situated ~430 m above the contact between the Willwood Formation and underlying Fort Union Formation (e.g., Koch et al., 1992; Bowen et al., 2001). Carbon isotopes from carbonate nodules and dispersed organic carbon show that the study section begins ~20 m below the PETM interval and ends ~20 m above the end of the carbon isotope excursion (CIE) that defines the PETM (e.g., Koch et al., 1992; Bowen, et al., 2001; Magioncalda, et al, 2004). Astronomically estimated time spans for the PETM range from 150 ka (Norris and Röhl, 1999) to 210–220 ka (Röhl et al., 2000), and extraterrestrial ^3He methods, indicate a total duration of ~217 ka (Murphy et al., 2010). The PETM section can be divided into three phases (Fig. 2). The onset phase is recognized by a geologically rapid decrease in $\delta^{13}\text{C}$ values (e.g., Bowen et al., 2006). The duration of the onset was 8—23 kyr based on continental strata (summarized in McInerney and Wing, 2011) or ~ 20 kyr based on a marine record (Cui et al., 2011). At Polecat Bench, the base of the PETM is just below a major purple paleosol (Purple 0), and the onset of the PETM extends from that level to ~26 m level in the section (Fig. 2).

The second phase – the body of the PETM – began when $\delta^{13}\text{C}$ values reached a minimum (e.g., Bowen, et al., 2006) (Fig. 2). Maximum warming occurred during the body of the PETM, and $\delta^{13}\text{C}$ values were relatively low but stable (e.g., Thomas et al., 2002; Bowen, et al., 2006).

The body of the PETM lasted ~115 kyr (e.g., Abdul Aziz et al., 2008; Murphy et al., 2010). Temperatures and isotope values returned to pre-PETM conditions during the recovery phase, which may have taken ~50 kyr to ~85 kyr (e.g., Bowen, et al., 2006; Röhl et al., 2007; Murphy et al., 2010). At Polecat Bench, the body of the PETM extends to the Purple 4 paleosol at ~62 m in the study section, and the recovery phase extends up to ~77m level (Fig. 2).

Two additional sections show the alluvial architecture lateral to the study section. A core, obtained as part of the Bighorn Basin Coring Project (BBCP; Clyde et al., 2013), includes channel sandstones as well as paleosols and avulsion deposits (Fig. 2B). A stratigraphic section from northwest of Polecat Bench includes the boundary sandstone, where it is thickest (Fig. 2C) (Foreman, 2014). Stratigraphic correlations in Figure 2 are based on isotope stratigraphy from carbonate nodules from sections 2B (Bowen et al., 2015) and 2C (Foreman, 2014).

Paleoclimate proxy evidence indicates that temperatures rose ~5°C during the PETM in the Bighorn Basin; mean annual temperature (MAT) may have reached 26°C (Fricke and Wing, 2004; Wing et al., 2005). Regional models developed by Sewall and Sloan (2006) indicate that early Paleogene Laramide basins, including the Bighorn Basin, received summer monsoonal precipitation related to the Mississippi embayment. Analysis of fossil plants suggests that mean annual precipitation (MAP) dropped at the onset of the PETM and gradually increased to pre-PETM values by the end of the warming episode (Wing et al., 2005). Similarly, studies of paleosols (Kraus and Riggins, 2007; Kraus et al., 2013) and paleosol ichnofossils (Smith et al., 2008) suggest drier conditions in the Bighorn Basin at the beginning of the PETM interval and a return to more humid conditions during the recovery phase.

3. Methods

Individual sections were measured by digging ~1 m wide trenches down to fresh rock. Section locations were chosen to minimize lateral offset. Sections were subdivided into field units based on differences in grain size, sedimentary structures, matrix and mottle color, pedogenic nodules, and other pedogenic features including slickensides and ped structures. Color was described from fresh hand samples using Munsell color charts. The composite section can be divided into five different facies based on field estimates of grain size, ichnology, sedimentary structures and pedogenic features. Quantitative grain sizes for 341 samples distributed through the study section were determined using a Malvern Longbed Mastersizer. Grain size is reported according to the terminology of Folk (1954). Samples were analyzed for major element oxides using a Kevex 0700 X-ray fluorescence spectrometer at the University of Colorado Laboratory for Environmental and Geological Studies (LEGS). Weight percents given by XRF were recalculated to molar percents for estimating MAP (Nordt and Driese, 2010).

4. Facies

The five facies recognized in the study section are described and interpreted in the sections that follow. In addition, channel-belt sandstones, as seen in the other sections in Figure 2, are described.

4.1. *Variegated mudrocks I*

This facies consists dominantly of mudstone and sandy mudstone (Fig. 3A; classification of Folk, 1954), and beds show pedogenic modification such as color banding and mottling; ichnofossils, redoximorphic concentrations and depletion features; and slickensides and mukkar structures (e.g., Retallack, 1988; Kraus and Hasiotis, 2006; Smith, et al., 2008). Each paleosol is

typically thicker than 50 cm, and grain size is relatively uniform downwards until a basal silty bed is reached (Fig.4A). A typical paleosol consists of an upper gray (5GY 5/2) unit that is underlain by a series of red, red-brown, purple or, less commonly, yellow-brown beds. Mottles in the upper gray unit are the same color as the matrix of the underlying unit, and gray rhizoliths (root traces) commonly extend downward from the topmost unit into the underlying unit of the paleosol (Fig. 4A). Trace fossils include rhizohaloes and rhizcretions, both of which are specific types of rhizoliths (Fig. 5A). Adhesive meniscate burrows (AMB) are locally associated with the root traces. Slickensides, which are commonly intersecting and up to several tens of cm in breadth, are commonly well developed (Fig. 5A). The red paleosol shown in Figure 4A has a red, high chroma matrix (5R 3/4 to 4/6), and small (<3 mm diameter) ferruginous nodules are present. Calcium carbonate nodules are present, but sparse (5-10%), in the lower B horizon. The diagnostic characteristics of the paleosols are: mottles and trace fossils that show gradual, systematic downward changes in color and abundance and grain sizes that show little vertical variability (Fig. 4A). Mottles become grayer (lower chroma colors) downwards. Neither the colors and abundances of mottles nor the abundance and types of ichnofossils show major shifts at recognizable unit boundaries.

