

Magnetostratigraphy of the Ludlow Member of the Fort Union Formation (Lower Paleocene) in the Williston Basin, North Dakota

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ABSTRACT

To determine the geomagnetic polarity stratigraphy and the duration and age of the Ludlow Member of the Fort Union Formation (Lower Paleocene), we constructed a 325 m composite lithostratigraphic section of the Upper Cretaceous Hell Creek Formation and the Lower Paleocene Ludlow and Tongue River Members of the Fort Union Formation in the Little Missouri River valley of North Dakota, USA. We analyzed paleomagnetic samples from nine of the logged sections. The principal magnetic carrier in the Ludlow Member sediments is likely titanomagnetite, as indicated by predominantly irreversible thermomagnetic curves measured from sandstone, siltstone, and carbonaceous shale samples. The analyzed paleomagnetic samples document a series of polarity zones that can be correlated from C29n to C27r on the geomagnetic polarity time scale (GPTS). We infer that the magnetization of the samples is primary because the characteristic directions are consistent with those of the Paleocene of North America, and the reversal stratigraphy from this section corresponds to the GPTS with reasonable sediment accumulation rates. By extrapolating the measured sediment accumulation rate from the Cretaceous-Tertiary (K-T) boundary to the top of C28n and then to the top of the Ludlow Member, we estimate the duration of the member to range from 2.31 to 2.61 m.y. This is the first estimate for the duration and age of the Ludlow Member, and it can be used as an important tool for interpreting rates of biotic recovery after the K-T extinction.

Keywords: Paleocene, magnetostratigraphy, magnetic mineralogy, Fort Union Formation, Ludlow Member, Williston Basin.

INTRODUCTION

In the Little Missouri River valley of North Dakota, a continuous succession of Cretaceous through lowermost Eocene terrestrial sediments contains a nearly complete Paleocene record (Fig. 1). These sediments contain a rich fossil record that has been the focus of many lithostratigraphic, magnetostratigraphic, and paleontologic studies relating to the biota before and after the Cretaceous-Tertiary (K-T) boundary extinctions (e.g., Belt et al., 1984; Hicks et al., 2003; Hunter and Archibald, 2002; Hunter et al., 2003; Johnson, 2002; Nichols, 2002; Nichols and Johnson, 2002; Pearson et al., 2001, 2002). Paleontologic and paleomagnetic research has focused on the K-T boundary interval, however, and no paleomagnetic studies have been conducted on the entirety of the Ludlow Member of the Fort Union Formation. Thus, the duration of the early Paleocene Ludlow Member, an important metric for estimating speciation rates and changes in biodiversity of plants and mammals following the K-T extinctions, has never been determined.

PREVIOUS STUDIES

Over the past 150 yr, the lower Fort Union Formation has been the focus of many lithostratigraphic studies. Due to the predominance of coal in the early Paleocene succession, most early studies were focused on the general stratigraphy of the region and the economic geology (Calvert, 1912; Laird and Mitchell, 1942; Leonard, 1908, 1911; Meek and Hayden, 1861; Thom and Dobbin, 1924). Since the middle of the twentieth century, researchers

have been more focused on the depositional environments and local stratigraphy of the upper Hell Creek Formation and the Ludlow and lower Tongue River Members of the Fort Union Formation (Belt et al., 1984; Brown, 1962; Chevron and Jacobs, 1985; Fastovsky, 1987; Fastovsky and McSweeney, 1987; Frye, 1969; Hartman, 1993, 1989; Johnson et al., 2002; Moore, 1976; Murphy et al., 1995, 2002; Warwick et al., 2004). More recently, studies have focused on recognizing the K-T boundary biostratigraphically and geochemically (Arens and Jahren, 2002; Johnson, 1992, 2002; Johnson et al., 1989; Nichols and Johnson, 2002; Pearson et al., 2002).

The upper Hell Creek Formation and the lower Ludlow Member have also recently been the focus of several paleomagnetic studies. The first magnetostratigraphic study of the upper Cretaceous-lower Paleocene sediments in the Williston Basin was conducted on four stratigraphic sections in eastern Montana (Archibald et al., 1982). The published magnetostratigraphy indicated two normal polarity intervals bracketing a reversed interval; however, the magnetostratigraphic sections were not long enough to be correlated to the geomagnetic polarity time scale (GPTS). Archibald et al. (1982) identified black, opaque minerals in the sediments, which they interpreted to be magnetite or titanomagnetite and inferred that magnetite or titanomagnetite was the detrital remanence-bearing magnetic mineral. Subsequent paleomagnetic work on the same area, coupled with a correction of plotting errors and a series of ⁴⁰Ar/³⁹Ar ages from the Paleocene sequence by Swisher et al. (1993), related the section to the intervals C30n through C28n. In addition, Swisher et al. (1993) further refined the interpretations of the detrital magnetic mineralogy and demonstrated that the dominant ferromagnetic mineral was an intermediate-

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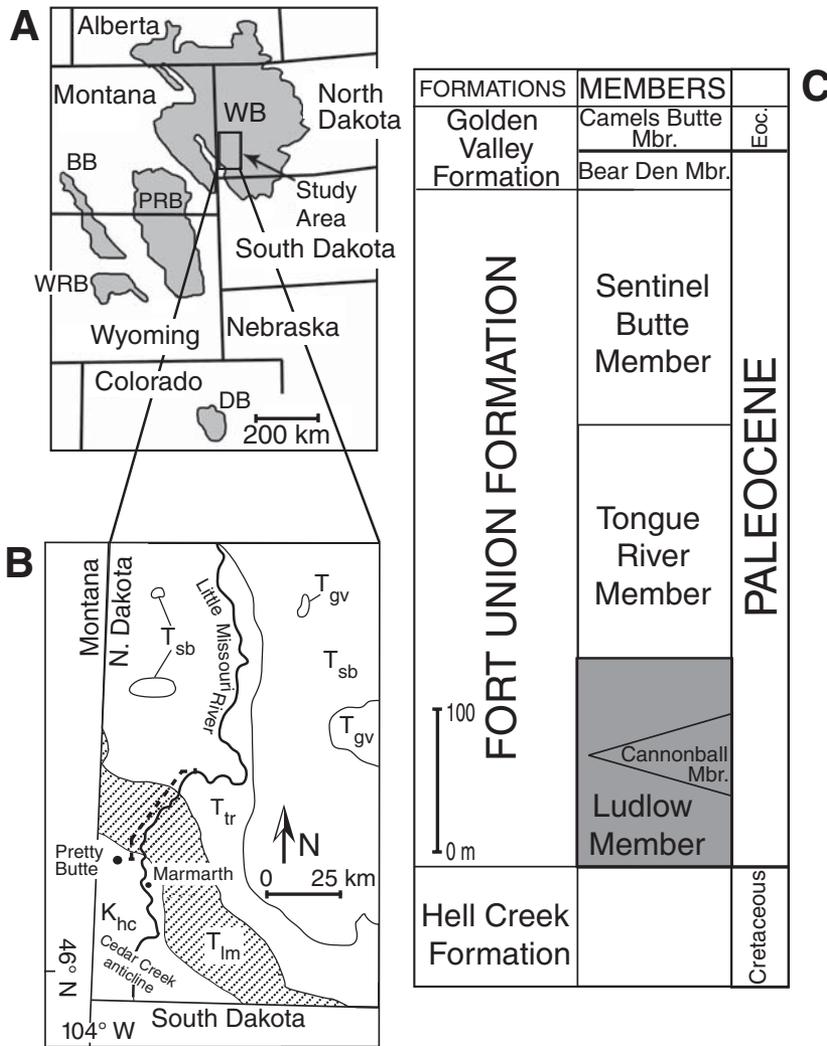


Figure 1. (A and B) Geologic map of study area in Williston Basin (WB), northern Great Plains, United States. Dashed black line indicates line of section measured and sampled through the Ludlow Member of the Fort Union Formation. K_{hc} —Hell Creek Formation, T_{lm} —Ludlow Member, T_{tr} —Tongue River Member, T_{sb} —Sentinel Butte Member, T_{gv} —Golden Valley Formation. (C) Generalized stratigraphy of Paleocene and Eocene section of the Williston Basin in southwestern North Dakota. BB—Bighorn Basin; DB—Denver basin; Eoc.—Eocene; PRB—Powder River Basin; WRB—Wind River Basin.

composition titanohematite. Lerbekmo and Coulter (1984) completed one paleomagnetic section around the K-T boundary in central Montana and recognized two normal and two reversed polarity intervals. They interpreted the lowermost reversed interval, which contained the K-T boundary, to be C29r. Lund et al. (2002) studied the magnetostratigraphy of four sections of the Fox Hills Formation, the Hell Creek Formation, and the lowermost Fort Union Formation in eastern Montana and central North Dakota. They then made regional correlations of the Hell Creek–Fort Union contact and the placement of the K-T boundary

within C29r in their sections. They also conducted rock-magnetic analyses and concluded that the primary detrital magnetic mineral carrying the sediments’ magnetic remanence was hemo-ilmenite and that secondary viscous and chemical overprints were routinely present. Hicks et al. (2002) conducted a magnetostratigraphic and geochronologic study on six sections that included the K-T boundary in southwestern North Dakota. That study noted the position of the K-T boundary within C29r and revised the age estimate for the K-T boundary using their magnetostratigraphy and new normalized isotopic ages.

LOCAL STRATIGRAPHY

In the Little Missouri River Valley of western North Dakota, the Paleocene Fort Union Formation conformably overlies the terrestrial Upper Cretaceous Hell Creek Formation. The Ludlow, Tongue River, and Sentinel Butte Members of the Fort Union Formation are almost entirely terrestrial, with the exceptions of two brief intervals of marine deposition: the Cannonball Member, which interfingers with the Ludlow Member, and an unnamed marine incursion in the Tongue River Member (Belt et al., 1984, 2005; Clayton et al., 1977; Cvancara, 1972, 1976; Kroeger and Hartman, 1997; Moore, 1976; Warwick et al., 2004). The remainder of the Paleocene section consists of fluvial deposits interbedded with abundant, thick, laterally continuous lignite deposits.

