Chapter 3

Long-eccentricity regulated upstream and downstream controls on fluvial incision and aggradation in the Paleocene of north-eastern Montana (USA)

ABSTRACT

Aggradation and fluvial incision controlled by downstream base-level changes at timescales of 10 to 500 kyr is incorporated in classic sequence stratigraphic models. However, upstream climate control on sediment supply and discharge variability causes fluvial incision and aggradation as well. Orbital forcing often regulates climate change at 10 to 500 kyr timescales while tectonic processes such as flexural (un)loading exert a dominant control at timescales longer than 500 kyr. It remains challenging to attribute fluvial incision and aggradation to up- or downstream processes or disentangle allogenic from autogenic forcing, as time control is mostly limited in fluvial successions.

The Paleocene outcrops of the fluvial Lebo Shale Member in north-eastern Montana (Williston Basin, USA) form an exception. Here, we use a distinctive tephra layer and two geomagnetic polarity reversals to create a 15-km long chronostratigraphic framework based on the correlation of 12 sections. Three aggradation-incision sequences are identified with durations of approximately 400 kyr, suggesting a relation with long-eccentricity. Our age control further reveals that incision occurred in the direction of -or during- a 405-kyr long-eccentricity minimum. A long-term relaxation of the hydrological cycle related to such an orbital phasing potentially exerts a downstream as well as an upstream control on river incision. Downstream, base-level rise can be expected following c. 30 m of global sea-level rise predicted by aquifer-eustasy. The resulting flattening longitudinal profile can be accompanied with incision upstream of the terrace intersection. Upstream, an expanding vegetation cover is expected because of an increasingly constant moisture supply to source areas. Because of entrapping by vegetation, sediment supply was significantly reduced relative to discharge, especially at times of low evapotranspiration. Hence, high discharges resulted in incision. We assess the long-eccentricity regulated upstream and downstream controls on fluvial aggradation and incision in a new aggradation-incision sequence model.
INTRODUCTION

Many fluvial successions are built by aggradation intermittently interrupted by degradation (incision), operating at 10 kyr - 10 Myr time scales. Aggradation occurs when sediment supply is significantly higher than maximum bedload transport rate, and incision occurs when sediment supply is significantly lower than bedload transport rate. Both aggradation and incision are controlled by internal (autogenic) and/or external (allogenic) factors on sediment supply (Mackin, 1948; Leopold & Bull, 1979; Blum & Törnqvist, 2000; Holbrook et al., 2006; Catuneanu, 2006; Miall, 2014). On 10 kyr - 10 Myr time scales, fluvial incision and aggradation are attributed to an interplay of allogenic controls: geomorphic base-level (i.e. sea-level, lake-level, or a drainage network trunk channel axis), climate, and tectonics (e.g. Schumm, 1993; Shanley & McCabe, 1994; Ethridge et al., 1998; Miall, 2014). Base-level change remains a major cause for fluvial incision and aggradation in sequence stratigraphic models (e.g. Posamentier & Vail, 1988; Wright & Marriott, 1993; Shanley & McCabe, 1994). Accordingly base-level fall causes incision and base-level rise aggradation. Nevertheless, the sequence stratigraphic model cannot simply be applied to every fluvial setting. If, and how, base-level change causes aggradation or incision depends in particular on the difference between the coastal plain and shelf gradient, the river’s ability to self-accommodate by adapting channel sinuosity, and the landward limit of base-level influence (Schumm, 1993; Blum & Törnqvist, 2000). Base-level change may be controlled downstream by eustasy and upstream by sediment supply. Greenhouse worlds feature only minor or no polar ice-caps which makes glacio-eustasy as a mechanism for base-level control unlikely. In contrast, aquifer-eustasy driven by eccentricity-forced climate changes may control higher-order global sea-level fluctuations of ca 30 m (Wendler & Wendler, 2016; Sames et al., 2016; Wendler et al., 2016). In addition to downstream control by aquifer-eustasy, orbital-forced climate changes in the hinterland (i.e. precipitation, weathering, vegetation), that influence sediment supply and discharge, exert an important upstream control on sedimentation at the basin-scale (Fielding & Webb, 1996; Abels et al., 2013; Noorbergen et al., 2018).

Orbital control is difficult to separate from time-overlapping tectonic forcing (e.g. Holbrook & Schumm, 1999) and autogenic processes (e.g. Hajek & Straub, 2017). This is mainly due to poor age control in pre-Quaternary archives (Blum & Törnqvist, 2000; Abels et al., 2013). It precludes proper assessment of timescales, lateral extents of facies, and inter-facies connections. Not surprisingly, only a few pre-Quaternary studies have documented impacts of orbital-forced climate changes on sedimentation in fluvial systems (Olsen, 1990; Olsen et al., 1994; Fielding & Webb, 1996; Abels et al., 2013; Noorbergen et al., 2018). These studies deal with fluvial stratigraphic architectures that are dominated by aggradation and lack valley-related unconformable surfaces. It thus remains unknown how orbital-forced climate change may influence aggradation and incision in pre-Quaternary greenhouse fluvial systems.

The Paleocene Lebo Shale Member of the Fort Union Formation, Williston Basin (north-eastern Montana, USA), provides the rare opportunity to investigate the control mechanisms related to fluvial aggradation and incision under greenhouse climate conditions and at the time scales of interest. For this purpose, a ca 15-km-long N-S oriented stratigraphic fence panel, based on detailed stratigraphic logging of parallel sections in combination with magneto- and tephr stratigraphic correlations and tracing of stratigraphic marker levels, such as coals and palaeosols, has been generated in McCone County (north-eastern Montana, USA).
GEOLOGICAL SETTING

The lower Fort Union Formation in the Williston Basin

During Cretaceous time, the intracratonic Williston Basin was encapsulated in the Western Interior Foreland Basin (DeCelles, 2004). Depositional environments in the Williston Basin were

Figure 3.1. Geographic and geological setting of the study area in McCone County, north-eastern Montana (USA). A: Environmental reconstruction of Montana and adjacent states (modified from Flores, 2003). Peat mires were dissected by low sediment load rivers that drained into the Cannonball Sea in the early Paleocene, approximately 65 Ma. Dashed white lines show map locations projected on the lithostratigraphic diagram of panel B. B: Along-profile cross-sectional diagram of lithostratigraphic units comprising Chron 29r to Chron 28n (modified from Johnson et al., 2002). Study interval of Lebo Shale Member in north-eastern Montana is indicated (red arrow), stratigraphically above the Tullock Member studied by Noorbergen et al. (2018; white arrow). C: Enlarged area from Panel 1D. Figure shows lenticular ridges belonging to the muddy point-bar deposits of Facies B1 (Table 3.1). D: Ortho-image of the study area with 24.4-m (80-feet) elevation contour lines. Facies B1 valley-fill deposits are coloured orange for aggradation-incision sequence 1 (AIS-1) and cyan for AIS-2. Thick black solid line shows field mapping of coal bed U by Collier & Knechtel (1939). The locations of the sections (yellow circles) are connected by a solid black line showing the transect of the fence panel (Fig. 3.8). Section abbreviations (from north to south): CMD-W = Coal Mine Divide West, CMD-M = Coal Mine Divide Main, RP-N = Rough Prong North, RP-E = Rough Prong East, RP-M = Rough Prong Main, RP-SW = Rough Prong Southwest, RT-NE = Radiotower Northeast, RT-E = Radiotower East, RT-M = Radiotower Main, RT-SE = Radiotower Southeast, HCR-N = Horse Creek Road North, HCR-M = Horse Creek Road Main. Orthoimagery, quadrangle and road data were obtained from the Montana Spatial Data Infrastructure (MSDI) (http://geoinfo.msl.mt.gov/Home/msdi).
mainly marine in the mid-Cretaceous but these became replaced by the alluvial systems of the Laramide orogeny taking place from the late Cretaceous into the Eocene (Cherven & Jacob, 1985). Approximately half-way the Laramide uplift, in the early Paleocene, fluvo-deltaic sediments of the Fort Union Formation were deposited. In the Williston Basin, the lower Paleocene part of the Fort Union Formation is exposed in the badlands of the Missouri River and its tributaries, broadly covering the region between eastern-Montana and central-Dakota (Fig. 3.1A). The base of the Fort Union Formation is represented by the first laterally extensive coal seams at -or close to- the K-Pg boundary (Fastovsky & Bercovici, 2016). The top of the lower Fort Union Formation, at the base of the Tongue River Member, is represented by a few meters thick bleached palaeo-weathering zone (i.e. the Rhame Bed) that might represent a hiatus of a few million years of middle Paleocene time (Warwick et al., 2004). The lower Fort Union Formation has been subdivided into five members that are partly lateral equivalents; the Tullock, Lebo Shale, Ludlow, Slope and Cannonball Members (stratigraphic relations, Fig. 3.1B).