The mudrocks are interpreted as overbank deposits with steady and relatively slow rates of sediment accumulation that allowed moderate to strong soil development (e.g., Kraus and Aslan, 1993; Kraus, 1997). The paleosols are interpreted as cumulative paleosol profiles (e.g., Marriott and Wright, 1993), which develop when sedimentation is steady and slow, such that pedogenesis accompanies profile aggradation. Johnson and Watson-Stegner (1987) referred to this incorporation of new material as developmental upbuilding. If soil development is cumulative, roots and mottles are distributed uniformly through the profile, and grain size tends

to be relatively uniform (e.g., Aslan and Autin, 1998). The gradational nature of features in the Willwood paleosols suggests that soil profiles developed on floodplain surfaces that received regular influxes of sediment that were thin enough to be assimilated into the upward building soil (e.g., Marriott and Wright, 1993; Daniels, 2003).

The high chroma matrix of red paleosol in Figure 4A indicates well drained to moderately drained conditions (e.g., Richardson and Daniels, 1993; Cornell and Schwertmann, 2003). The ferruginous and calcium carbonate nodules as well as slickensides, yellow-brown mottles, and calcareous cores in rhizoliths indicate seasonal fluctuations in soil moisture (e.g., Wilding and Tessier, 1988; Vepraskas, 1994; Pipujol and Buurman, 1997). The B horizon is designated a Bss horizon because of slickensides (ss); the Bk designation reflects the presence of carbonate nodules (k) in that bed.

4.2. Variegated mudrocks II

This facies also consists largely of mudstone and sandy mudstone beds (Fig. 3B) that show evidence for pedogenic modification including ubiquitous mottles, ichnofossils, and slickensides (Fig. 4B). Each paleosol can be subdivided into multiple beds, based on abrupt vertical changes in mottle and/or matrix color, nodule or ichnofossil abundance, and/or grain size. In addition, a number of profiles show evidence for erosion between the different parts of the profile. A representative profile with multiple horizons is shown in Figure 4B. Gray units form the base and top of the profile. Between the gray beds, matrix colors and grain size allow six B horizons to be identified and to be grouped into three couplets. The upper horizons of the couplets (Bssg1, Bssg3, Bssg horizons) have grayish purple (5P 3/2 to 4/2) matrix, whereas the lower horizons (Bssg2, Bss, Bw) are redder (5RP 4/2 to 5/2); grain size tends to be finer in the redder horizons, except for the lower part of the paleosol. Green (10G 6/2) rhizoliths up to 8 cm

long and ferruginous nodules are more common in the upper horizons (Bssg1, Bssg3, Bssg). Ferruginous nodules are either powdery or uncommon in the Bssg2, Bss, and Bw horizons. Contacts between successive B horizons are sharp except in the lower part of the paleosol where Bssg and Bw have gradational lower contacts.

The abrupt changes in matrix color, nodule content, and rhizoliths at unit boundaries as well as the sharp nature of the contacts are attributed to an irregular depositional history accompanied by multiple, distinct stages of soil development. The paleosols in this facies are interpreted as compound to composite paleosol profiles that developed on unstable surfaces (e.g., Marriott and Wright, 1993). As defined by Morrison (1978), composite soils show overlap between two profiles, whereas compound soils show no overlap. Thus, the balance between the rates of deposition and pedogenesis are somewhat different. Features of the Willwood paleosols indicate that an initial stage of deposition and pedogenesis was interrupted by erosion of the top of the developing profile or by an addition of sediment that was too thick to be simply incorporated into the developing profile. If erosion occurred, renewed deposition and pedogenesis produced an upper profile that did not completely overprint the older profile below but was welded to it (Ruhe and Olsen, 1980). If sufficient sediment was added to the top of the developing profile, probably as avulsion or crevasse deposits, that sediment and continued pedogenesis either produced an upper profile that was welded or partially welded to the profile below. Johnson and Watson-Stegner (1987) referred to the later as retardant upbuilding.

Most of the purple horizons of Fig. 4B are interpreted as a Bssg horizons based on the purple, low chroma (5RP 4/2) matrix color (g) and well-developed slickensides (ss). The matrix color and various redoximorphic features – yellow-brown iron accumulations in depletion zones, gray rhizoliths – indicate somewhat poorly drained conditions but with seasonal drying of the

solum (e.g., Vepraskas, et al., 1992; Wright et al., 2000; Jacobs, et al., 2002). Changes in features such as mottle colors are related to variation in the redox conditions of the older soil profiles: closer to the surface and more oxidizing in its original position and further below the surface in the new profile (e.g., Vepraskas, 1994; Violante, 2003). Similarly, rhizcretions, rhizohaloes, and local burrows, which become more abundant upwards through the lower part of the paleosol, are significantly less abundant in the upper part of the welded profile. Some profiles included in this facies have less abrupt variations, but all profiles can be subdivided into at least two distinct parts based on changes in mottle colors and/or ichnofossil abundance.

Unusually thick paleosols with numerous subdivisions characterize the middle of the PETM interval (Figs. 2A, 5B, 6A). Figures 5B and 6A show the Upper Double Red paleosols: two red paleosols separated by a thin gray interval assigned to the heterolithic facies. Each paleosol consists of six or more beds with contacts between beds placed at changes in grain size or soil features, as described above. Downward grain size changes are not systematic through the paleosols, but vary irregularly. The Lower Double Red paleosols (Fig. 5B) and Purple-2 paleosol are also unusually thick, and grain size variations and morphologic features allow them to be subdivided into multiple horizons. These paleosols reflect welding due to both erosion and retardant upbuilding.