In southwestern North Dakota, the K-T boundary is approximately synchronous with the Hell Creek–Fort Union formational contact. In Bowman County, the K-T boundary and the formational contact are coincident, and to the north in Slope County, the boundary is within the basal three meters of the Fort Union Formation (Nichols and Johnson, 2002). The lithostratigraphic contact between the Hell Creek and Fort Union Formations is typically placed at the base of the first laterally extensive lignite bed, or, if a lignite is not present, at the sharp transition from smectite-rich, “popcorn” weathered, drab gray claystone of the Hell Creek to carbonaceous shale and laminated yellow and brown mudstone or sandstone of the Fort Union (Brown, 1962; Fastovsky, 1987; Frye, 1969; Hares, 1928; Laird and Mitchell, 1942; Leonard, 1911; Moore, 1976; Murphy et al., 1995, 2002; Nichols and Johnson, 2002). Based on the close proximity of the K-T boundary to the easily recognizable formational contact, the approximate location of the boundary can be readily located in the field.

There is a distinct lithologic difference between the upper and lower parts of the Ludlow Member (see following discussion). However, there are few thick exposures of both the upper and lower parts of Ludlow Member. For this reason, Moore (1976) chose a laterally extensive lignite to correlate sections of predominantly lower Ludlow strata to sections of predominantly upper Ludlow deposits. In his correlation, Moore incorrectly attributed this bed to the T-Cross Coal of Hares (Hares, 1928; see Hartman [1989] for a detailed discussion). This name has been used many times in subsequent publications on the Ludlow Member (e.g., Belt et al., 2004; Warwick et al., 2004). Therefore, in this paper, we will refer to the marker bed between the upper and lower Ludlow Member

as the “T-Cross” coal with quotation marks to denote incorrect but common usage.

Early in the study of the Ludlow Member (e.g., Leonard, 1908), brackish units were recognized within the Ludlow Member, and many subsequent studies continued to informally recognize them as tongues of the marine Cannonball Member of the Fort Union Formation (e.g., Belt et al., 1984; Cvancara, 1976; Van Alstine, 1974). Hartman (1993) proposed formal stratigraphic names for these brackish tongues: the Boyce Tongue and the Three V Tongue. The Boyce Tongue is stratigraphically above the “T-Cross” coal and is recognized by the presence of the bivalve *Corbicula*. The Three V Tongue is a thick siltstone deposit full of oyster fossils stratigraphically near the middle of the upper Ludlow Member. It is laterally continuous over a few kilometers.

The Rhame zone (Belt et al., 2004), also known as the Rhame bed (Wehrfritz, 1978) or the “white marker zone” (Belt et al., 1984; Clayton et al., 1977; Moore, 1976), consists of laterally extensive white-colored sediments and interdispersed silcrete beds that mark the contact between the Ludlow and Tongue River Members.

Many laterally extensive lignites have been given names and have been used to make correlations across great distances (see, for example, Belt et al., 1984, 2004; Warwick et al., 2004). In this study, we did not attempt to use any of the names given to the lignite deposits, other than the easily distinguished “T-Cross” coal, to prevent unintentionally misnaming or incorrectly determining a deposit and thus causing miscorrelation of stratigraphic units.

LITHOSTRATIGRAPHY

Section Measurement

Using the Hell Creek–Fort Union formational contact as the basal reference datum for the composite section, 35 lithostratigraphic sections were measured during field work in 2004, 2005, and 2006. Using these sections, we constructed a 325 m composite section of the upper Hell Creek Formation and the Ludlow and lower Tongue River Members of the Fort Union Formation (Fig. DR1¹). All sections were trenched at least 50 cm deep to remove weathered material and expose the lithologic contacts and undeformed bedding. Each section was logged systematically to the nearest centimeter. Stratigraphic relationships between sections were

determined by physically tracing marker beds and measuring numerous parallel and overlapping sections. Construction of such a composite section is necessary in the Little Missouri River valley because strata dip at most a few degrees (generally to the northeast), and badland relief is typically only 50–100 m.

Paleomagnetic samples were systematically collected from nine parallel and overlapping lithostratigraphic sections along a southwest-northeast transect during field seasons in 2004 and 2005. These sections were chosen in order to reduce the likelihood of missing any short-lived subchrons and to determine the full lithostratigraphic thickness of all magnetozones.

The sampled sections have been given the following informal names for clarity of discussion (global positioning system [GPS] locations are listed in Table DR1 [see footnote 1]; numerical codes refer to field designation of each section): Bald Butte (05–40), Never-Ending Butte (04–15), John’s Nose (04–01), Upper John’s Nose (04–04), Lonesome Bull (04–30), Bug Butte (05–06), Three V Butte (05–33), Three V Amphitheater, (05–14), and Far Butte (05–19) (Fig. DR1, see footnote 1). For Bald Butte, Never-Ending Butte, and John’s Nose, the reference horizon for correlation between sections was defined as the formational contact between the Hell Creek Formation and the Ludlow Member. At Lonesome Bull and Bug Butte, the “T-Cross” coal was used as the reference horizon for correlation. At Far Butte, the Tongue River–Ludlow lithologic contact was used as the reference horizon, and at Upper John’s Nose, Three V Butte, and Three V Amphitheater, local distinctive marker beds were used as the reference horizons.

Stratigraphic Sections

(1) At Bald Butte, the position of the K-T boundary is uncertain because there are no definitive Paleocene fossils above the formational contact. The majority of the section is interpreted to be Cretaceous based on the presence of numerous dinosaur fossils and Cretaceous megafloreal taxa. The section is 41 m thick; 37 m are in the Hell Creek Formation, and 4 m are in the Ludlow Member. Twenty-two horizons were collected through the section. The mean sampling interval was 1.8 m, with a maximum interval of 7.5 m and a minimum interval of 0.40 m.

(2) At Never-Ending Butte, the approximate stratigraphic position of the K-T boundary was determined by the presence of dinosaur bones ~5 m below the lithologic contact and the presence of Paleocene megafloreal taxa ~5 m above the contact. Furthermore, this section was the New Facet Boundary section in Nichols and

Johnson (2002), who noted that the K-T boundary was 199 ± 1 cm above the formational contact. The complete section is 77 m thick; 7 m are in the Hell Creek, and 70 m are in the Ludlow Member. The upper 6 m of the section are interpreted to be the “T-Cross” coal. Seventeen horizons were collected through the section. The mean sampling interval was 4 m, with a maximum interval of 11 m and a minimum of 0.80 m.

(3) At John’s Nose, we estimated the K-T boundary to be within 1.5 m of the formational contact because Paleocene leaves and vertebrate fossils were found at 1.5 and 15 m above the formational contact, respectively (Pearson et al., 2004; Peppe et al., 2006). The section is 30 m thick; 1 m is in the Hell Creek Formation, and 29 m are in the Ludlow Member. Ten horizons were collected from this section. The mean sampling interval was 2 m, with a maximum interval of 3 m and minimum of 1 m.

(4) Upper John’s Nose is 12.5 m thick, is laterally adjacent to John’s Nose, and is easily related by several beds. The entire section is within the Ludlow Member. Four horizons were collected from this section. The mean sampling interval was 3 m. The maximum interval was 6 m, and the minimum interval was 1 m.

(5) Lonesome Bull is 64 m thick and entirely within the Ludlow Member. The section can be related to the other sections based on the presence of the “T-Cross” coal near the base of the section. Furthermore, this section contains both marine tongues of the Cannonball Member. Eighteen horizons were collected from this section. The mean sampling interval was 3.2 m, with a minimum interval of 0.25 m and a maximum interval of 15 m.

(6) Bug Butte is 87 m thick and is entirely within the Ludlow Member. The section contains both tongues of the Cannonball Member and has the “T-Cross” coal at its base. Twenty-three horizons were collected from this section. The mean sampling interval was 3.4 m, the maximum sampling interval was 18.95 m, and the minimum interval was 0.50 m.

(7) Three V Butte is 38.5 m thick and entirely within the upper Ludlow Member. This section was related to sections Lonesome Bull, Bug Butte, and Three V Amphitheater by several laterally continuous lignite deposits found in all four sections. Fifteen horizons were collected from this section. The mean interval was 2.3 m, the maximum interval was 12 m, and the minimum interval was 0.15 m.

(8) Three V Amphitheater is 60 m thick and entirely within the upper Ludlow Member. This section was related to sections Lonesome Bull, Bug Butte, and Three V Butte by several laterally continuous lignite deposits found in all four

¹GSA Data Repository Item 2008172, which contains additional material related to this article, is available at www.geosociety.org/pubs/ft2008.htm. Requests may also be sent to editing@geosociety.org.

sections. Twenty-four horizons were collected from this section. The mean sampling interval was 1.6 m, the maximum interval was 7.5 m, and the minimum was 0.45 m.

(9) Far Butte is 62 m thick; 35 m are within the upper Ludlow Member, and 27 m are within the Tongue River Member. The Tongue River contact was used as the reference datum for this section, and it was related to the Three V Amphitheater by two widespread and laterally continuous lignite and sandstone beds. Thirty-three horizons were collected from this section. The mean interval was 1.75 m, the maximum interval was 6.85 m, and the minimum sampling interval was 0.35 m.