The Lebo Shale Member
In the Williston Basin, the Lebo Shale Member of the lower Fort Union Formation consists of greyish-yellow sand-dominated intervals and grey mud-dominated intervals, both containing several coal seams and bleached zones (Collier & Knechtel, 1939; Rigby & Rigby Jr., 1990). The sand-dominated intervals are laterally extensive and form erosion-resistant caps on top of buttes and ridges. These have been interpreted as channel-splay systems (Rigby & Rigby Jr., 1990). The mud-dominated intervals contain coaly organic material and locally contain lenses of coal and carbonaceous shale. If enriched in shale, these units typically show a sombre and bentonitic (as ‘dirty-popcorn’) surface weathering. They have been deposited in moderately saline playas with low carbonate concentrations (Rigby & Rigby Jr., 1990). The coals are subbituminous to lignite in rank (Collier & Knechtel, 1939; Rigby & Rigby Jr., 1990) and contain abundant vitrain alternated by dull-black powdery material (Rigby & Rigby Jr., 1990). Coals in the Lebo Shale Member may have been formed in raised mires (Flores & Keighin, 1999; Flores et al., 1999). Bleached zones are whitish laterally extensive markers and have been interpreted as the leaching-horizons of palaeosols (Rigby & Rigby Jr., 1990).

The presence of dark-grey shales is the lithological criterion to distinguish the Lebo Shale Member from the underlying grey-gold-brownish thinly banded siltstones of the Tullock Member. Accordingly, the contact between the two members is placed at the base of the first distinct dark-grey shales just above the W-coal zone (Noorbergen et al., 2018). In McConne County, distinct coal beds in the Fort Union Formation have been locally traced and regionally mapped based on topographic extrapolations (Collier & Knechtel, 1939). In their mapping, Collier and Knechtel (1939) assigned these coal beds reversed alphabetic labels, from Z to P, with respect to their inferred regional stratigraphic extrapolations. Given the Tullock-Lebo lithostratigraphic contact as defined in Noorbergen et al. (2018), the Lebo Shale Member in McConne County includes coal beds V, U, and T of Collier & Knechtel (1939). In the study area (Fig. 3.1D), bed U has been mapped by Collier & Knechtel (1939), while bed V was not indicated and bed T was only indicated at a few locations. Because each coal bed of Collier & Knechtel (1939) is generally part of a larger cluster consisting of multiple coals, they are referred to as zones in this work. Individual coal beds can be up to 2 m thick but still contain cm-scale detrital partings and cm- to dm-scale intervals with volcanic ash or sharp mm to cm-scale ash intercalations (i.e. tephras). The tephras contain euhedral crystals which enables their distinction from detrital partings that contain rounded minerals as the result of abrasion. The U-coal zone, containing abundant tephras (> 20) and detrital partings, has been
correlated to the Big Dirty coal zone (Collier & Knechtel, 1939), originally defined in the Bull Mountain coal field (Woolsey et al., 1917), ca 300 km southwest of McCone County.

METHODS

Sections and palaeo-flow measurements
In McCone County, twelve sections have been logged in four isolated outcrop areas that are from north to south: Coal Mine Divide (CMD), Rough Prong (RP), Radiotower (RT), and Horse Creek Road (HCR). There are at least two sections in each outcrop area; a central main section (last character is the M, e.g. CMD-M) and one or more other sections at a maximum distance of ca 3 km from the main section (last character is the compass direction with respect to main section, e.g. CMD-W is west of the main section at Coal Mine Divide).

Palaeo-flow directions were measured from the dip planes of different sets of decimetre-scale cross-bedded sandstone that were interpreted as mid-channel bars. Palaeo-flow orientations were also determined perpendicular to the dip plane of lateral accretion planes that were interpreted as point-bars. Both sources of palaeo-flow data were plotted onto separate rose diagrams. No corrections for tectonic dip were required because the layers of the Fort Union Formation in north eastern Montana are dipping less than 1˚.

Palaeomagnetism
All samples for palaeomagnetic analysis were taken in trenches, from unweathered rock at least one meter below the surface. A total of 243 levels were sampled within eight sections. Standard core samples (diameter = 2.54 cm) were taken with an electric, battery-powered drill using water as a coolant and oriented with a Brunton compass fixed to a Pomeroy orientation shaft. For each level it was attempted to drill multiple cores or one core of sufficient length so that the sample could be split into at least two specimens, enabling both thermal (TH) and alternating field (AF) demagnetization.

Previous rock magnetic analyses in the lower Fort Union Formation showed that the dominant magnetic remanence carriers are magnetite and maghemite (Swisher III et al., 1993; Sprain et al., 2016). Intermediate-composition titanohematite could also be present and, if abundant, this mineral might complicate palaeomagnetic interpretations because of its ability to self-reverse (Sprain et al., 2016). Previous magnetostratigraphic results of the lower Fort Union Formation in this area (Swisher III et al., 1993; LeCain et al., 2014; Noorbergen et al., 2018; Sprain et al., 2018) are mutually consistent and in line with the Geomagnetic Polarity Timescale (Ogg, 2012). Therefore, self-reversal of intermediate titanohematite is a minor issue. In this study, TH demagnetization was done on 219 samples and AF demagnetization on 205. Of the total of 243 sampling horizons, 172 could be processed with both TH and AF demagnetization. More detailed procedures for the TH and AF demagnetization experiments are described in Noorbergen et al. (2018).

Inclination and declination components of the Characteristic Remanent Magnetization (ChRM) were determined using the Remasoft 3.2 software program (Chadima & Hrouda, 2006). Plots were exported from Paleomagnetism.org (Koymans et al., 2016). ChRM directions were determined by anchored principal component analysis (PCA) (Kirschvink, 1980) if the sample clearly trended towards the origin along at least four consecutive demagnetization levels. Fisher mean statistics (Fisher, 1953) was used if a sample showed no clear trend towards the origin, but a clustering of higher coercivity vector end-points. Data points with a mean angular deviation (MAD) < 15˚ of the
anchored fit or Fisher mean are connected (Fig. 3.3). Samples with a ChRM, but with a MAD > 15˚, are displayed with open symbols.

**Scaling, vertical optimization and correlation of sections**

From north (left) to south (right), along the transect (Fig. 3.1), stratigraphic logs of the sections were plotted on a 1 : 80 000 horizontal scale and a 1 : 600 vertical scale (Fig. 3.8). Two geomagnetic polarity reversals and one distinctive tephra ("the Sugar Ash") were used to optimize their vertical positions with respect to a backbone section (Coal Mine Divide West, CMD-W) in which all three chronostratigraphic markers are present. For the polarity reversals the mid-point of the reversal interval was used as absolute value (Figs 3 and 8). From the five sections that contained all three markers, CMD-W was selected as backbone, because the stratigraphic spacing of the markers in this section best represented the entire study area. The position of polarity reversals in sections lacking palaeomagnetic data was estimated by lithostratigraphic correlations to nearby sections that recorded the reversal to increase the number of optimization tie-points. Prediction of polarity reversals include the lower reversal in Rough Prong North (RP-N), Rough Prong Main (RP-M), and Radiotower Northeast (RT-NE) and both the lower and upper reversals in Rough Prong Southwest (RP-SW), Radiotower East (RT-E), and Horse Creek Road North (HCR-N). For the vertical optimization, the same approach was used as in Noorbergen et al. (2018). After scaling and vertical optimization, sections were correlated, using the three chronostratigraphic markers. On the basis of this chronostratigraphic fence panel, the stratigraphic architecture of the three facies associations (Table 3.1) was interpreted.