4.3. *Heterolithic facies*

This facies consists primarily of sandy siltstone and siltstone beds (Fig. 3C) that are typically thinner than 30 cm and contain local relict bedding (Figs. 4C, 5C). The average sand percentage is 20%, but sand can reach 50%. Matrix colors tend to be gray to yellow gray (5GY 7/2). Some beds have sparse (5-10%) to locally common (10-20%) mottles that can be yellow-brown, brown, red, or purple (Figs. 4C, 5C, 5D). In other beds of this facies, mottles are rare.

Carbonate nodules are locally common. Rhizoliths, usually as rhizohaloes or carbonate rhizocretions, are rare (<5%) to common (Fig. 5D). Burrows are generally less common, but better preserved, than in the two variegated mudrock facies, and include passively-filled burrows (cf. *Cylindricum* isp).

The heterolithic facies shows evidence for limited pedogenesis. The presence of only local ichnofossils and mottles and the preservation of relict bedding indicate limited bio- or pedoturbation, and, thus, relatively rapid deposition and burial of the sediment (e.g., Daniels, 2003; Hembree and Hasiotis, 2008). Beds that contain rhizoliths, well-preserved burrows, and mottles indicate slightly stronger, but still limited, pedogenic development. Because of limited pedogenesis, indicating relative rapid sediment accumulation, this facies is interpreted as crevasse-splay/avulsion deposits. The variety of mottle colors indicates that drainage conditions likely ranged from poorly to moderately well-drained (e.g., Kraus and Aslan, 1993; Vepraskas, 1994).

4.4. Laminated mudrocks

The laminated facies consists of beds of mudstone to sandy mudstone in colors that range from gray to dark gray to brown (Fig. 3D; Fig. 5E). Mudstone laminations or interlaminae of sandstone, siltstone, and claystone are the diagnostic character of the facies. Organic matter is common (10-20%), usually as fragments <1 mm in length. Gypsum crystals are locally common, whereas ichnofossils are typically sparse (5-10%) to absent. Where ichnofossils are present, passively-filled burrows dominate and include burrows with *in situ* mollusks (Fig. 5E). Yellow brown rhizocretions are locally present. This facies is typically underlain and overlain by the heterolithic facies.

The preservation of finely detailed sedimentary structures and burrows indicates a lack of significant flow energies and low amounts of biologic reworking (Hasiotis et al., 2001). The unionid bivalve fossils, particularly in life position, suggest standing water (Good, 1993). The gypsum crystals are probably relatively recent features developed in response to oxidation of a sulphur-rich source such as organic matter or minerals such as pyrite, which indicate initial waterlogged conditions (e.g., Dent, 1980; Aslan and Autin, 1998). This combination of features suggests that the laminated facies represents slack water deposition in ponds or other quiet waters.

4.5. *Channel-form facies*

The channel-form facies consists largely of sandy siltstone beds that fill a channel-form shape. Sand percentages range from ~10 to >60% (Fig. 3E). The channel-form bodies are up to several m thick and usually less than 15 m wide. Each channel-form body has a sharp lower contact that typically scours the underlying unit (Fig. 5F). Relict bedding and small- to medium-scale crossbedding are common. Channel-form sandstones are not common in the section, but, where present, are associated with heterolithic deposits.

This facies is interpreted as the fill of small crevasse-splay channels (e.g., Smith, et al., 1989; Perez-Arlucea and Smith, 1999). The range in grain sizes and sedimentary structures of beds filling the scours suggest variable flow energy as the channel filled.

4.6. *Channel-belt sandstones*

Although the study section does not contain channel-belt sandstones, both the core section (Fig. 2B) and section of Foreman (2014) (Fig. 2C) contain multistory sandstones deposited by channel belts. The channel-belt sandstones in the core are similar to those

described by Jones and Hajek (2007) for the Willwood Formation. Foreman (2014) also points out that, despite its unusual thickness (maximum thickness of 33m), the boundary sandstone resembles others in the Willwood Formation with no evidence for deeper scouring or greater erosion. Paleocurrent measurements show that the fluvial system in the northern Bighorn Basin flowed northward (Kraus and Middleton, 1987; Foreman, 2014).

4.7. Overall depositional environment

The boundary sandstone and channel-belt sandstones recognized in the core represent deposits of trunk rivers, and the northerly flow indicates they were part of an axial river system. The strong pedogenic development of the variegated mudrocks indicates slow rates of sediment accumulation, consistent with deposition as overbank deposits (e.g., Kraus and Aslan, 1993; Kraus and Gwinn, 1997). The gray mudrocks, heterolithic facies, and channel-form facies are interpreted as splay and avulsion deposits. These facies show a range of soil development with minimal pedogenesis for the gray mudrocks and somewhat stronger pedogenesis for the mottled mudrocks of the heterolithic deposits. Although those differences suggest variable rates of deposition, overall the mudrocks in both facies reflect rapid deposition compared to the variegated mudrocks. The association between the heterolithic facies and channel-form deposits is characteristic of Willwood avulsion deposits (e.g., Kraus, 1996; Kraus and Gwinn, 1997). The pond deposits, although sparse, are also associated with the avulsion deposits and probably represent local depressions on the avulsion belt surface. The vertical transition from overbank to avulsion/splay deposits is usually gradual, which is consistent with progradational avulsion in which a sediment wedge originates from the channel and develops in a downstream direction (Slingerland and Smith, 2004). The thicker and more laterally extensive avulsion deposits are probably the result of regional rather than local avulsion (Abels, et al., 2013).