Lithostratigraphic Interpretations

Our lithostratigraphic work demonstrates that the Ludlow Member is 190–210 m thick, and that the Boyce Tongue and the Three V Tongue are 75 and 115 m above the Hell Creek–Fort Union formational contact, respectively (Fig. DR1, see footnote 1). The “clinker” or “scoria” that caps many of the buttes along Cannonball Creek is ~75 m above the Hell Creek–Fort Union formational contact. This clinker horizon can be correlated to the “T-Cross” coal and, thus, the Boyce Tongue.

As noted in many previous studies (e.g., Belt et al., 1984; Chevron and Jacobs, 1985; Hartman, 1989, 1993; Johnson et al., 2002; Moore, 1976; Murphy et al., 1995, 2002; Warwick et al., 2004), our section logging indicates that the Ludlow Member is composed of alternating beds of poorly lithified light yellow to brown sandstone and siltstone, rare mudstone, numerous carbonaceous shale beds, and thick (>0.5 m) lignite deposits. The Ludlow Member is also highly fossiliferous with plant remains. Our data also indicate that there is a demonstrable shift between the lower and upper Ludlow Member from thin lignites (<1.0 m) and common siltstones and sandstones in the lower Ludlow to thick lignite deposits (>1.0 m) and thick coarse- and medium-grained sandstones in the upper Ludlow Member.

METHODOLOGY

Paleomagnetism

Samples were collected by excavating a deep pit to remove any weathered material. At each stratigraphic horizon described herein, four to five separately oriented block samples were collected. A flat face was shaved on the in situ samples with a hand rasp, and then the strike and the dip of the face were measured with a Brunton pocket transit compass. Samples were collected

from a range of lithologies and grain sizes from mudstone to medium sandstone. In the laboratory, at least three cuboid specimens from each interval were either hand cut with a diamond-bit saw or dry sanded into ~10 cm³ volumes.

Magnetic moment measurements were done at Yale University using an automated three-axis DC-SQUID cryogenic magnetometer housed inside a three-layer magnetostatic shield with a background magnetic field less than 200 nT. Samples were demagnetized using a combined alternating-field (AF) and thermal demagnetization strategy (Schmidt, 1993). First, we applied a low-AF pretreatment to remove any low-coercivity viscous or isothermal remanence. Next, 10–20 thermal demagnetization steps were performed from 75 °C to the maximum unblocking temperature (typically 250–400 °C). Thermal demagnetization was done in a nitrogen atmosphere, and AF demagnetization was conducted with in-line automated static coils. Instability of magnetization above 200 °C has been reported in previous studies, which we reproduced when heating pilot samples in air. However, when the samples were heated in nitrogen, the stability field extended to well above 300 °C. Progressive demagnetization was carried out until the magnetic intensity of the samples fell below noise level or until the measured directions became erratic and unstable.

The characteristic remanence for samples with quasi-linear trajectories was isolated using principal-component analysis (PCA) (Kirschvink, 1980). The best-fit line was used if defined by at least three consecutive demagnetization steps that trended toward the origin and had a maximum angle of deviation (MAD) less than 20° (Figs. 2A and 2B). Specimens that were analyzed by great circles were used if they had a MAD less than 20° (Figs. 2C and 2D). In specimens with directions that clustered around one point (e.g., Fig. 2E) but did not decay toward the origin, we selected at least four consecutive points, anchored to the origin, to define the characteristic remanence component. These data were also filtered with a cutoff MAD value of 20°. If a single horizon (i.e., paleomagnetic “site”) had one or two specimen directions that were calculated by PCA, the great circles and/or best-fit lines were combined using the method of McFadden and McElhinny (1988) to compute a mean direction. The mean direction of each horizon with three or more statistically significant directions was then calculated using Fisher statistics (Fisher, 1953). Sites that had an alpha 95 (α_{95}) value greater than 35°, which exceeds the cutoff value based on the randomness criteria of Watson (1956), were excluded. Data from specimens that had erratic demagnetization behavior were also excluded (Fig. 2F).

Reversal boundaries were placed at the stratigraphic midpoints between samples of opposing polarity. The stratigraphic position of each reversal was calculated relative to the Hell Creek–Fort Union formational contact (Table DR2, see footnote 1). The reversals were placed with a precision of 0.5–7.5 m, depending on sample spacing due to outcrop exposure constraints. The resulting polarity stratigraphy was then correlated to the geomagnetic polarity time scale (GPTS) (Ogg and Smith, 2004).

Rock Magnetism

Isothermal remanent magnetization (IRM) acquisition was performed using a 2G capacitor relay box, modulated by custom-built transformer boxes, and measured using the cryogenic magnetometer. Temperature dependence of low-field magnetic susceptibility, $k(T)$, was measured upon cycling through the range of –192 to 700 °C in argon, using an AGICO KLY-4S magnetic susceptibility meter equipped with a high-temperature furnace and a cryostat. Room-temperature magnetic hysteresis loops were obtained using a Princeton Measurement vibrating sample magnetometer at the Institute for Rock Magnetism (University of Minnesota). We also examined the magnetic mineralogy of our samples using an XL-30 Environmental Scanning Electron Microscope (e-SEM) at the Yale Department of Geology and Geophysics. Backscattered electron (BSE) imaging was used to identify oxide grains. The compositions of these grains were determined by means of energy-dispersive spectrometry (EDS).

RESULTS

Magnetostratigraphy

Many specimens’ demagnetization trajectories turned toward the origin after a few steps and were fully demagnetized by 250–400 °C (Figs. 2A and 2B). A subset of the samples had demagnetization trajectories that were best characterized by a great circle, and thus only a plane containing the characteristic component of the samples was defined (Figs. 2C and 2D). Other samples clustered around one point, behaved erratically before trending toward the origin, or did not display any coherent behavior (Figs. 2E and 2F).

We analyzed 432 samples from 159 sampling horizons. 348 of those samples, from 143 sampling horizons, passed our selection criteria. Twenty-three of the sampling horizons (e.g., sites with three or more samples with statistically significant directions that could be used to calculate a site mean with an $\alpha_{95} < 35^\circ$) passed

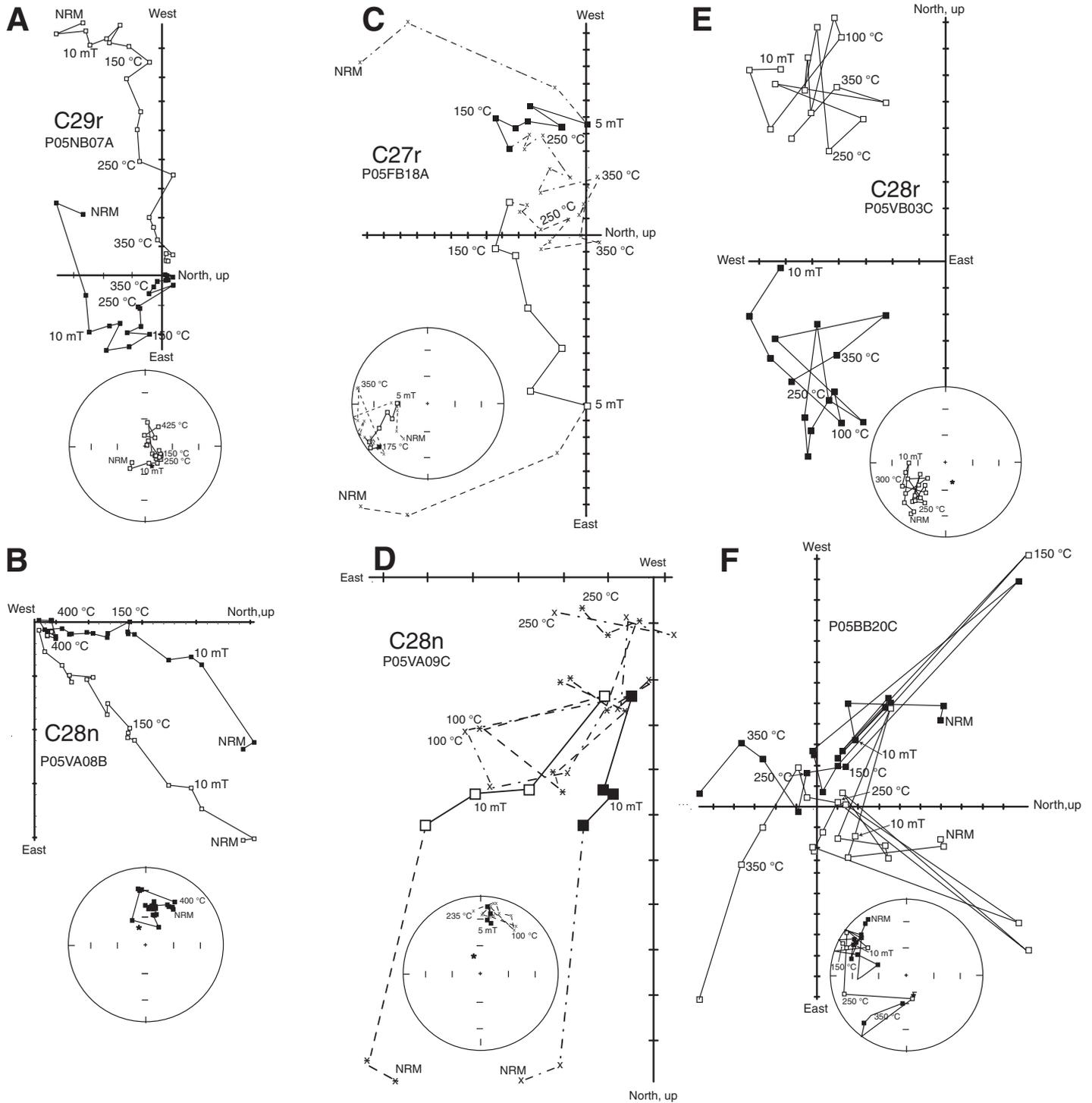


Figure 2. Representative Zijderveld diagrams and equal-area plots for each subset of data. (A) Demagnetization trajectory of reversed polarity sample from C29r (representative of ~25% of the data). (B) Demagnetization trajectory of normal polarity sample from C28n (representative of ~35% of the data). (C) Reversed polarity sample with demagnetization direction characterized as a great circle from C27r (representative of ~20% of the data). (D) Normal polarity sample with demagnetization direction characterized as a great circle from C28n (representative of ~5% of the data). (E) Reversed sample clustered around one point from C28r. Characteristic remanence component was calculated by selecting at least four consecutive points (representative of ~5% of the data). (F) Representative of ~10% of the data that yielded no useable data. NRM—natural remanent magnetization.