**RESULTS**

**Magnetostratigraphy**

Typical demagnetization results of the Natural Remanent Magnetization (NRM) are presented in Figure 3.2. The initial NRM intensities range from ca 500 to ca 1500 x 10^-6 A/m. The majority of the samples are demagnetized at ca 400 °C or 100 mT, after which they generally have a remaining NRM intensity < 150 x 10^-6 A/m. Palaeomagnetic directions have been determined from unblocking of the ChRM between 120 and 400 °C for TH and between 10 and 100 mT for AF demagnetizations. The polarities can be divided into four groups: (1) reversed polarity (24.3 %), (2) normal polarity (56.1 %), (3) uncertain polarity (14.6 %) and (4) undetermined polarity (5 %). A stable primary ChRM component can be isolated in most reversed polarity samples (Fig. 3.2). In normal polarity samples, the difference between the present-day overprint and the ChRM is difficult to discern. On the reasonable premise of similar unblocking behaviour of reversed and normal polarity samples, the 120 to 400 °C or 10 to 100 mT range was also used to isolate the ChRM of normal polarity samples (Fig. 3.2). Samples with an uncertain polarity show inclination and declination angles in between normal and reversed samples and/or clustering of vector end-points straddling the origin. The undetermined polarity group includes all samples without a reliable ChRM component.

In total 47.6 % of the acquired directions show a clear ChRM component trending towards the origin. These components are calculated with an anchored-PCA fit. The remaining directions (52.4 %) show a clustering of higher coercivity vector end-points and are calculated with a Fisher mean. Directions calculated from demagnetization of samples derived from bleached palaeosols are also calculated with a Fisher mean (Fig. 3.2). These samples showed directions that significantly deviate from trends in samples that were unaltered by soil leaching (Fig. 3.3) probably due to the eluviation
Figure 3.2. Zijderveld diagrams of TH and AF demagnetization of characteristic samples underpinning the magnetostratigraphy of the Lebo Shale Member in the study area. Left panels show examples of normal polarity (C29n or C28n) samples, middle panels show examples of reversed polarity (C28r) samples, and right panels show examples of polarity data measured from bleached palaeosols. In samples with a clear reversed polarity, stable ChRM components of inclination (inc) (open circles) and declination (dec) (solid circles) were isolated from the secondary present-day overprint component upward from 120-240 °C (TH) and 10-28 mT (AF). Points included in the anchored-PCA fit are indicated; declination (blue) and inclination (green). Equal area plots have been included to the panels on the right, since directions of these samples were calculated with a Fisher mean. Points included in the Fisher mean are indicated in red.
Legend lithostratigraphic columns
characteristic Munsell-colours lithologies:
- N 1.5/0 (coaly) black
- 2.5Y 7/4 (sandy) dull orange yellow
- 7.5YR 8/1 (bleached) pale beige
- 5Y 3/1 (muddy) grayish olive
sedimentary and stratigraphic symbols:
- undulating surface
- cross-bedded
- wavy bedded
- heterolithic strata
- tephra layer
- tephra correlation

Legend magnetostratigraphic columns
- inc° TH demag. (MAD<15)
- dec° TH demag. (MAD<15)
- dec° TH / inc° AF demag. (MAD>15)
- TH / AF demag. duplicate site level
- TH / AF demag. unreliable polarity
- TH / AF demag. whitish palaeosol
- Sugar Ash (SA): tie-point Fig. 3.8
- mid-height reversal: tie-point Fig. 3.8
Fig. 3.38 Rough Prong Main inclination

Rough Prong East (RP-E)

field trace

Rough Prong Southwest (RP-SW)

Rough Prong North (RP-N)
of the primary magnetic clay minerals. TH and AF magnetostratigraphic data are provided in Tables S1 and S2. Raw TH and AF palaeomagnetic data are available online in the data repository of this article.
The resultant magnetostratigraphy reveals a single reversed polarity interval with normal polarities stratigraphically both below and above. The lower normal polarities correspond to Chron C29n, as they represent the stratigraphic continuation of normal polarities recorded in the top part of the underlying Tullock Member that have previously been correlated to C29n by Noorbergen et al. (2018) in accordance with the 40Ar/39Ar radioisotope dating of Swisher III et al. (1993) and Sprain et al. (2015). The reversed interval thus corresponds to C28r and the two recorded reversals to the C29n/C28r and C28r/C28n Chron boundaries.

Figure 3.3. Sedimentological logs and magnetostratigraphic columns of the sections. A: Coal Mine Divide. B: Rough Prong. C: Radiotower part 1. D: Radiotower part 2. E: Horse Creek Road.
Facies associations

Three facies associations are recognized in the Lebo Shale Member: (A) channel-splay, (B) valley-related, and (C) peat mire facies association. A description of the different facies within the facies associations along with their depositional interpretation is given below and summarized in Table 3.1. Figure 3.4 provides a field photo of an outcrop in which the three Facies Associations A, B, and C are identified. Facies-specific field photographs of the facies associations are provided in Figures 3.5, 3.6, and 3.7, respectively.

Figure 3.4. A: View to the NW from the RP-M section showing the Lebo Shale Member in between the RP-N and RP-E sections respectively ca 1 km left and ca 500 m right outside the photo (section locations in Fig. 3.1D). White line marks area interpreted in B. B: Sedimentary facies belonging to the three Facies Associations A, B and C identified for the Lebo Shale Member in north-eastern Montana (Table 3.1).

Facies Association A – Channel-splay

Description. Facies Association A (Fig. 3.5) is composed of Facies A1 (ca 20 %), A2 (ca 10 %) and A3 (ca 70 %). Facies A1 consists of up to 10-m thick light-yellow-weathered sandstone bodies that are up to 0.25-km wide in cross-sectional dimension. From aerial view they show nearly straight ridges (Fig. 3.1C). The bodies consist of up to m-scale trough-cross-stratified medium-grained...
sandstone sheets (Figs 5C and 5D) gradually changing-upward to cm-scale trough cross-laminated and climbing ripple cross-laminated fine- to very-fine-grained sandstone.

Facies A2 is represented by ca 1 – 15 m wide and ca 1 – 3 m thick, locally isolated, brownish-yellow-weathered concentric sandstone bodies with average total width-depth ratios of 4.8 (σ=1.4, n=3; Fig. 3.5B). The bodies consist of 10 – 30 cm thick concentric internal bedding composed of slightly muddy sandstone in the lower part and very muddy sandstone in the upper part (Fig. 3.5B).
Facies A3 is composed of ca 0.5-m to 5-m thick brownish-yellow weathered, horizontally stratified packages consisting of dm- to cm-scale silty sandstone (ca 80 %) to mudstone (ca 20 %) sequences that show fining-upward (Fig. 3.5A). The silty sandstones are dm- to cm-scale rippled (Fig. 3.5E) and horizontally laminated. The mudstones contain dm- to cm-scale horizontal laminations. The sand-mudstone sequences show km-scale lateral extent. Locally mudstones can be up to 1-m thick. Within the horizontally stratified packages, up to 10-m wide lenticular intercalations of up to 2-m thick fine-grained sandstone bodies occur (Fig. 3.5A).