5. Soil drainage and grain size

Willwood strata contain a diverse and abundant group of vertebrate, invertebrate, and plant trace fossils. The assemblage present in a particular depositional setting or particular paleosol reflects both moisture levels and temperature. Consequently, Smith et al. (2008) used particular assemblages of ichnofossils in conjunction with associated lithologic features, including paleosol features, to assess paleodrainage for paleosols and non-paleosol units through the Polecat Bench section. The combination of lithologic and ichnologic features indicates that environmental settings varied from poorly drained to well drained conditions at different stratigraphic levels (Fig. 7D). In general, conditions ranged from poor to intermediate drainage with pulses of moderate to good drainage between Lower Double Red paleosols and Purple-4.

Drier and wetter paleosols exist, even when the climate is drier or wetter. Grain size and landscape position both influence the drainage of alluvial soils. In general, more poorly drained soils form in floodbasins where clayey sediments accumulate in topographic depressions, and better drained soils form on coarser-grained deposits in topographically raised areas (e.g., Kraus and Aslan, 1999). Analysis of grain size reveals that the study section can be subdivided into 4 intervals (Table 1, Fig. 7C, Appendix 1). Statistical analysis confirms differences among the four subdivisions in terms of % sand and % clay (Supplementary Information). Subdivision 1 is much finer grained (has more clay) than subdivision 2. Sand content increases and clay content decreases between subdivisions 1 and 2. Between intervals 2 and 3, sand content decreases and clay content increases. Similar conclusions can be obtained by examining Figure 8 with boxplots of sand, silt, and clay percentages in the four stratigraphic subdivisions (see the Supplementary Information - Statistics for details).

Comparison between the paleodrainage curve and grain size subdivisions of the stratigraphic section show that paleosol drainage patterns appear to be independent of grain size trends. Between grain size intervals 1 and 2, as grain size increases, paleosols became more poorly drained. As grain size decreased between intervals 2 and 3, paleosols became drier, and, as grain size became coarser in interval 4, paleosols again became more poorly drained. The lack of correlation between grain size and paleodrainage suggests that the upward variations in paleosol drainage reflect changes in paleoclimate.

6. Paleoclimatic interpretations

A high resolution record of temperature change through the study section is not available. However, an assessment of temperature derived from $\delta^{18}\text{O}$ values in fossil mammal teeth collected in the southeastern Bighorn Basin shows that temperature rose steadily from the pre-PETM into the PETM interval and then declined during the recovery phase (Secord, et al., 2012). Intersecting slickensides in both Variegated Facies I and II and muklara structures indicate that paleosols in the study section are ancient analogs to Vertisols (e.g., Nettleton, et al., 2000; Driese and Ober, 2005). Consequently, changes in precipitation through the section were estimated using the quantitative approach developed by Nordt and Driese (2010) for Vertisols

Following their approach, we sampled subdivisions of the B horizon to a depth of ~1 m, if possible, and then used a weighted average of the base oxides to calculate MAP for a particular soil. The standard error for this method is ± 108 mm. Estimates were calculated for 13 different paleosols (Table 2; Appendix 2). More than one estimate was calculated for 9 of the paleosols by collecting samples from lateral equivalents of the same paleosol in outcrop and in the BBCP core. MAP values for an individual paleosol are relatively consistent with the exception of the upper double red paleosols (Fig. 7E). Additional MAP estimates were calculated using data

from Kraus and Riggins (2007). Although additional paleosols are present in the section, they were too thin and weakly developed to be used for MAP estimates (e.g., Adams, et al., 2011).

MAP values are plotted against stratigraphic level (Fig. 7E). Estimates are generally around 1100 mm; lower values and greater variability in MAP estimates occurs between ~30 and 50 m in the study section, thus, in the middle of the PETM interval. Figure 7 shows that trends between the MAP estimates and soil drainage proxy are relatively consistent. Conditions were generally poorly to imperfectly drained from the base of the section up to ~28 m level, which extends from the pre-PETM to a few meters into the body of the PETM interval (Fig. 7D). Relatively wet conditions are consistent with MAP values over ~1100 mm (dashed purple line in Fig. 7E). Between ~30m and ~50 m in the section, corresponding to the main body of the PETM, drainage fluctuated, but was much drier than below with moderately to well drained environments (Fig. 7D). In particular, a peak of better soil drainage and a lower MAP value corresponds to the Lower Double Red A paleosol. Although conditions are wetter at the overlying Lower Double Red B paleosol, Purple-2 paleosol also shows drier conditions and a lower MAP (Fig. 7E). More complicated are the Upper Double Red paleosols because the ichnofossils indicate moderately- to well-drained conditions, whereas MAP values for the upper paleosol show a significant range with both dry (~800 mm rainfall) and moderately wet (>1100 mm) values, and the estimate for the lower paleosol is moderately wet (~1100 mm). Above the 50 m level and into the PETM recovery, climatic indicators suggest a return to wetter conditions with some drying at the top of the section.

The results are consistent with paleoclimatic reconstruction for the PETM in the southeastern Bighorn Basin. Using paleosol proxies, Kraus et al. (2013) concluded that the PETM interval showed overall drier climates, but with fluctuations. The paleofloral record from

that part of the basin also shows a dramatic change at the onset of the PETM that indicates a decrease in soil wetness during the PETM (Wing, et al., 2005; Peppe et al., 2011). Latest Paleocene floras in the southeastern basin have a large proportion of taxa with relatives that today live in mesic temperate climates; those taxa are absent from the overlying PETM interval. Instead, the PETM floras are dominated by several kinds of Fabaceae (the legume family), and those that have been identified suggest an increase in temperature and water stress during the PETM (Wing, et al., 2005; Kraus, et al., 2013). A paleofloral record is not known from PETM strata in the northern Bighorn Basin.