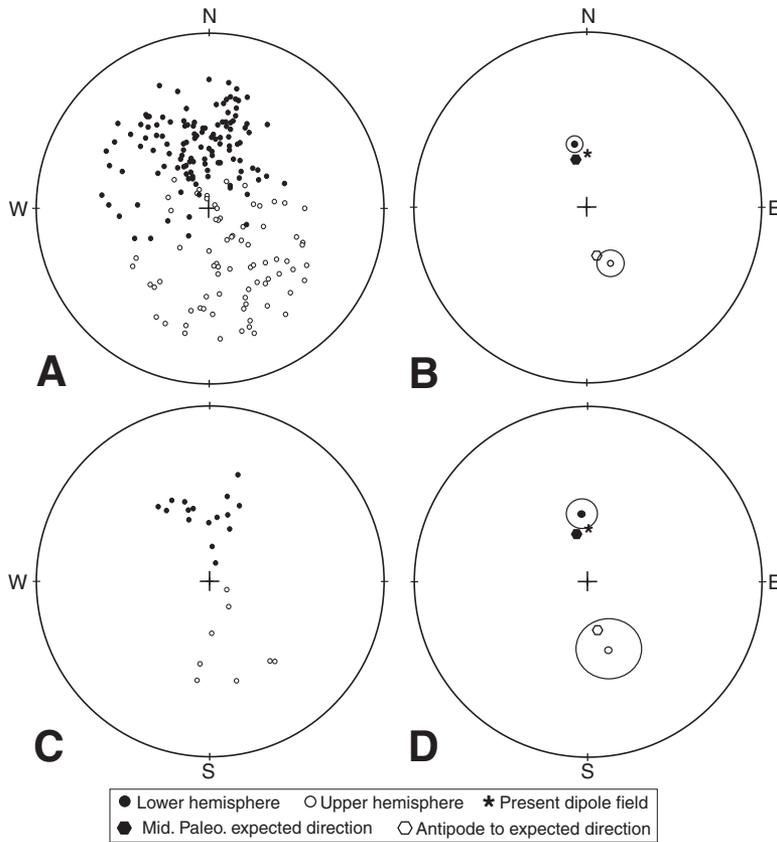


Figure 3. (A) Equal-area plot of characteristic magnetization directions calculated from all lines measured in this study. (B) Mean normal and reversed polarity directions from lines, plotted with present-day field direction, the expected middle Paleocene direction, and the antipode to the expected middle Paleocene direction. Circle around mean direction represents 95% confidence limit (Fisher, 1953). (C) Equal-area plot of all site mean directions. (D) Mean normal and reversed directions of site means, plotted with present-day field direction, the expected middle Paleocene direction, and the antipode to the expected middle Paleocene direction. Circle around mean direction represents 95% confidence ellipse (Fisher, 1953).

TABLE 1. MEAN PALEOMAGNETIC DIRECTIONAL DATA FROM THE LUDLOW MEMBER OF THE FORT UNION FORMATION

Subset	<i>n</i>	D (°)	I (°)	k	α_{95} (°)	Pole (°N)	Pole (°E)	K	A95 (°)
C30n - lines	11	355.5	59.5	15.5	12	85.7	108.1	9.4	15.7
C29r - lines	32	150.7	-54.4	8.86	9.1	66.6	158.7	5.6	11.8
C29n - lines	31	340.1	62.2	11.5	8.0	77.4	170.3	6.5	11.0
C28r - lines	30	151.4	-69.7	8.8	9.4	68.5	209.3	4.5	14.1
C28n - lines	82	351.1	58.6	11.1	4.9	82.3	132.0	6.5	6.6
C27r - lines	13	180.4	-56.2	7.0	16.8	84.9	069.6	4.2	23.2
All normal - lines	124	349.0	59.7	11.5	3.9	81.9	144.6	6.7	5.3
All normal - sites	16	355.7	58.1	28.5	7.0	83.0	101.3	16.7	9.3
All reversed - lines	75	156.6	-61.3	7.9	6.2	73.2	178.5	4.6	8.7
All reversed - sites	8	162.3	-56.1	15.0	14.8	75.3	146.6	9.6	18.8
Dual polarity - lines	165	342.4	61.4	10.0	3.7	77.9	169.1	5.7	5.1
Dual polarity - sites	23	350.5	57.4	21.5	6.7	80.8	126.2	13.0	8.7

Note: *n*—number of lines or sites included in the mean; D—declination; I—inclination; k—Fisher's (1953) precision parameter; α_{95} —radius of 95% confidence cone around mean (Fisher, 1953); pole N and E—mean of virtual geomagnetic poles calculated from each line or site mean; K and A95—Fisher statistics of paleomagnetic pole. Reversed polarity paleomagnetic poles have been inverted to the Northern Hemisphere for comparison with normal polarity poles. Dual-polarity lines and sites are for C29r (Paleocene only)–C28n, inclusive.

our site mean selection criteria. The site means and data from all lines and the site means (combining lines and planes; McFadden and McElhinny, 1988) at each statistically robust sampling horizon (α_{95} value < 35°) are plotted on the equal-area projections in Figure 3 and Table 1 (see Tables DR3 and DR4 for data from all specimens and site means [see footnote 1]). The mean normal and reversed directions calculated using the Fisher (1953) statistics are also shown, surrounded by their 95% confidence circles. The mean normal declination and inclination for lines are 349.0° and 59.7° ($n = 124$; $\alpha_{95} = 3.9^\circ$). For sites, the mean normal declination and inclination are 355.7° and 58.1° ($n = 16$; $\alpha_{95} = 7.0^\circ$). The mean reversed declination and inclination for lines are 156.6° and -61.3° ($n = 7.9$; $\alpha_{95} = 6.2^\circ$); for sites, the mean reversed declination and inclination are 162.3° and -56.1° ($n = 8$; $\alpha_{95} = 14.8^\circ$). The dual-polarity mean directions (e.g., all declinations and inclinations converted to normal polarity, chrons C29r–C28n, inclusive) for lines are 342.4° and 61.4° ($n = 165$; $\alpha_{95} = 3.7^\circ$), and for sites, mean directions are 350.5° and 57.4° ($n = 23$; $\alpha_{95} = 6.7^\circ$). Using the mean normal and reversed directions for either lines or sites, the null hypothesis of antiparallelism cannot be rejected at the 95% confidence level (i.e., positive reversals test). Mean directions from lines yield a class B positive reversals test (McFadden and McElhinny, 1990). Due to the smaller number of data points and consequently larger α_{95} values, the reversals test for mean sites is a class C positive reversals test.

Using the dual-polarity mean direction, we calculated a paleomagnetic pole for both lines and sites. The pole for lines is 77.9°N, 169.1°E ($A95 = 5.1^\circ$), and for sites, it is 80.8°N, 126.2°E ($A95 = 8.7^\circ$). Our preferred mean is from sites rather than lines because the larger uncertainty of the site-mean pole reflects a more conservative choice that is less susceptible to type I statistical error when comparing it to other poles and because a two-tiered averaging procedure would be needed to treat sites that yielded more than one line. These pole locations for lines and sites are significantly distinct from the Paleocene reference paleomagnetic pole for North America at 60 Ma (Besse and Courtillot, 2002, 2003) (Fig. 4). This difference can be explained by two possibilities. First, it could result from inclination shallowing in our data set, with a flattening factor of ~10%. If we were to correct for this possibility, our pole and the reference pole would become statistically indistinct. Second, because the reference pole is the result of a sliding time window that incorporates data from both North America and other continents linked by the global plate circuit (in this particular instance, poles from the Deccan Traps in India), there is

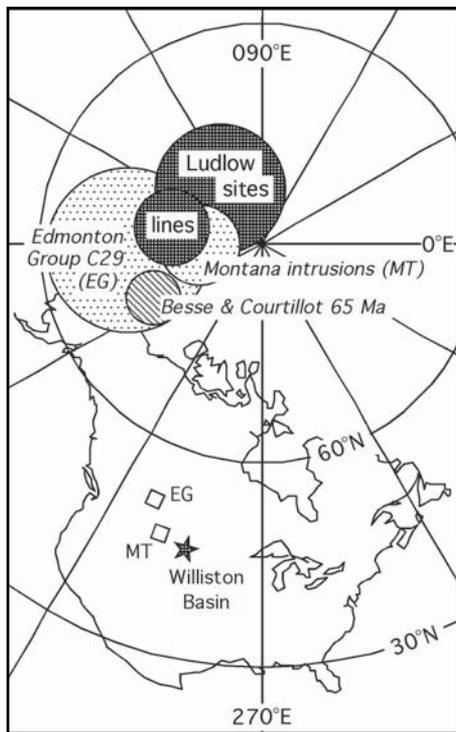


Figure 4. Polar equal-area projection of Ludlow Member (chrons 29r through 28n) paleomagnetic poles computed from quality-filtered lines and sites (Tables DR3 and DR4, see text footnote 1) compared with relevant early Paleocene paleomagnetic poles from interior North America. Edmonton Group pole (chron 29n/r boundary mean) is from Lerbekmo and Coulter (1985); Montana intrusions pole (61–67 Ma) is from Diehl et al. (1983); Besse and Courtillot pole (65 Ma, sliding window) is from Besse and Courtillot (2003).

the possibility that it is biased by small errors in either the seafloor-spreading reconstructions or the assumption of a pure geocentric axial dipole (GAD) field model. Our pole positions for sites and lines do overlap with poles from the North American craton in the early Paleocene (Diehl et al., 1983; Lerbekmo and Coulter, 1985) (Fig. 4). Additional analysis is needed to determine the extent to which inclination shallowing affects the paleomagnetic pole direction.