Dm-scale cross-stratified sandstone units of Facies A1 are found with a sharp contact on top of underlying facies. Laterally, sands of Facies A1 pass gradually into the laminated silty sandstones and mudstones of Facies A3. The concentrically bedded local sandstone deposits of Facies A2 have a sharp lower contact. Laterally, sediments in the upper part of Facies A2 pass gradually into those of Facies A3 (Fig. 3.5B).

**Interpretation.** The sharp bases, relatively coarse grain sizes compared to A3 and channelized geometries lacking major concave truncation-surfaces, point toward slightly erosive-based aggrading channels for the origin of Facies A1 and A2. Channel deposits of Facies A1 and A2 that laterally pass into the finer-grained deposits of Facies A3 indicate a crevasse-splay origin for the latter. The elongated cross-sectional nature, dm-scale cross-bedded sheets, and low dispersive palaeo-flow (Fig. 3.1C) of Facies A1 are indicative of large-sized, low-sinuosity, sandy river channels (Rigby & Rigby Jr., 1990). The sedimentation of sands in such channels possibly led to super-elevation with respect to Facies A3, initiating subsequent river avulsion and compensational stacking (e.g. Straub et al., 2009; Hajek & Straub, 2017).

The symmetric cross-sectional nature, isolated occurrence, and concentric internal bedding of Facies A2 argue for the presence of small-sized, relatively narrow channels with the filling composed of multiple individual stories (Gibling, 2006). Such a filling style may be comparable to modern ephemeral rivers in central Australia (Gibling et al., 1998).

The grain-size range, sedimentary structures, and organic content of the horizontally stratified packages of Facies A3 suggest the presence of most crevasse-splay elements: i.e. crevasse-channel (lenticular sandstone intercalations), proximal to distal splays (respectively dm- to cm-scale fining-up sequences), and very distal splays (slightly laminated to structureless mudstones). The km-scale lateral extent of the sand-prone (ca 80 %) crevasse splay deposits indicate mature crevasse-splay systems. The high sand (relative to mud) fraction may be due to relatively low-accommodation, mainly offered by compaction of underlying peats and/or high amounts of sand in suspension during flood events (Burns et al., 2017).

**Facies Association B – Valley-related**

**Description.** Facies Association B (Fig. 3.6) is composed of Facies B1 (ca 50 %), B2 (ca 40 %), and B3 (ca 10 %). Facies B1 consists of up to 20-m thick, light-grey- to grey-weathered heterolithic bodies (Figs 6A and 6B) that are maximal 1-km wide in cross-sectional dimension and show multiple WNW-ESE oriented lenticular ridges from aerial view (Fig. 3.1C). The bodies consist of an up to 10-m thick, dm-scale cross-stratified, medium- to fine-grained sandstone-dominated lower part (Fig. 3.6D) gradually changing-upward into a ca 10-m thick heterolithic upper part. The upper part is composed of low-angle inclined- to horizontal dm-scale alternations of ca < 40 % fine-grained sandstone and ca > 60 % muddy sandstone (Fig. 3.6C). Within the dm-scale inclined sandstone-mudstone alternations, a few shallow truncation surfaces have been observed, laterally extending up to 200 m (Figs 4 and 6A). These contacts mark a slight < 2˚ change in dip directions from
underlying to overlying bedding. Up to m-scale coal rafts and petrified tree trunks (Fig. 3.6F) occur within the bodies of Facies B1.

Facies B2 consists of up to 6-m thick mud-dominated units that are composed of dm-scale horizontal alternations between poorly structured, massive sandy claystone and mm- to cm-scale, vaguely ripple laminated, muddy-sandstone layers (Fig. 3.6E). Cm- to dm-scale pieces of petrified wood are generally aligned with the sand- and mudstone bed surfaces of Facies B2 (Figs 7B, 7C and 7D).

Facies B3 is characterized by up to 3-m thick white- to light-grey-weathered (bleached) heterolithic (clastic to coal) marker zones as these can be traced in the field over distances of several kilometres (Fig 6H). When siliciclastic-prone, the horizons occur in the top parts of Facies Association A. When coaly, the horizons occur in the top of Facies Association C. Sandstone-prone horizons show white weathering, mudstone-prone horizons show light-grey weathering, and coal-prone horizons show bright-black weathering. Randomly oriented root traces, sometimes coalified, are common in the sand- and mudstones. Many slickensides can be found in the mudstones. The sand- and mud-prone marker zones contain one or multiple gold (2.5Y 7/8) and/or dark purplish grey (5P 4/1) mottled (up to 50 %) zones and internally cm- to dm-scale concretions may be present (Fig. 3.6I). Locally, just below the undulating top of Facies B3, an up to 20-cm thick, dark (10YR 1.7/1) sandy, or muddy horizon occurs with cm-scale coaly patches, burrows, and/or ash lenses (Figs 6E and 7A).

Heterolithic sandstone/mudstone bodies of Facies B1 are found with a sharp contact on top of other facies and along truncated sides of sand-dominated intervals. Such contacts are concave-up in case the truncated sediments belong to Facies B3 or to Facies Association A and C. The concave-up nature of the contact may abruptly end and become bedding-parallel when it reaches the top of a
coal seam (Fig. 3.6A). Only a few sites show evidence for coal seams being truncated by a concave-up boundary (Fig. 3.6B). The upper inclined sand-mudstone alternations of Facies B1 pass laterally into the horizontally layered muddy sand-sandy claystone alternations of Facies B2 (Figs 6A and 6B). Facies B2 becomes more dominated by massive sandy claystones some hundreds meters away from the concave-up truncation of Facies B1, while it may gradually pass into coal seams ca 1 to 5 km from the truncation (Facies C1). The bleached zones of Facies B3 laterally thin toward the valley-fill of Facies B1. Stratigraphic boundaries between Facies B3 overlying Facies B2 and Facies Association C, are undulating (Figs 6E, 6H, 6I and 7A).

**Interpretation.** The sharp up to 15-m high concave-up boundaries of Facies B1, abruptly stopping on top of coal seams, argues for river channel incision creating fluvial valleys, while further incision was impeded or halted at underlying cohesive peat layers (Smith & Pérez-Arlucea, 2004; Van Asselen et al., 2009). The dm-scale cross-stratified sand-dominated lower part of Facies B1 is indicative of upper-flow regime dune bedform aggradation (e.g. Miall, 1985), when post-incisional waning of flow velocities in the channel prevented bypassing of coarser sands. Likely, such waning flow velocities also caused relatively large transported organic components, such as peat slabs (later compacted to coal rafts) and tree trunks, to be contained and preserved within the sand dunes. The peat slabs probably were cut from underlying peat by the downward force of incision. During incision, the petrified tree trunks were uprooted from levees at times of bank-full discharge or they tumbled into the channel at times of over-saturated and/or critically steepened valley sides.

The up to 3-m thick bleached marker zones of Facies B3 laterally thinning toward the valley-fill of Facies B1 argues for intense soil leaching on inter-valley areas both during and after incision, when post-incisional sedimentation was restricted to the adjacent lower-lying valleys (e.g. Kraus, 1999). The mainly white to light-grey soil profiles, developed in the top parts of the yellowish-grey sand- and mudstones of Facies Association A, point toward soil eluviation (E horizon). The cm- to dm-scale gold and/or dark-purplish-grey mottled and locally concreted horizons may represent the illuviation in B horizons. Just below the undulating top of Facies B3, the dark organic-rich patchy horizons are likely indicative of the local preservation of humic topsoils (i.e. A and O horizons). The bright-black horizons with undulating tops that developed over coal seams which laterally pass into the valley-fills of Facies B1, are interpreted to be the result of inter-valley soil-leaching over peat parent material, although distinct soil horizons were not recognized in this lithology.