7. Stratigraphic variations in facies

7.1. Statistical analyses of facies associations

Figure 7B shows the vertical distribution of the five facies in the measured composite section. The section can be subdivided into 5 stratigraphic intervals based on facies abundance (Tables 3 and 4, Appendix 3). The association between facies types and the 5 vertical subdivisions was analyzed in three ways (Supplementary Information). The first approach tests for a difference in the frequency of different facies across different stratigraphic intervals. The other two methods test for a difference in the total “amount” of different facies across different intervals, where “amount” is defined as (1) total thickness of each facies type or (2) percentage of each facies within the different subdivisions (Tables 3, 4). All three approaches show statistically significant differences in the kinds of facies present and their abundances across the 5 vertical subdivisions (see Supplementary Information for details).

7.2. *Variable rates of sediment accumulation*

Avulsion deposits and the two kinds of paleosol indicate different rates of sediment accumulation. Limited pedogenic development indicates that the avulsion deposits accumulated very quickly compared to the paleosols; for example, Daniels (2003) observed that, if floodplains aggrade faster than 0.5 cm/yr, recognizable soils do not develop. Similarly, study of modern avulsion deposits indicates that roughly one meter of sediment can accumulate in only several hundred years (e.g., Smith et al., 1989; Farrell, 2001). Other studies show that the time required for complete avulsion of a channel can be several millennia (e.g., Jones and Schumm, 1999), suggesting a somewhat slower, but still quite rapid, rate of floodplain deposition. Thus, parts of the study interval containing significant thicknesses of avulsion deposits, especially those that show minimal soil development, represent relatively rapid deposition.

Subdivisions of the study section that contain significant thicknesses of paleosols represent slower rates of floodplain accretion, and the type of paleosol provides additional information on deposition rates. As pointed out by Wright and Marriott (1996), floodplain deposition can be the result of frequent, but thin, influxes of sediment; gradual up-building of the floodplain accompanied by pedogenesis produces a cumulative paleosol (e.g., Morrison 1978, Marriott and Wright, 1993). Development of Willwood cumulative paleosols was eventually halted by an avulsion event; here, the frequency of depositional events is low but sufficient thickness is deposited quickly enough to bury and preserve the underlying cumulative soil. If the rate of floodplain accumulation is slow relative to the rate of pedogenesis, vertically successive soil profiles may partly overlap, producing composite paleosols. In particular, welded soils can form if the sediment between two vertically successive soils, deposited in the Willwood strata by avulsion, is so thin that the overlying soil merges with that below (Ruhe and Olsen, 1980). And

if those paleosols are composite in nature, they indicate slower rates of accumulation than do cumulative paleosols.

7.3. Integration of facies, grain size, and paleoclimate

The lowest 8 m of the section are dominated by both cumulative and composite/compound paleosols, and silt and clay are the dominant grain sizes (Fig. 7). The proportion of the interval represented by paleosols suggests that accretion rates were moderately slow. Climate proxies indicate relatively humid conditions. The presence of pond/slack water deposits (laminated mudrock) supports poorly-drained conditions (e.g., Smith et al., 1989; Morozova and Smith, 2003).

Relatively humid conditions are also interpreted for the overlying ~5 m interval (Fig. 7). This interval is dominated by avulsion deposits, including channel-form sandstones, and these facies imply relatively rapid rates of sediment accumulation. Consistent with the facies, grain size is sandier and siltier than for underlying strata (Table 1).

Although grain sizes continue to be sandy and silty and avulsion deposits continue to dominate, the interval between 13 and 31 m also includes both cumulative and composite paleosols (Table 1, Fig. 7B, C). Avulsion deposits are relatively thick (1 m or more), especially in the lower part of the interval. Despite the total thickness of avulsion deposits, average accretion rates for the interval are interpreted to be slower than for the underlying 5 m interval because of the presence of paleosols. Drainage and MAP estimates continue to indicate humid conditions.

A marked change to less sand, more clay, and dominance by composite paleosols occurs at ~31 m and extends upwards for 20 m. This interval, situated in the middle of the main body of the PETM, is characterized by the prominent red bands of Lower Double Red and Upper Double

Red paleosols and the densely-spaced paleosols associated with Purple-2 (Figs. 6, 7A). The total thickness of welded paleosols and decrease in thickness of avulsion deposits indicate that average floodplain accretion was relatively slow. Climate proxies show fluctuations between drier and wetter conditions. In general, the wetter periods are characterized by avulsion deposits and thin, non-welded paleosols. Times of better drainage and lower MAP estimates correspond to Lower Double Red and Purple-2 paleosols (Fig. 7D, 7E). Upper Double Red is also associated with better drainage; however MAP results are more equivocal. The interval dominated by welded paleosols is stratigraphically equivalent to the Boundary Sandstone in areas northwest of Polecat Bench (Fig. 2C).

Grain size coarsens (more sand and less clay) above 51 m, and, above the ~58m level, avulsion deposits replace welded paleosols as the dominant facies. Paleosols are still common in the uppermost part of the section; however, they are likely to be well separated by avulsion deposits. The relative thicknesses of paleosols and avulsion deposits indicate a return to more rapid floodplain aggradation; drainage and MAP indicate a return to wetter environments.

8. Controls on alluvial architecture

8.1. Depositional patterns

Depending on location in the northern Bighorn Basin, the middle 20 m of the PETM is characterized by unusual deposits relative to the stratigraphic section above and below: thick, welded paleosols, thin avulsion deposits, and the abnormally thick channel-belt sandstone (Fig. 2). The welded paleosols and thin avulsion deposits indicate relatively slow rates of floodplain accretion during deposition of the middle of the PETM interval. The stratigraphic pattern above and below this interval is marked by paleosols that are thinner than the welded paleosols and more typically cumulative in nature. The paleosols are widely separated by avulsion deposits,

some of which are quite thick (Figs. 2, 7). Comparison to the paleoclimatic proxies indicates that the changes in alluvial architecture during the PETM are associated with fluctuating, but overall drier, climatic conditions. In particular, several of the thick welded paleosols correspond to better drained floodplains and lower MAP estimates.