The individual specimen directions, the site means, and the magnetic polarity stratigraphy of all analyzed sections are plotted in Figures 5, 6, and 7. The composite summary magnetostatigraphy is plotted in Figure 8. Polarity zones are labeled from A+ at the bottom to F- at the top. The stratigraphic position of each section is shown in Figure DR1 and listed in Table DR3 (see footnote 1).

One of the difficulties in interpreting the polarity stratigraphy of the Ludlow Member is the existence of anomalous, normal polarity intervals (e.g., within zone D-, Lonesome Bull and Bug Butte, Fig. 6). This feature was also noted by Hicks et al. (2002) in their study of the Hell Creek Formation and lowermost Ludlow Member. None of these anomalous, normal intervals is recognized within the GPTS, or near the K-T boundary as hypothesized by Lerbekmo et al. (1996), so they seem unlikely to represent “real” short-lived normal polarity chrons.

More likely, they are the result of dominant overprinting due to secondary precipitation of goethite and/or hematite as suggested by Hicks et al. (2002). In many of the anomalous samples, there was evidence of groundwater penetrations, outcrop cracking, and/or subsurface weathering, which likely allowed the secondary precipitation of goethite and/or hematite to take place. Additionally, the presence of goethite in the anomalous samples is demonstrated in Figure 9, which shows characteristic J/J_0 plots and Zijderveld diagrams of typical reversed polarity and anomalous normal polarity samples. In the anomalous samples, there is a noticeable drop in magnetic intensity between 100 and 200 °C (Fig. 9C), and the samples demonstrate random demagnetization behavior above 150 °C (Fig. 9D). These data indicate that goethite is the dominant ferromagnetic mineral in many of the anomalous samples. Furthermore, some of the anomalous samples were collected from medium- to coarse-grained sandstones. In addition to being porous and allowing water penetration and precipitation of secondary goethite and/or hematite, it is also likely that these coarser-grained sediments contain large, multidomain magnetic minerals that further complicate the demagnetization behavior. Based on the evidence for water penetration and subsurface weathering and the typical demagnetization behavior of the anomalous samples, we conclude that the apparent normal polarity directions within our D- and F- magnetozones represent secondary overprints.

Because we controlled the stratigraphic position of each magnetostatigraphic horizon at the centimeter scale, we can precisely place the upper and lower bounds of each polarity zone, within the limits of our ability to sample the outcrops. For example, at both the Lonesome Bull and Bug Butte sections, the upper and lower reversal patterns are almost identical, whereas the anomalous overprints are found in different stratigraphic positions. Furthermore, the occurrence of the D-/E+ reversal in the same stratigraphic position in sections at Lonesome Bull, Bug Butte, and Three V Butte demonstrates the likely accuracy of our magnetostatigraphic interpretation (Fig. 6).

We also noted two, single-site reversed polarity intervals within polarity zone E+ (Three V Amphitheater, Fig. 7). Both of these reversed horizons are represented by one best-fit line. We did not reproduce these results in any other overlapping sections and consider it unlikely that they represent true, short-lived reversed polarity chrons.

Rock Magnetism

All measured samples showed irreversible behavior of low-field magnetic susceptibility versus temperature (Fig. 10). Most samples show a dominant Curie temperature (T_c) between 450 and 580 °C. The cooling curves of most samples also show a large increase in susceptibility at 400–500 °C. In some samples, there is a slight inflection of slope at ~180–200 °C. The irreversible $k(T)$ curves are consistent with the inversion of initial titanomaghemite (e.g., Dunlop and Özdemir, 1997).

The EDS analyses of grains identified by BSE imaging demonstrate high Ti, Fe, and O peaks (Figure DR2A, see footnote 1). These grains were all submicron scale (~300–500 nm) and made up ~1%–2% of matrix of the sample (Figure DR2B, see footnote 1). The high titanium content is consistent with titanomaghemite as the most likely magnetic remanence carrier mineral in our samples.

Anhyseretic remanent magnetization (ARM) acquisition curves (Figure DR3, see footnote 1) and magnetic hysteresis analyses (Figure DR4, see footnote 1) indicate the presence of a significant fraction of pseudo-single domain, weakly interacting magnetic grains, consistent with the SEM observations.

IRM acquisition values of samples with carbonaceous shale, siltstone, and sandstone lithologies all demonstrate nonsaturation above 100 mT (Fig. 11), suggesting the additional presence of a hard, antiferromagnetic component, such as goethite or hematite. Thermal demagnetization data, which show a drop in magnetic intensity between 100 and 150 °C, also suggest the presence of goethite in most samples.

DISCUSSION

Magnetic Mineralogy

The results of our rock magnetic and SEM analyses demonstrate that most specimens contain weakly interacting, pseudo-single domain titanomaghemite grains. It is likely that the magnetic grains are detrital and were generated during weathering of the source terrain during Laramide uplift (e.g., Butler, 1982; Butler and Lindsay, 1985), or that they are the result of in

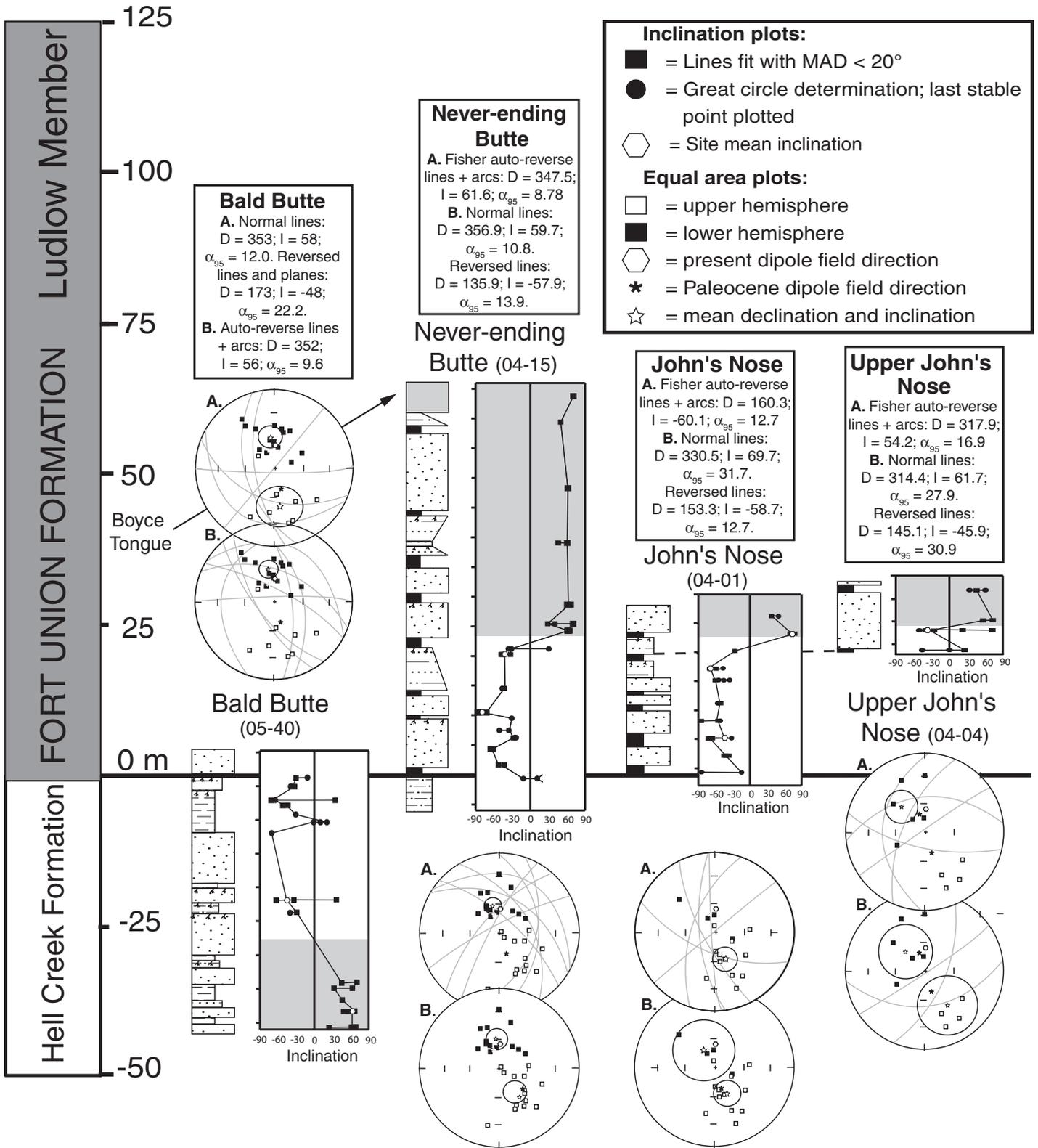


Figure 5. Magnetostratigraphic sections and equal-area plot for Bald Butte, Never-Ending Butte, John's Nose, and Upper John's Nose. Sections are shown in approximate geographic alignment from southwest to northeast. Inclination versus stratigraphic level is plotted for all magnetostratigraphic sections. Interpreted polarity for each section is overlain on inclination plot (gray indicates normal, white is reversed). MAD—mean angle of deviation.