The horizontally stratified mud-dominated units of Facies B2 that overlie the undulating tops of Facies B3 and laterally pass into the shallow inclined sand- and mudstone alternations of Facies B1 are interpreted in terms of a low-energy unconfined meandering bayou depositional system (e.g. Guccione et al., 1999). The shallow truncation surfaces found within the muddy low-angle inclined upper deposits of Facies B1 may be formed by intra-pointbar erosion and rotation (Durkin et al., 2015). The pieces of petrified wood horizontally aligned in the sand- and mudstone bed surfaces of Facies B2 may reflect driftwood that drifted in the bayou flood-basin during overbank flooding. Any lack of tidal indicators (e.g. ichnofacies, tidal-bundling) suggest that the inclined- and horizontally heterolithic sandstone/mudstone deposits of respectively Facies B1 and B2 were mainly fluvial-influenced (Eberth, 1996).

**Facies Association C – Peat Mire**

**Description.** Facies Association C (Fig. 3.7) is composed of Facies C1 (ca 10 %), C2 (ca 80 %), and C3 (ca 10 %). Facies C1 include up to 50-m wide, ca 0.5-m thick, dark-greyish-brown, lenticular, coaly shale layers (Figs 7A and 7B) and massive, dark-grey carbonaceous (muddy) shale (top hill,
Figure 3.7. Field photos of Facies Association C – Peat Mire. A: Facies C1. Several lenticular dark-greyish-brown coaly shale layers intercalated in Facies B2 ca 500 m E of the CMD-W section. Layers may have been locally formed on shallow-water vegetated islands during decreasing bayou water depths. B, C and D: Facies C1. Pieces of wood parallel to bedding of Facies B2 reflecting the settling of driftwood that was introduced in the flood-basin after high discharge events. E: Facies C2. Two coal seams of the U-coal zone exposed along Highway 24 at the RT-M section. Strong lateral continuity reflects a palaeo-environment dominated by peat
Fig. 3.7A). The coaly shales are mm-scale laminated and contain mm- to cm-scale vertical root traces, slicken-sides, and mm-scale plant fragments.

Facies C2 consists of up to 2-m thick, brownish-black (10YR 1.7/1) to black coal seams (Figs 7E, 7F and 7G). The coals consist of cm- to dm- scale alternations of bright (vitrain) bands with a conchoidal fracture (ca 75 %) and dull-black dusty coal (ca 25 %). Semi-round mm- to cm-size pieces of amber are present in the coals.

Facies C3 is made up of up to 5-cm thick, pinkish beige (2.5YR 7/3) to pinkish light-grey (5YR 8/1), very fine to coarse ash layers or whitish ash-dispersal in a coaly matrix (Fig. 3.7G). The ash contains white-transparent, euhedral, very-fine to coarse-grained crystals. Closely spaced individual ash layers may be clustered in zones up to 60-cm thick.

Lenticular coaly shales of Facies C1 are intercalated in Facies B2 and C2 (Fig. 3.7A). Coaly and carbonaceous shales of Facies C1 gradually overlie Facies B2, and gradually underlie and overlie Facies C2 and A3. Ash layers of Facies C3 are intercalated in Facies C1 and C2. Coaly and carbonaceous shales (C1) and coal seams (C2) sharply overlie Facies B3 and may gradually pass into Facies B2.

Interpretation. The coaly shale lenses of Facies C1 within the coal seams of Facies C2 may reflect the influx of clay from very distal splays in a shallow water peat mire environment. Coaly shale lenses of Facies C1 within the massive mudstones of Facies B2 may be interpreted as shallow water vegetated islands within a deeper water bayou flood-basin environment (e.g. Guccione et al., 1999).

Coals of Facies C2 and carbonaceous shales of Facies C1 laterally passing into the massive mudstones of Facies B2, that are laterally continuous, not being interrupted by interfingering channels of Facies Association A, and that do not pass into the massive mudstones of Facies B2, were likely formed in regionally extensive peat mires in which drainage was mainly diffusion-dominated in the absence of well-defined channels (Whitfield et al., 2009). The cm- to dm-scale alternations between dull-black and bright-black coal may reflect wet-dry cycles respectively (Potter et al., 2008; Holdgate et al., 2016; Korasidis et al., 2016).

Within Facies C1 and Facies C2, the intercalation of discrete ash layers and zones composed of euhedral, very-fine to coarse-grained crystals indicates a volcanic ashfall (tephra) origin of Facies C3 (Bohor & Triplehorn, 1993). The discrete ash layers may reflect tephra preservation in low-relief, possibly drowned peat mires while the ash dispersal intervals may point to rapid peat-forming conditions in raised mires (Triplehorn et al., 1991).
### Facies Associations of the Lebo Shale Member in north-eastern Montana