The combination of reduced floodplain accretion and climate change during the PETM indicates that sediment supply from source areas, which were the surrounding Laramide mountain ranges, varied through time in response to climate change. Other factors that influence sediment supply, tectonic activity and changes in source lithology, are highly unlikely given the short time span ($\sim 10^4$ years) over which the sediment changed character. Whereas climate can impact an entire drainage basin quickly, tectonic changes take time to propagate through the system (e.g., Tucker and Slingerland, 1997; Castelltort and Van Den Driessche, 2003).

8.2. *Boundary sandstone*

Foreman (2014) concluded that genesis of the boundary sandstone associated with the PETM interval was climatically controlled. He eliminated a climatically-driven change in river size because reconstructed channel depths and widths are no different for this channel-belt sandstone than for others in the Willwood Formation. Two scenarios were considered possible. First, if sediment source areas underwent climate changes similar to those recorded in the Polecat Bench section, fluctuations in river discharge and gradient could have caused upstream reaches to alternately store and release sediment. Several authors (e.g., Hovius, 1998; Simpson and Castelltort, 2012) have suggested that periods of climate change, especially from wet to dry, have the potential for creating large sediment fluxes. Fluvial systems store sediment in upstream reaches if water flux has decreased in response to drier climates (e.g., Hooke, 2000). The stored sediment then moves to the depositional basin when water flux increases with a return to wetter

conditions. Thus, during a return to wetter conditions following dry climates, sediment was released to the Polecat Bench depositional site to generate the boundary sandstone.

Alternatively, climate in source areas was not synchronized with climate and climate changes in the Bighorn Basin. In this interpretation, Foreman (2014) suggested that the boundary sandstone was generated during episodes of PETM drying, which caused reductions in soil moisture and vegetation, and, in turn, a weakening of channel stability. Channel banks become more prone to erosion, in part, because root density is less (e.g., Gran and Paola, 2001; Katuwal, et al., 2013), and vegetation tends to reduce the shear stress of overbank flows such that, with lesser vegetation, floodplains are more prone to erosion (Griffin and Smith, 2004). Less stable channels and reduced floodplain vegetation led to reworking of floodplain deposits and preferential deposition of sand at the expense of fine-grained deposits, which were carried further downstream and out of the basin. As shown by Gouw and Berendsen (2007), bank erodibility controls the width and width/thickness ratio of channel-belt sands through its influence on the ability of a channel to migrate laterally.

8.3. Climate and sediment supply

The study section, with its record of successive paleosols, can help untangle the controls on alluvial architecture at Polecat Bench during the PETM. The welded paleosols and thin avulsion deposits reflect periods during which sediment supply to the floodplain decreased, and the intervals with paleosols separated by thick avulsion deposits indicate periods when sediment supply increased. In addition, the middle of the PETM interval (~32 – 50 m interval) shows somewhat less sand than the intervals above and below. Sediment production in catchment areas is controlled by the degree of vegetative cover, which, in turn, is controlled by climate. Diminished plant cover promotes erosion and sediment yield by decreasing water infiltration and

exposing more soil to overland flows (e.g., Bull, 1979; Bautista, et al., 2007; Collins and Bras, 2008; Zhongming et al., 2010). Erosion, and thus sediment production, can also increase if rainfall intensity increases (e.g., Nearing, et al., 2005). Modeling by Kirkby and Cox (1995) suggests that, in warm, seasonal climatic settings, soil erosion in catchment areas increases as MAP decreases.

Climate, including its precipitation characteristics, can modify sediment supply through its effect on the rate of sediment production in source areas and the rate at which that clastic debris is eroded and transported to depositional basins (e.g., Hovius, 1998; Leeder et al., 1998; Hooke, 2000; Macklin, et al., 2012). Changes in precipitation patterns can modify discharge, weathering in source areas, and the type and extent of vegetation cover in source areas and the depositional basin (e.g., Bull, 1991; Leeder, et al., 1998; Macklin, et al., 2012). If, for example, vegetation cover decreases in sediment source areas because conditions became drier, runoff and sediment yields increase, and more sediment can enter the upstream reaches of fluvial systems (e.g., Bosch and Hewlett, 1982; Fuller et al., 1998). In particular, reduction in forest canopy influences evapotranspiration and leads to increases in peak discharges (e.g., Jones, 2000). Various studies suggest that sediment supply is controlled not only by total precipitation, but also by how strongly seasonal that precipitation is (e.g., Tucker and Slingerland, 1997; Hooke, 2000; Cecil et al., 2003).

Fluvial deposits studied by Nador et al. (2003), which record fluctuations in sediment supply in response to climatic fluctuations between glacial (cold, dry) and interglacial (warm, wet) times, provide an analog for the Willwood PETM section. During glacial episodes, sediment, especially coarser sediment, was stored in source areas because vegetation decreased and because the transport ability of upstream reaches were diminished by dry conditions.

Consequently, basin locations received finer sediment, sediment supply was reduced, alluvial ridges underwent little building, and few crevasse sands were deposited on floodplains. The welded paleosols, relative sparse avulsion/splay deposits, and diminished sand in the middle of the PETM interval resemble the deposits that formed in basinal positions during glacial episodes.

During interglacial times when climate was wetter, vegetation cover increased and capacity of upstream reaches was greater (Nador et al., 2003). Stored sediment was flushed to downstream areas, where rivers developed more prominent alluvial ridges, and avulsion/splay deposition on the floodplain was more common. The sections below and above the middle of the PETM interval show common avulsion deposits, coarser (sandier) grain sizes, and paleosols that are not welded, but separated by the avulsion deposits. Increased sediment supply promotes super-elevation of the channel above its floodplain and makes channels more prone to avulsion (e.g., Bryant et al., 1995; Jones and Schumm, 1999; Leier et al., 2005). This interpretation is also consistent with Foreman's (2014) first scenario for climatic origin of the boundary sandstone.