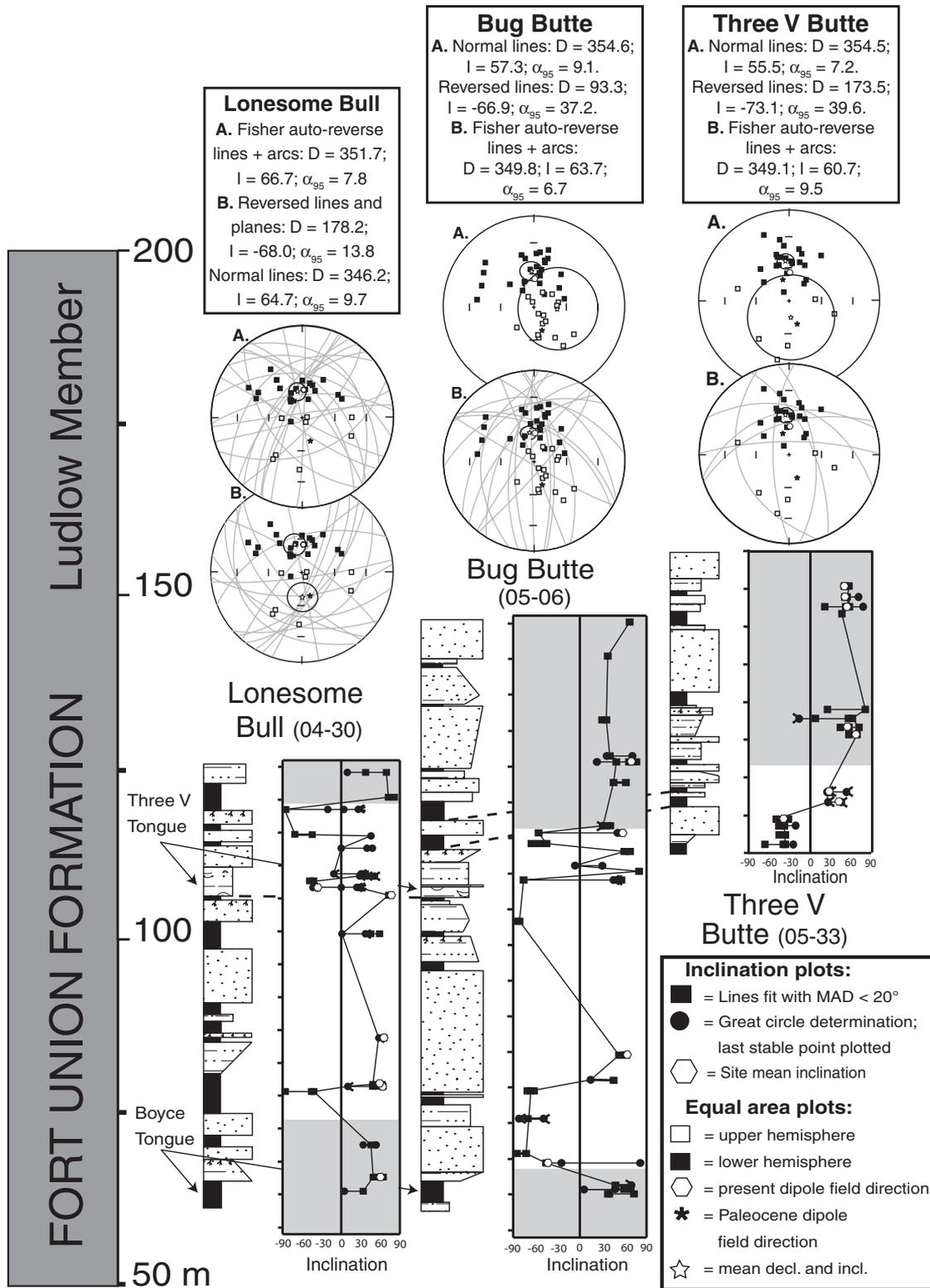


Figure 6. Magnetostratigraphic sections and equal-area plot for Lonesome Bull, Bug Butte, and Three V Butte. Sections are shown in approximate geographic alignment from southwest to northeast. Inclination versus stratigraphic level is plotted for all magnetostratigraphic sections. Interpreted polarity for each section is overlain on inclination plot (gray indicates normal, white is reversed). MAD—mean angle of deviation.

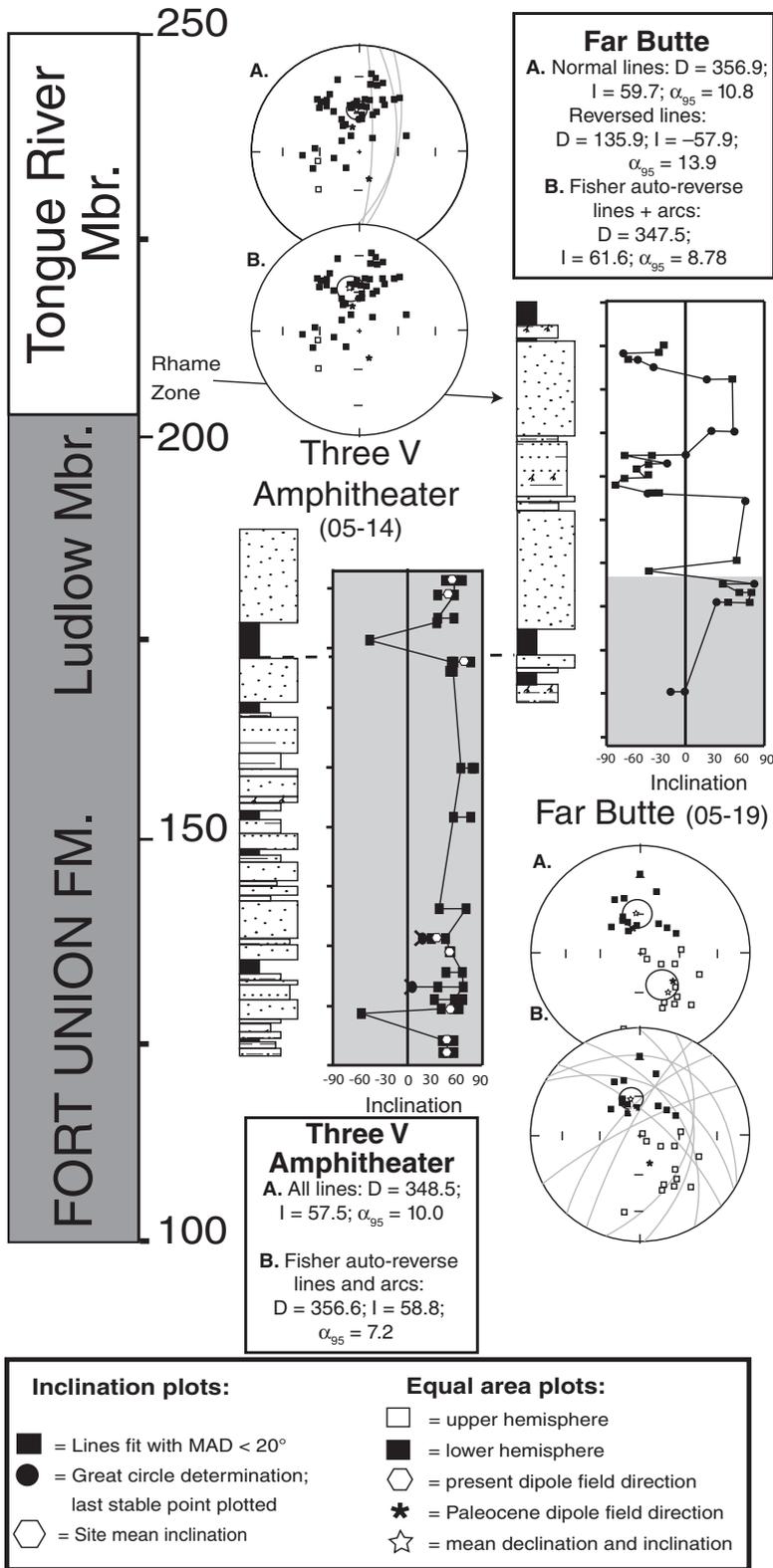
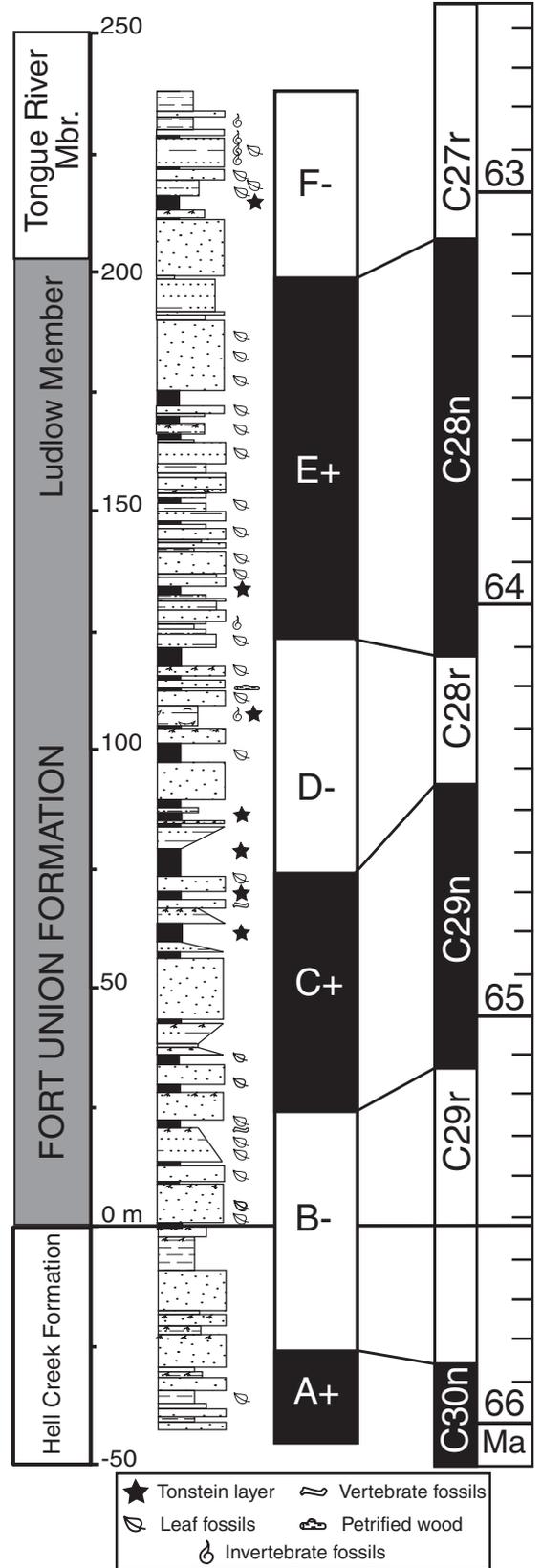