<table>
<thead>
<tr>
<th>Facies Association A – Channel-splay</th>
<th>Lithology, geometry and other contents</th>
<th>Sedimentary structures</th>
<th>Colour and present weathering</th>
<th>Unit boundaries</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>A1</strong> Large-sized channel with low-sinuosity, poorly confined flow, high clastic loads and rapid infilling leading to super-elevation and subsequent avulsion</td>
<td>Very-fine-, fine- and medium-grained sandstone. Up to ca 10-m thick and ca 0.25-km wide sandstone bodies. Cross-sectional geometry shows elongated sheets. Aerial image shows a nearly straight ridge (Fig. 1C).</td>
<td>In lower part dm-scale trough-cross-stratified sandstone. In upper part cm-scale trough cross-laminated layers and, climbing rippled cross-laminated layers</td>
<td>Greyish-yellow (2.5Y7/2), light-yellow weathering, locally indurated ribbons. Erosional resistant cover sandstone</td>
<td>Sharp erosional base, up to ca 4-m downward erosion, generally not through underlying coal. Generally a gradual transition when passing into overlying and juxtaposing A3</td>
</tr>
<tr>
<td><strong>A2</strong> Small-sized channel with confined flow, episodically high clastic loads and active to abandoned infilling</td>
<td>Muddy very-fine to fine-grained sandstone. Concentric fill: 1 – 15 m wide and 1 – 3 m thick. Width-depth ratio ca 4:8</td>
<td>Sets of 10 – 30 cm thick concentric muddy sandstone fills. Set contacts marked by concentric cemented levels.</td>
<td>Dull-orange-yellow (2.5Y 7/4), brownish-yellow weathering. Indurated at gold (10YR 6/8) concentric cemented levels</td>
<td>Sharply bounded along margins of concentric fill. Gradual transition with overlying A3</td>
</tr>
<tr>
<td><strong>A3</strong> Crevasse-splay; crevasse-channel, proximal, medial, distal to very-distal spays</td>
<td>Ca 0.5-m to 5-m thick packages consisting of dm- to cm-scale silty sand to mudstone fining-up sequences. Mudstones up to 1-m thick. Intercalation of up to 2-m thick lenticular fine-grained sandstone bodies. Small rootlets, leafs</td>
<td>Dm- to cm-scale horizontal- to slightly rippled-cross laminations. Cm- to mm- thick cemented levels (variegated). Mudstones slightly laminated or structureless</td>
<td>Dark-yellow (2.5Y 6/4) silty sandstone, brownish-yellow weathering. Greyish-olive (2.5Y 4/3) mudstone, light-grey weathering</td>
<td>Gradual transition to overlying carbonaceous shale layers and coal seams. Sharply overlying coal seams</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies Association B – Valley-related</th>
<th>Lithology, geometry and other contents</th>
<th>Sedimentary structures</th>
<th>Colour and present weathering</th>
<th>Unit boundaries</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>B1</strong> Fluvial valley created by channel incision. Lower valley fill by low-sinuosity, high-energy meandering sandy channels. Upper- and overfill of the valley by high-sinuosity low-energy muddy meandering bayou’s</td>
<td>Fine- medium sandstone-dominated lower part. Dm-scale alternations of very-fine- (&gt; 40 %) and muddy sandstone (&lt; 60 %) in upper part. Up to ca 25-m thick and ca 1-km wide sandstone bodies. Concave-up sides. Petrified trunks, peat rafts and organic residue in sandstones</td>
<td>Lower part: dm-scale trough-cross-stratified sandstone that can be overlain by rippled cross-laminated muddy sandstone (up to 30 % muddy laminae). Upper part: up to ca 60 % mud in dm-scale horizontal- and shallow inclined stratification. Inclined bedding up to ca 25°. Few truncation surfaces</td>
<td>Greyish-yellow (2.5Y 8/1) sandstone, light-grey weathering, locally indurated. Greyish-olive muddy sandstone (5Y 3/1), grey weathering, mud cracks. Aerial image shows a parabolic ridge (Fig. 1C)</td>
<td>Sharp erosional base, down-cutting, occasionally through underlying coal. Concave-up side, generally at boundary with Facies A. Gradual transition with overlying B2</td>
</tr>
<tr>
<td><strong>B2</strong> Bayou flood-basin receiving dominantly muddy sediments and driftwood from adjacent channel flooding</td>
<td>Up to ca 6-m thick mud-dominated units. Massive sandy clay (&gt; 80 %) and muddy sandstone (&lt; 40 %). Coalified root traces and slickensides in some levels. Petrified wood aligned with bed surface</td>
<td>Dm-scale alternations between slightly laminated, mainly massive sandy clay and slightly mm- to cm-scale rippled laminated muddy sandstone</td>
<td>Dark-greyish-olive (2.5Y 2/1) to brownish olive (2.5Y 4/4) sandy clay/ mud. Greyish-olive (5Y 5/2) muddy sandstone. ’Dirty-popcom’ weathering</td>
<td>Sharply overlying B3, at undulating surface between B2 and B3. Gradual transition if over lain by C1 and underlain by B1. Laterally passes in Facies C</td>
</tr>
<tr>
<td><strong>B3</strong> Inter-valley soil with sporadically-preserved A, thick E and thin B horizons. Soil formation in between valleys took place during incision and confined valley flow</td>
<td>Up to 3-m thick light-coloured (bleached) zones, marking areas for several km’s horizontal distance. Heterolithic: from sandstone, mudstone, to coal. Many (coalified) root traces visible in sand- and mudstones. Slickensides in mudstones</td>
<td>Internal structures. Poorly-developed, locally-concreted horizons with up to ca 50 % mottles in bleached sandy matrix. Locally, just below undulating top, an up to ca 20 cm thick dark sandy or muddy horizon with coaly patches, burrows and small ash lenses</td>
<td>Pale-beige (7.5YR 8/1) whitish-weathered sandstone. Beige (7.5Y 6/1) light-grey-weathered mudstone. Gold (2.5Y 7/8) and dark purplish grey (5P 4/1) mottles and concretions. Brownish black (10YR 1.7/1) in top</td>
<td>Laterally passes into B1. Undulating surface at top of bleached zone or dark horizon. Frequently sharply overlain by B2 and sporadically sharply by D2. If darker horizon is overlain by D2 the undulating contact between them is barely visible</td>
</tr>
<tr>
<td>Facies Association C – Peat mire</td>
<td>Lithology, geometry and other contents</td>
<td>Sedimentary structures</td>
<td>Colour and present weathering</td>
<td>Unit boundaries</td>
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<tr>
<td>C1 Shallow lacustrine, locally (pond) or regionally (lake) inundated mire with little clastic input</td>
<td>Up to ca 0.5-m thick lens-shaped organic coaly shale to more laterally extensive massive carbonaceous (muddy) shale. Root traces, slicks and, plant remains in coaly shale</td>
<td>Mm-scale laminated coaly shale. No clear sedimentary structures in massive carbonaceous shale</td>
<td>Dark-greyish-brown (5YR 2/2) coaly shale. Dark-greyish-olive (2.5Y 2/1) massive carbonaceous shale</td>
<td>Coaly shale lenses in B2 and C2. Coaly and carbonaceous shale sharply overlying B3, gradually overlying B2 and gradually underlying or overlying C2 or A3</td>
</tr>
<tr>
<td>C2 Peat mire, mostly extensive swamp environments dissected by low sediment-load peat-drained rivers</td>
<td>Lignite rank coal in up to ca 2-m thick seams consisting of cm- to dm- scale alternations between bright (vitrain) bands with a conchooidal fracture (ca 75 %) and dull-black dusty coal (ca 25 %). Presence of amber. Lateral tracing of coal seams up to several km’s</td>
<td>No sedimentary structures</td>
<td>Black dusty and bright coal to brownish black (10YR 1.7/1) bright coal</td>
<td>Relatively sharply overlying and underlying all other facies except for a more gradual transition with C1</td>
</tr>
<tr>
<td>C3 Volcanic ashfall from single volcanic eruptions in the western Cordilleran Thrust Belt. Preservation of ashfall layers particularly across low-relief mires</td>
<td>Up to ca 5-cm thick discrete ash layers or ash intervals. Grain size range from very fine to coarse. Euhedral crystals. Ashy layers laterally pinch out and merge, up to lateral distances of tens of km’s. Closely-spaced individual ash layers make up ash zones</td>
<td>No sedimentary structures</td>
<td>Pink beige (2.5YR 7/3) and pinkish light-grey (5YR 8/1) ash. Platy weathering of light-coloured, indurated ash intercalations within softer and darker-coloured coaly matrices</td>
<td>Sharply-bounded intercalations in coal or organic shales and mudstones</td>
</tr>
</tbody>
</table>
Fence panel

Chronostratigraphic markers

Three major chronostratigraphic markers have been used for correlation of the sections. The stratigraphically lowest marker is a pinkish light-grey (2.5YR 7/2), up to 3-cm thick, medium- to coarse-grained ashfall layer containing euhedral-shaped and transparent sanidine minerals. It is consistently found in the top part of a ca 1-m thick coal seam in the upper part of the C29n magnetochron and can be traced over tens of kilometres. Sieving of grain-size fractions in the laboratory shows that this tephra mainly has a grain-size range of 250 to 550 μm, whereas grain sizes of other tephras in the study area range between 62.5 and 200 μm. The coarse tephra layer is unique in its grain size and represents a key layer for stratigraphic correlation. The transparent light-grey crystals and the grain-size resemble crystals of granulated sugar. Therefore, we named this tephra “Sugar Ash”.

The second marker is the C29n/C28r geomagnetic polarity reversal. This polarity reversal occurs ca 6 m above the Sugar Ash at Coal Mine Divide West (CMD-W), Coal Mine Divide Main (CMD-M) and Radiotower Northeast (RT-NE). At CMD-W and CMD-M, the C29n/C28r reversal occurs within a 1- to 2-m-thick coal seam (Figs 3A and 8), labelled the U-coal by Collier & Knechtel (1939) (Fig. 3.1D). In the southernmost part of the transect, at Horse Creek Road Main (HCR-M), the C29n/C28r reversal occurs in the lower part of a clastic interval directly above a whitish mature, 0.75-m-thick palaeosol, ca 1.5 m above the Sugar Ash (Figs 3E and 8). The Horse Creek Road sections are located in the type area of the U-coal of Collier & Knechtel (1939) (Fig. 3.1). The U-coal at HCR-M occurs 17 meters higher than the C29n/C28r reversal, at the stratigraphic position of the C28r/C28n reversal (Fig. 3.8).

The stratigraphically highest marker is the C28r/C28n geomagnetic polarity reversal. It occurs in a carbonaceous shale and mudstone interval, ca 20 m above the U-coal of Collier & Knechtel (1939) at Coal Mine Divide Main (CMD-M), Rough Prong East (RP-E), and Rough Prong Main (RP-M) (Fig. 3.8). At Radiotower Main (RT-M), Radiotower Southeast (RT-SE) and Horse Creek Road Main (HCR-M), the C28r/C28n reversal occurs within a 1- to 2-m-thick coal seam that was mapped as the U-coal by Collier & Knechtel (1939) (Fig. 3.8).