Foreman's second scenario – decoupling of PETM climate change between the catchment and the northern Bighorn Basin – may explain the boundary sandstone; however, that scenario, by itself, cannot explain various characteristics of the paleosol section. In particular, the episodes of reduced accretion represented by the welded paleosols and the finer grain sizes in the middle of the PETM interval are difficult to reconcile with enhanced erosion and migration of the channel belt. Undoubtedly, reduced vegetation in the depositional basin during dryer periods affected the paleosol section. With sparser vegetation, floodplain surfaces would have been more susceptible to erosion due to increased shear stress from overbank flows (e.g., Griffin and Smith, 2004; Smith, 2004). The abrupt vertical changes and the presence of some erosion

surfaces in the welded paleosols are attributed to reduced vegetation. Additionally, floodplain vegetation can influence grain sizes deposited on the floodplain (e.g., Griffin, et al., 2014). In particular, sand in suspension in overbank flows tends to be transported across sparsely vegetated floodplain surfaces and carried back to the channel rather than being deposited on the floodplain (Griffin et al., 2014). This would have contributed to the lower sand content of the middle of the PETM interval, although the low proportion of avulsion/splay deposits was probably the primary factor influencing grain size of that interval.

Consequently, the overall alluvial architecture – both the nature of channel-belt sandstones and attributes of the paleosols – suggests that PETM climate changes also influenced source areas. This study emphasizes that floodplain paleosols are important for constraining interpretations of alluvial architecture and for understanding the factors that influenced alluvial architecture. As Macklin et al. (2012) pointed out, alluvial deposits not only contain information on how fluvial systems responded to climate change but also provide a record of how the climate changed through time. That is especially true for paleosol-dominated alluvial deposits such as those at Polecat Bench.

9. Conclusions

The PETM interval in the Bighorn Basin provides a remarkable record of changes in the fluvial sedimentary record that can be tied to a high resolution climate record. The trends suggest that changing climate played a significant role in the stratigraphic architecture of the Willwood Formation. Slower rates of sediment accumulation and less sandy sediment in extra-channel settings correspond to the time of warmest temperatures and driest and probably most seasonally dry conditions. Sediment supply depends on sufficient runoff to erode and transport sediment but too little plant cover to impede erosion. We hypothesize that the change to dryer

and more seasonally dry conditions caused reduced vegetation cover in the sediment source area and depositional basin. Decreased plant cover promotes erosion and sediment yield. But, because precipitation was reduced, discharge from source areas was diminished and much of the sediment was stored in upstream reaches of the fluvial system rather than moving to the depositional basin at Polecat Bench.

A return to wetter climates at the end of the PETM and increased discharge caused stored sediment to be flushed to the basin. Increased sediment supply resulted in rivers with more prominent alluvial ridges and encouraged channel avulsion. Higher rates of accretion produced more widely spaced paleosols and thicker avulsion deposits.

Our results demonstrate how vertical sections of alluvial paleosols can provide information on how climate fluctuated through time as well as how the fluvial system responded to that climate change. The paleosols and avulsion deposits, including their contained ichnofauna, provide a high resolution record of climate fluctuations through the PETM interval. The nature and stratigraphic architecture of those deposits records how the alluvial system responded to the climate fluctuations.

Acknowledgements

This material is based upon work supported by the National Science Foundation under Grant Numbers EAR-0228858 and EAR-0958583 to MJK and National Science Foundation Grant Number EAR-02293000 to JJS. We thank two anonymous reviewers for constructive reviews that improved the manuscript.

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Figure Captions

Fig. 1. Map showing exposures of the Willwood Formation (shaded) in the Bighorn Basin and
mountain ranges surrounding the basin. Stars on map and field area inset map (lower left) mark
the locations of the PETM section and core location, both at Polecat Bench, and location of
boundary sandstone in Sand Coulee area studied by Foreman (2014). Index map in upper right
corner shows location of the Bighorn Basin in northern Wyoming, USA. Base map and location
of basin axis modified after Finn et al. (2010).

Fig. 2. Lithologic sections for the PETM interval in northern Bighorn Basin, WY. A. Study
section at Polecat Bench showing paleosols (red, purple, brown, and yellow-brown mudstones)
interbedded with avulsion deposits (gray bands). Marker paleosols of Gingerich (2001) to left
of lithologic column. Dispersed organic matter $\delta^{13}\text{C}$ curve (blue) from Magioncalda et al. (2004)
and pedogenic carbonate (orange) $\delta^{13}\text{C}$ curve from Bowen et al. (2001) and Bains et al. (2003).
B. Bighorn Basin core section from Polecat Bench; stratigraphic levels of PETM subdivisions
based on pedogenic carbonate $\delta^{13}\text{C}$ (Bowen et al., 2015). Marker paleosols of Gingerich (2001)
to right of column. C. Boundary sandstone section from Sand Coulee area with stratigraphic
levels of PETM subdivisions based on pedogenic carbonate $\delta^{13}\text{C}$ (Foreman, 2014).

grain size shown to of left measured section. ss= slickensides; k = carbonate accumulations. See text for details. B. Field photograph of Lower Double Red paleosols showing closely spaced red beds and thin gray units representing splay/avulsion deposits with minimal pedogenesis. Red beds contain abundant gray rhizohaloes. Scale bar is 25 cm.