Figure 7. Magnetostratigraphic sections and equal-area plot for Three V Amphitheater and Far Butte. Sections are shown in approximate geographic alignment from southwest to northeast. Inclination versus stratigraphic level is plotted for all magnetostratigraphic sections. Interpreted polarity for each section is overlain on inclination plot (gray indicates normal, white is reversed). MAD—mean angle of deviation.



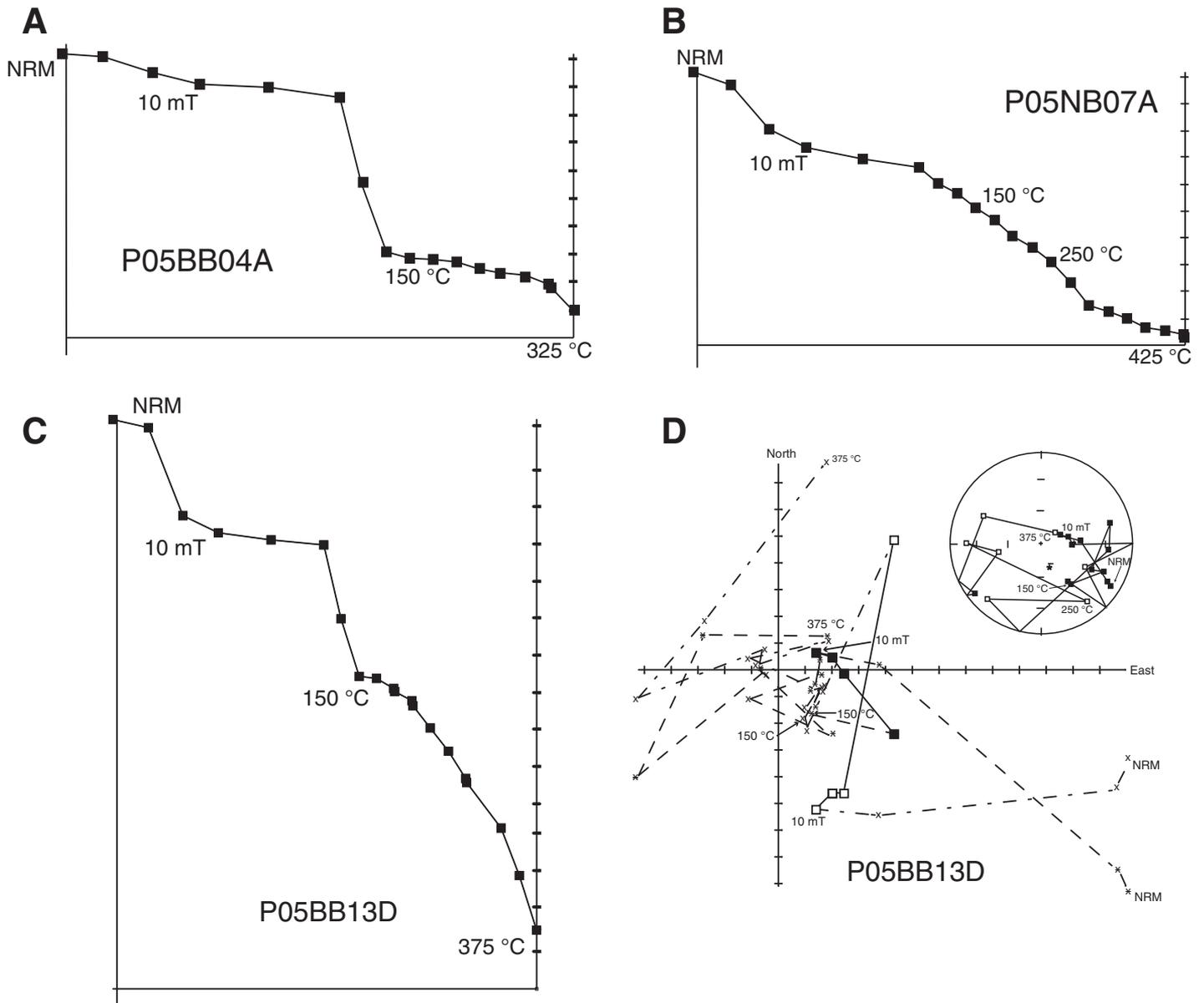


Figure 9. Characteristic J/J_0 plots of paleomagnetic samples. (A and B) Typical J/J_0 plots of reversed polarity samples. (A) Sample P05BB04A shows a large drop in magnetic intensity between 100 and 150 °C and a gradual drop in intensity between 150 and 325 °C. (B) Sample P05NB07A shows a gradual drop in magnetic intensity between natural remanent magnetization (NRM) and 425 °C. Zijderveld diagrams and equal-area plot for this sample are shown in Figure 2A. (C) Typical J/J_0 plot of anomalous normal polarity samples. Sample P05BB13D shows a significant drop in magnetic intensity between 100 and 150 °C, suggesting the presence of goethite. (D) Typical Zijderveld diagram and equal-area plot of anomalous normal polarity samples. Sample P05BB13D shows random demagnetization behavior above 125 °C, which indicates that the dominant detrital magnetic mineral is goethite, and only a normal polarity overprint direction was determined.

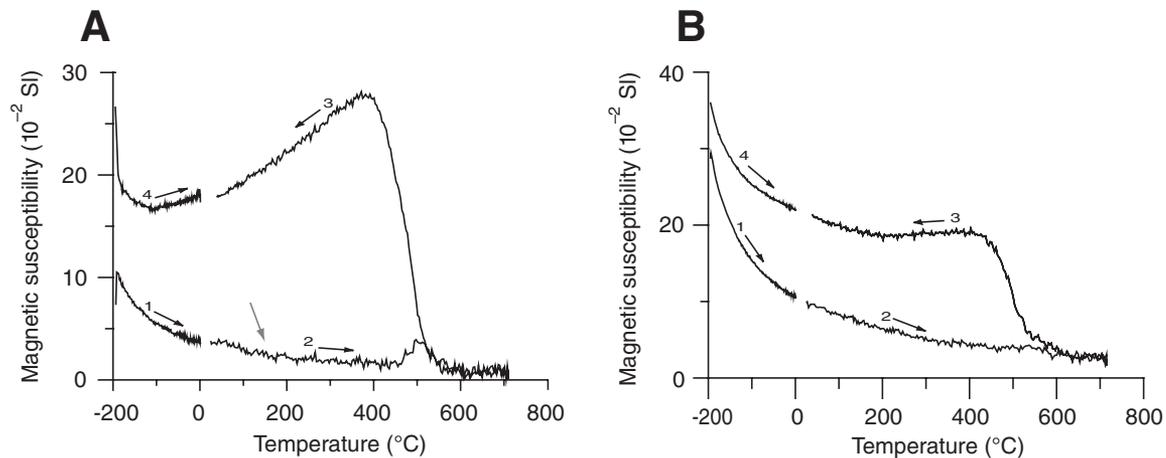


Figure 10. (A and B) Typical dependences of low-field magnetic susceptibility versus temperature for samples P05NB09D (A) and P05NB05D (B). The arrows show the direction of temperature change, and the numbers show the order in which the heating and cooling runs were measured. Both samples manifest irreversible behavior. Sample P05NB09D (A) shows a slight inflection at ~180–200 $^{\circ}\text{C}$ (indicated by gray arrow) and a dominant Curie temperature at ~550 $^{\circ}\text{C}$ (heating curve 2). No clear Curie temperature is defined in heating curves 1 or 2 for sample P05NB05D (B). Both samples manifest a Curie temperature at ~550–580 $^{\circ}\text{C}$ on cooling (curves 3).