Coal nomenclature: the U-coal revisited

Correlation of the polarity reversals shows that the U-coal of Collier & Knechtel (1939) in the north and the U-coal of Collier & Knechtel (1939) in the south cannot be the same (Figs 1D and 8). They are different coals stratigraphically separated from each other over a time interval that encompasses the entire Chron C28r, ca 224 kyr (Dinarès-Turell et al., 2014). Thus, the mapping of the U-coal bed in the study area by Collier & Knechtel (1939), is inconsistent with respect to current chronostratigraphic results and needs to be revised. In updating the U-coal labelling across the study area, the underlying V-coal and overlying T-coal also have to be re-assigned. On the premise of keeping the original U-coal label in its type area (Collier & Knechtel, 1939), the U-coal in the south of the fence panel, at the stratigraphic position of C28r/C28n, is retained. It is now, however, defined as a zone consisting of multiple closely spaced coals (Fig. 3.8). As a result, the U-coal as mapped by Collier & Knechtel (1939) in the north of the fence panel, at the stratigraphic position of C29n/C28r, is now untenable. In that region, this coal is included in the V-coal zone consisting of a cluster of three coal seams with the Sugar Ash in the middle seam (Fig. 3.8). The V-coal zone overlying the upper part of the W-coal zone (i.e. #8-W coal) is consistent with Noorbergen et al. (2018). The T-coal zone is re-assigned to the coal zone above the bleached palaeosol at RTSE.
Coal correlation

Chronostratigraphic correlation of the three markers (Fig. 3.8) shows that some coal seams can be correlated along the whole transect of the fence panel while others cannot. The basal coal in the upper W-zone shows lateral continuity, except just south of RT-NE where it is truncated by a sandstone complex over a distance of ca 500 meter (white arrow, Fig. 3.6B). Overlying the W-zone, in the V-zone, two coal seams also show lateral continuity: the coal seam of the Sugar Ash and the coal seam below that. Correlation of the C29n/C28r reversal shows that the revised U-coal of Collier & Knechtel (1939), now in the upper V-zone, is laterally discontinuous. This coal is ca 2-m-thick in the north. To the centre of the fence panel, it becomes thinner, with a thickness of 0.5 m. It is absent in the south of the fence panel at the contact of a whitish palaeosol (Facies B3) disconformably overlain by clastic fluvial deposits (Facies B2).

Correlation of the C28r/C28n reversal shows that the stratotype U-coal of Collier & Knechtel (1939) is also laterally discontinuous. This coal is ca 2-m-thick in the south. To the centre of the fence panel, it becomes thinner with a thickness of 1 m. It is absent in the north of the fence panel where, instead, massive mudstones of Facies B2 occur. Laterally continuous coal seams are shown above the C28r/C28n reversal (upper part, Fig. 3.8). Above the C28r/C28n reversal, additional chronostratigraphic correlations are lacking. Coal correlations in this interval are mainly based on lithostratigraphic interpretations that are locally supported by field tracing of coals and bleached zones.

Successions and sequences

The new chronostratigraphic correlations allow for further interpretation of the stratigraphic architecture of the sediment successions and sequences (Table 3.1) at m-scale resolution (Fig. 3.8). Two scales of alternations are identified. The first are ca 5-m-thick coal-clastic aggradational successions (CCS) consisting of a peat mire facies association (C) overlain by a channel-splay or valley-related facies association (A or B). A total of ten CCS are identified in the fence panel. Superimposed on the ca 5-m-thick CCS are ca 15-m-thick aggradation-incision sequences (AIS). At the base these sequences start with the aggradation of Facies B1 and B2 deposits on top of ca 20-m thick valley-base and inter-valley undulation surfaces. Overlying Facies B1 and B2, are the aggradations of ca three to four CCS. The top of the AIS are in the last CCS where the aggradation stops at the undulating top surface of a bleached zone (Facies B3) laterally passing into the valley-base of Facies B1 representing the incision. A total of three AIS are identified in the fence panel (Fig. 3.8). The lowermost AIS (AIS-1) only shows its uppermost part. The overlying AIS-2 has the most complete architecture with 5- to 15-m aggradation. AIS-2 is laterally truncated in its top part by two
valleys filled by Facies B1 and related to soil zones of Facies B3 occurring in between the valley-fills on elevated platforms. On top of AIS-2, AIS-3 is composed of 15- to 20-m aggradation in the south where it is covered by a bleached zone in the top (Facies B3). To the north, the soil formation zone exceeds the upper fence panel margin because the full thickness of AIS-3 is not exposed in that area. Chron C28r comprises the uppermost aggradation phase of AIS-2, the incision phase of AIS-2, and the lowermost aggradation phase of AIS-3 (Fig. 3.8).

A first-order quantification of minimal incision depth is calculated by the difference between the thickness of the full valley fill (Facies B1 and B2) at the place of deepest incision and the thickness of Facies B2 above the inter-valley soil (Facies B3). The latter locality has seen significantly less erosion during the incision phase. We cannot account for differences resulting from sand/mud compaction and depositional thickness due to palaeo-topographic effects at the time of deposition. The minimal incision depth of AIS-1 is 8.3 m and AIS-2 is 5.1 m.

Compensational stacking patterns
Both at the scale of CCS and AIS compensational stacking patterns are observed (Fig. 3.8). Facies become thicker at places where underlying facies were thinner and vice versa. For the successive CCS, these patterns are shown by the Facies Association A sediments. Examples are in the V-coal zone of RP-SW and in the U-coal zone of RP-M, RP-SW, RT-M, and RT-SE (Fig. 3.8). An example of compensational stacking for AIS is observed between RP-SW and RT-NE, where Facies Association B deposits of AIS-3 thin from the north and from the south above thick Facies Association B deposits of AIS-2. These lateral thickness variations of facies likely reflect the effect of palaeo-topography on sedimentation. The lateral extent of compensational stacking for CCS is at ca 10 km and for AIS at ca 15 km (Fig. 3.8), showing that sedimentation evened out over these distances.

With the compensational stacking of the CCS, coals (Facies C2) do not gradually pass into lateral clastics of Facies Association A. This suggests the absence of syndepositional channel-belts of Facies A1 coeval with peat formation. The coals in the lower U-zone do gradually pass into the massive mudstones of Facies B2 (Fig. 3.8), suggesting that these facies co-existed.

Time control and duration
The two geomagnetic polarity reversals spanning chron C28r and the Sugar Ash provide constraints on durations of coal-clastic successions (CCS) and aggradation-incision sequences (AIS). The duration of chron C28r is 291 kyr in the Geological Time Scale 2012 based on astronomical tuning (Vandenberghe et al., 2012). More recent estimates on the duration of C28r arrive at 224 kyr by combining several astronomically tuned chronologies in Spain, the Atlantic, and the Pacific (Dinarès-Turell et al., 2014), at ~ 358 kyr based on 206Pb/238U radioisotope ages of tephras in the Kiowa Core of the Denver Basin (Colorado, USA) in combination with a revised magnetostratigraphy (Clyde et al., 2016), and at 252 kyr using 40Ar/39Ar data of two ash-layers below (= the Sugar Ash) and above C28r in the Fort Union Formation, Garfield County, Montana (Sprain et al., 2015). Although these estimates range from 224 – 358 kyr, it is important to note that chron C28r corresponds to a 405-kyr eccentricity minimum as evident from the integrated stratigraphic framework of Dinarès-Turell et al. (2014). A closer look at the short-eccentricity scale further reveals that C28r apparently starts in a 100-kyr eccentricity maximum or on the transition from a 100-kyr maximum to minimum, and that C28r contains two 100-kyr minima (Fig. S3.1 A-C; green arrows) and the maximum in between.