Fig. 7. Composite lithologic section and facies, grain size, and paleoclimate data for the PETM study interval. Dispersed organic matter $\delta^{13}\text{C}$ curve (blue) from Magioncalda et al. (2004) and pedogenic carbonate $\delta^{13}\text{C}$ curve (orange) from Bowen et al. (2001) and Bains et al. (2003). A. Study section at Polecat Bench showing paleosols (red, purple, brown, and yellow-brown mudstones) interbedded with avulsion deposits (gray bands). Marker paleosols of Gingerich (2001) to left of lithologic column. B. Stratigraphic distribution of five different facies recognized in the lithologic column. Horizontal red lines show five subdivisions based on facies abundance. C. Grain size data for the lithologic section with four subdivisions indicated by blue lines. D. Estimated drainage for paleosols and non-paleosol units based on assemblages of ichnofossils and lithologic features (from Smith et al., 2008). E. MAP estimates for 13 paleosols in the study section. Red triangles show paleosols sampled in outcrop; black circles show paleosols sampled in the core. Error bars for MAP values show standard error for the method (± 108 mm). See text for details. Raw grain size data included in Appendix 1 and facies data in Appendix 3.

Fig. 8. Boxplots of sand, silt, clay, and sand + silt percentages in the four stratigraphic subdivisions of the study section that are based on grain size.

Fig. 3. Ternary diagrams showing quantitative grain size results for the 5 different facies in the study section (classification of Folk, 1954). Figure D shows lithologic designations for different fields of the ternary diagram. S = sandstone; M = mudstone; C = claystone; Z = siltstone; c = clayey, m = muddy, s = sandy.

Fig. 4. Measured stratigraphic sections through representatives of the major facies. Vertical variations in grain size shown to left of each measured section. A. Variegated mudrocks I represented by Top Red A paleosol. B. Variegated mudrocks II represented by Purple 4 paleosol. C. Heterolithic facies represented by multiple beds above Top Red A paleosol. Meter levels to left of each column indicate thickness and stratigraphic level of each measured section. ss= slickensides; k = carbonate accumulations; g = gley features; w = weakly developed B horizon. See text for additional description.

Fig. 5. Facies features. A. Iron depletion zones representing rhizohaloes and well-developed slickensides (yellow arrows) in a red bed of the variegated mudrock I facies. Scale bar is 5 cm. B. Purple 2 paleosol and Upper Double Red paleosols representing variegated mudrock II facies. Scale bar is 5 m. C. Heterolithic facies. Gray silty mudstone with local relict bedding and yellow-brown mottles. Scale shows centimeter markings. D. Heterolithic facies. Gray mudstone with prominent carbonate rhizoliths (red arrows) and pale yellow-brown mottles. Scale bar is 5 cm. E. Laminated facies showing dark gray color and well-preserved unionid bivalve fossil (yellow arrow) with valves still closed suggesting in situ preservation. Scale shows centimeter markings. Orange flecks are paint used to mark field units. F. Channelform facies (black outline). Scale bar is 5 m.

Fig. 6. A. Measured stratigraphic section through Upper Double Red paleosols. Meter levels to left of column indicate thickness and stratigraphic level of the section. Vertical variations in

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Distribution of Sand Percents

Distribution of Silt Percents

Distribution of Sand+Silt Percents

Distribution of Clay Percents

Table 1

Table 1
Median grain size percentages for the four grain size subdivisions of the study section.

Grain Size Subdivisions	Weight %			
	Sand	Silt	Clay	Sand + Silt
4	12.71	54.68	29.51	70.0
3	9.29	54.74	33.59	66.41
2	18.79	56.38	22.84	77.16
1	8.41	49.47	40.79	59.21

See Figure 7C for stratigraphic positions of the four grain size subdivisions. Appendix 3 includes all grain size data.

Table 2

MAP (mm) estimates for the Polecat Bench paleosols.

Stratigraphic Level	Paleosol	MAP (mm)					
		Kraus/ Riggins*	BBCP core	Paleosols analyzed - this study			Average
Post-PETM	Top Red B		1036±108	820±108	1094±108		983±108
Recovery	Top Red A		1094±108	1091±108			1093±108
	Purple-4	1301±108	1240±108	1171±108			1237±108
	Purple-3			1085±108			1085±108
	Upper Double Red B	1186±108	832±108				1009±108
Main Body of PETM	Upper Double Red A		1157±108				1157±108
	Red above Purple-2		850±108				850±108
	Purple-2	1008±108	1034±108	987±108	1017±108	1015±108	1012±108
	Lower Double Red B	1114±108	1045±108	828±108	1097±108	1070±108	1031±108
	Lower Double Red A	755±108	516±108				636±108
Onset	Purple-0	1065±108		1110±108			1088±108
Pre-PETM	Red below Purple-0	1163±108					1163±108
	Purplish-red	1139±108		1208±108	1153±108	1181±108	1170±108

*Results from Kraus/Riggins (2007). See Appendix 2 for raw data and MAP calculations.

Table 3
The stratigraphic section is subdivided into five vertical intervals based on differences in abundance of the five different facies as measured by thickness in cm.

Vertical Interval	Five Facies				
	Variegated Mud I	Variegated Mud II	Heterolithic	Laminated Mud	Channel-form
5	612	565	1584	31	16
4	193	1599	666	0	0
3	316	314	1051	20	0
2	0	0	338	52	304
1	287	250	232	69	0

Appendix 3 contains facies data for all stratigraphic levels.

Table 4
The five vertical intervals of the stratigraphic section based on differences in abundance of the five facies as measured by thickness percent in a particular vertical interval.

Vertical Interval	Five Facies				
	Variegated Mud I	Variegated Mud II	Heterolithic	Laminated Mud	Channel-form
5	21.8	20.1	56.4	1.1	0.6
4	7.8	65.1	27.1	0	0
3	18.6	18.5	61.8	1.2	0
2	0	0	48.7	7.5	43.8
1	34.2	29.8	27.7	8.2	0

Appendix 3 contains facies data for all stratigraphic levels.