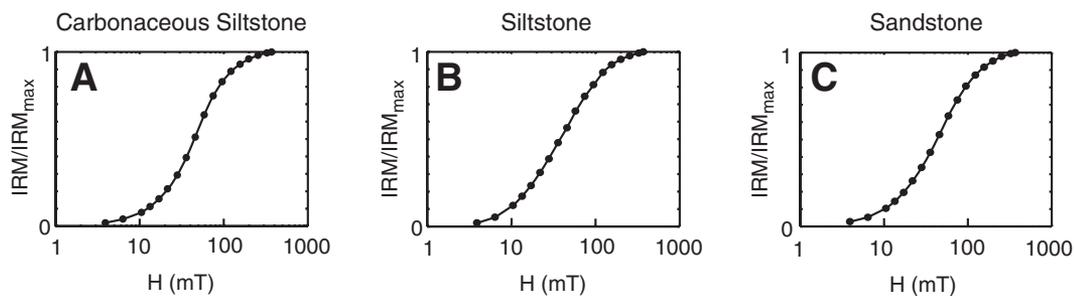


Figure 11. Typical isothermal remanent magnetization (IRM) acquisition curves measured for carbonaceous siltstone (A), siltstone (B), and sandstone (C). The curves are normalized by the maximum IRM value (IRM_{max}). All samples show nonsaturation by a magnetic field exceeding 100 mT.

situ chemical weathering prior to diagenesis or alteration after burial. These results are different than other magnetic mineralogy studies in the Williston Basin (e.g., Lund et al., 2002), suggesting that the provenance for the magnetic grains was different in the Cretaceous and the Paleocene. IRM acquisition data on mudstone, siltstone, and sandstone samples coupled with thermal demagnetization results suggest the additional, minor presence of goethite in most samples. It is likely that the goethite in the samples is the result of secondary precipitation due to water penetration within the sediments. Our combined AF and thermal demagnetization data, which typically show drops in magnetic intensity after a low-AF treatment and between 100 $^{\circ}\text{C}$ and 150 $^{\circ}\text{C}$, indicate that secondary overprints are typically removed early in the demagnetization process.

Relationship of Polarity Stratigraphy to GPTS

The K-T boundary has been demonstrated to be in the upper part of C29r (e.g., Alvarez et al., 1977; Cande and Kent, 1995; Dinares-Turell et al., 2003; Hicks et al., 2002; Lerbekmo and Coulter, 1984; Ogg and Smith, 2004; Swisher et al., 1993). We know the precise position of the K-T boundary within one of our nine sections, its approximate position within another, and the relationship of these two sections to the rest of our magnetostratigraphic sections. Thus, we can relate our polarity stratigraphy to the GPTS. Furthermore, there is an imprecise $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic age of 64.4 ± 1.8 Ma (Warwick et al., 2004) that can be related to the Lonesome Bull and Bug Butte sections, and thus the date is directly tied to our magnetostratigraphic sections. A

comparison of our magnetostratigraphy with the recent version of the GPTS (Ogg and Smith, 2004) indicates that our B- polarity zone corresponds to chron C29r, and subsequent zone/chron correlations are as follows: C+ to C29n, D- to C28r, E+ to C28n, and F- to C27r. All of the early Paleocene geomagnetic polarity chrons are thus represented in our composite section.

Duration of the Ludlow Member

Based on our stratigraphic data, we can calculate an average thickness for each polarity zone within our magnetostratigraphic section relative to the Hell Creek–Fort Union formational contact and the K-T boundary (Table 2). Using these data, plus age estimates for the early Paleocene chrons from seafloor-spreading models (Cande

TABLE 2. AVERAGE CHRON THICKNESS AND AVERAGE STRATIGRAPHIC POSITION OF CHRON BOUNDARY RELATIVE TO THE HELL CREEK-FORT UNION FORMATION CONTACT

Chron	Average chron thickness (m)	Average chron boundary stratigraphic position relative to formation contact (m)
C29r (Cretaceous)	26	-26
C29r (Paleocene)	24	24
C29n	50	74
C28r	50	124
C28n	72	196

and Kent, 1995; Ogg and Smith, 2004) and precessional cyclicity calculations (Dinares-Turell et al., 2003; Preissinger et al., 2002; Westerhold et al., 2008), we can calculate the duration of deposition for the Ludlow Member (Table DR5, see footnote 1). The primary assumption in the calculation is that there are no significant sedimentary unconformities in our sections. This assumption seems reasonable, based on our detailed stratigraphic logging and the ability to trace semicontinuous exposures between measured sections. It is also supported by consistent sediment accumulation rates for each polarity chron as estimated from the GPTS (Fig. 12).

Our primary calibration points are the K-T boundary, which was recalibrated by Hicks et al. (2002) to 65.51 Ma, and the top of C28n, which ranges from 63.31 Ma (Preissinger et al., 2002; Westerhold et al., 2008) to 63.02 (Cande and Kent, 1995). The duration of the Ludlow Member was then calculated by extrapolating from the two reference data upward to the Ludlow-Tongue River contact.

The calculated sedimentation rates for the Ludlow Member range from 79 to 89 m/m.y. Using these sedimentation rates, and a total thickness of the Ludlow Member of 210 m, its calculated duration is estimated between 2.31 m.y. and 2.61 m.y. Thus, the minimum age for the top of the Ludlow Member is 62.90 Ma, and the maximum age is 63.20 Ma.

The marine tongues of the Cannonball Member correlate as follows: the Boyce Tongue within C29n and the Three V Tongue within C28r. Their range of age estimates are interpreted as 64.67–64.58 Ma for the Boyce Tongue and 64.22–64.09 Ma for the Three V Tongue. This suggests that deposition of the Ludlow Member sediments prior to ca. 64.50 Ma was the result of, or significantly influenced by, rising base level due to the transgression of the Cannonball Sea. Furthermore, it indicates that by ca. 64 Ma, the seaway had fully regressed from southwestern North Dakota. These results

suggest that the sedimentological differences between the lower and upper Ludlow Member are likely related to a change in the environments of deposition due to the regression of the Cannonball Seaway, which may also explain the increase in large sandstone bodies and thick lignite deposits in the upper Ludlow Member.

These new paleomagnetic data for the Ludlow Member, combined with published mammal biostratigraphic studies from the Pita Flats locality and the Brown Ranch localities (Hunter, 1999; Hunter and Hartman, 2003; Hunter et al., 2003), document the occurrence of Puercan 2 and Puercan 3 mammals within C29r and Torrejonian mammals within C29n. These mammalian occurrences are ~500 k.y. to 1 m.y. earlier in the Williston Basin than in the type areas for the Puercan and Torrejonian in the San Juan Basin. These occurrences suggest that either the Puercan and Torrejonian North American Land Mammal Ages are compressed in the Williston Basin, the temporal ranges of the Williston Basin mammals are different and/or need to be extended, or there are biogeographic differences between the Puercan and Torrejonian faunas across the Western Interior of the United States. Each of these hypotheses suggests that the post-K-T speciation of mammals may have occurred earlier and/or more rapidly than previously supposed. Further detailed study of the mammalian faunas in the Williston Basin needs to be undertaken to address these possibilities.

CONCLUSIONS

Paleomagnetic and rock magnetic analyses of the Ludlow Member of the Fort Union Formation indicate that the dominant detrital magnetic mineral is titanomaghemite with minor secondary goethite. We conclude that the characteristic magnetization is primary based on a calculated mean polarity direction consistent with expected directions for the Paleocene of North America, and a reversal stratigraphy that corresponds to the GPTS with reasonable sedimentation rates. Magnetostratigraphic analysis of nine sections demonstrates that the Ludlow Member can be correlated to C29r–C27r of the GPTS.

Using these magnetostratigraphic data and age estimates for the duration of the magnetic polarity chrons (Cande and Kent, 1995; Dinares-Turell et al., 2003; Ogg and Smith, 2004; Preissinger et al., 2002; Westerhold et al., 2008), we have, for the first time, made estimates of the duration and the age of the Ludlow Member. Based on our calculated sedimentation rates for the member, the duration is at minimum 2.31 m.y. and at maximum 2.61 m.y. Using detailed lithostratigraphy and magnetostratigraphic cor-

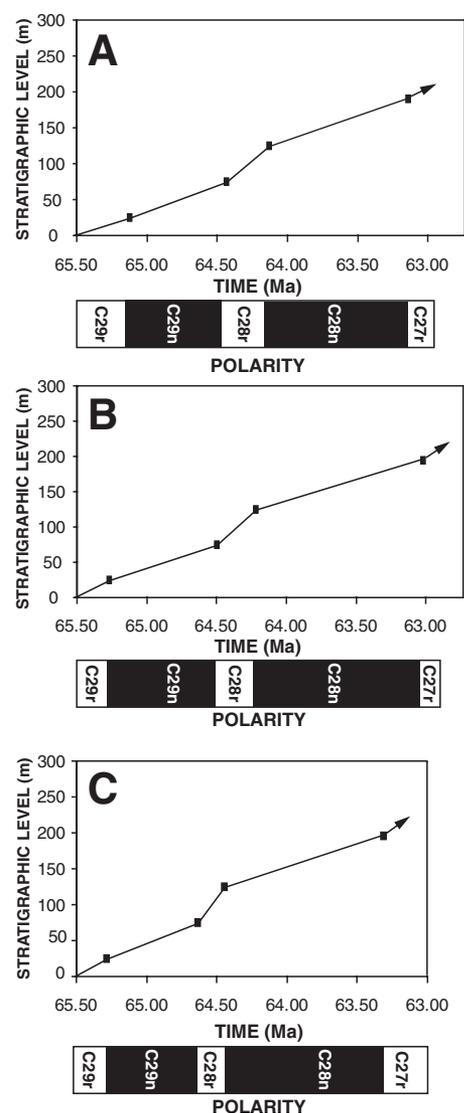


Figure 12. Sedimentation rates of the Ludlow Member of the Fort Union Formation. Slight change in slope at the base and top of C28r most likely represents the transition from predominately terrestrial to marginal marine sedimentation then back to terrestrial sedimentation. All data for plots are given in Table DR5 (see text footnote 1). (A) Ogg and Smith (2004). (B) Cande and Kent (1995). (C) Preissinger et al. (2002).

relation to the GPTS, we have shown that it is highly unlikely that substantial unconformities are present in the Ludlow Member. Therefore, the interpretations presented here constitute the best age estimates for examinations of rates of biotic recovery subsequent to the Cretaceous-Tertiary extinction in the Williston Basin. Further study of the Ludlow Member should focus

on isotopic dating of the numerous ash beds that we have identified in our measured sections.

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