The above implies that the valley-related Facies Association B sediments of AIS-3 are deposited somewhere during the 400-kyr minimum. To develop a conceptual model of long-eccentricity
forced fluvial stratigraphy, it is relevant to locate the moment of incision followed by aggradation in relation to this 405-kyr long cycle. Based on Figure S3.1 it seems that the incision of AIS-2 occurs during the transition of a long-eccentricity maximum to minimum or in the minimum itself.

Uncertainties in the exact position of C28r chron boundaries (i.e. grey intervals) in the Lebo Shale do not allow to decipher the exact phase-relation of the sedimentary succession with the 100-kyr short-eccentricity cycle, as known from the marine realm (Dinarès-Turell et al., 2014). Besides, it is important to note that the 100-kyr eccentricity cycle in contrast to the 405-kyr cycle is not reliably represented in the astronomical solution, due to chaotic behaviour of the Solar System (Laskar et al., 2011). Therefore the timing of incision and the subsequent onset of valley infill with respect to the short-eccentricity-cycle remains elusive.

DISCUSSION

Controls on the aggradation-incision sequences (AIS)

Tectonics
At 400-kyr timescales flexural foreland-basin tectonism may exert a control on building reciprocal fluvial stratigraphy (Miall, 2014). Foreland-basin flexural tectonism is the flexural behaviour of the foreland lithosphere in response to successive stages of orogenic loading and unloading (Beaumont, 1981; Catuneanu et al., 1997). As a result, creation and disappearance of accommodation space and associated sequence patterns will be out of phase between the proximal and distal regions of the foreland basin which separation is marked by a foreland system hinge line. Based on bentonite and ammonite-zone correlations in Campanian to Paleocene strata in the Western Interior Foreland Basin (particularly Canada area), Catuneanu et al. (2000) made a reconstruction of the hinge line position during seven consecutive time intervals within the Campanian to Paleocene, providing an indication of ca 3-Myr stages of foredeep migration. Sediments of the Lebo Shale Member were deposited in the study area when the hinge line was located ca 150 km westward (Catuneanu et al., 2000). The associated westward retreat of the foredeep towards the Cordilleran orogeny during the early Paleocene may reflect an episode of orogenic quiescence (Catuneanu et al., 2000). This quiescence is in agreement with eastward advancing Laramide thrust-sheets later in time, representing the middle Paleocene Bighorn Uplift and the early Eocene Black Hills uplift (Belt, 1993; Belt et al., 2004).

Tectonic control on foreland accommodation, with its associated positive or negative space for sedimentation, may overwhelm time-overlapping orbital controls (i.e. long to very long eccentricity cycles) or it may remove stratigraphy that was formed by shorter-term controls (e.g. short-eccentricity cycles or autogenic processes) due to upwarping. Although acting on 1- to 10-Myr timescales (Catuneanu et al., 2000), it seems less likely that flexural tectonism in the Western Interior Foreland Basin has exerted a dominant control on fluvial aggradation or incision at 400-kyr timescales.

Autogenic compensational stacking
The absence of syndepositional channel-belts of Facies A1 coeval with peat formation (Facies C2) suggests that alternations between these facies (CCS) are not formed by autogenic avulsion. More specifically, an autogenic avulsion control on peat formation in fluvial systems would start with a syn-depositional channel-belt system adjacent to backswamp peat formation; with time clastic-peat
interfingering would be observed (e.g. McCabe, 1984). If the channel-belt becomes super-elevated with respect to the peat, avulsion may be initiated (e.g. Fielding, 1984; Van Asselen et al., 2009). The peat formation of Facies C2 decoupled from autogenic channel avulsions possibly reflect regionally extensive peat mires in which drainage was mainly diffusion-dominated in the absence of high sediment load, well-defined channels (Whitfield et al., 2009). The 100-kyr time scale involved with the recurrence of Facies C2 argues for a cyclic allogenic control. Similar to the underlying Tullock Member, the major peat-forming phases may originate from changes in the amount of sediment supply allogenically driven by short-eccentricity-paced climate changes (Noorbergen et al., 2018).

Nevertheless, autogenic avulsions or channel migrations played an important role in the stratigraphic stacking within the clastic parts of the CCS, i.e. within the channel-splay dominated systems (Facies Association A, Table 3.1). This is supported by the 10-km scale compensational stacking patterns following lateral thickness variations of Facies Association A (Fig. 3.8). Hajek & Straub (2017) show experimental and statistical evidence indicating that compensational stacking in clastic fluvial systems can be fully explained by autogenic self-organization, along with avulsion or lateral migration processes transporting sediment into available lateral space. The patterns of compensational stacking along the successive CCS as observed in the fence panel is probably a consequence of local accommodation space created by early-stage peat compaction. As a result, existing clastic reliefs were relatively enhanced promoting autogenic self-organization within short-eccentricity-paced renewed stages of clastic sediment supply. Palaeo-topographic differences associated with compensational stacking locally could have determined the thickness and timing of later peat formation, such as delayed peat formation over an alluvial ridge as reflected by thinning of the upper V-coal over a sand-body in section RP-SW (Fig. 3.8).

On a smaller scale an autogenic avulsion model may be valid for specific intervals, for example the lower U-coals that gradually passes into the lateral, muddy sediments of Facies B2. Here, peat compaction or channel infilling may have caused super-elevation, triggering muddy channel avulsions. Nevertheless, the increasing lateral continuity of coal seams overlying Facies B1 and B2 points to expansion of peat mires. This can be explained by channel (and avulsion) starvation as the consequence of climatically driven reduced sediment supply.

**Downstream base-level impact on aggradation and incision**

The stratigraphic architecture of the three AIS (Fig. 3.8) resembles the sequence stratigraphic model of Wright & Marriott (1993). This model shows the maturity of soil development in relation to a geomorphic base-level control on fluvial aggradation and incision. Applied to the Lebo Shale Member, the aggradational part of the AIS, consisting of the successive CCS, would correspond to the high base-level/relative sea-level stages of Wright & Marriott (1993). Enhanced floodplain clastic sedimentation and weak soil development (conform Facies Association A) would be favoured if accommodation rates increased during base-level rise. In the subsequent highstand, reduced accommodation rates caused decreased clastic floodplain sedimentation, favouring increased soil development (conform FA-C). The incision part of the AIS representing the valley formation (lower part Facies B1) and inter-valley soil-leaching (Facies B2) would then correspond to the low base-level stages of Wright & Marriott (1993). Namely, increased erosion creating fluvial valleys and coeval strong soil development on the terraces would be favoured if accommodation rates were low, during a lowstand. In the subsequent early-stage of base-level rise, slightly increased accommodation caused channel amalgamation and hydromorphic soil formation (upper part Facies B1 and Facies B2).
The fluvial sequence stratigraphic model of Wright & Marriott (1993), however, is not universal. We have to consider several issues before we can deem it a valid model for the observed AIS in the Williston Basin. Firstly, base-level fall causing incision applies to settings where coastal plain gradients are significantly smaller than shelf gradients. If, vice versa, base-level fall may lead to aggradation (Schumm, 1993; Blum & Törnqvist, 2000). Secondly, experimental studies have shown that rivers can reorganise themselves in response to base-level changes by adapting their channel sinuosity, especially in settings with small differences between coastal plain and shelf gradients (Lane, 1955; Schumm, 1993). Thirdly, the landward limit of base-level control on fluvial aggradation and incision varies with coastal plain gradients, ranging from ca 350 km for low-gradient, large river systems to ca 40 km for steep-gradient, small rivers (Blum & Törnqvist, 2000).

To determine the specific response of the Lebo Shale fluvial system to base-level change a detailed palaeo-environmental reconstruction is required. A basin-scale examination of fossil flora preserved within the Paleocene sediments of the Fort Union Formation indicates that this area was covered with lowland swamp vegetation extending for over ca 300 kilometres, from the Cannonball Sea in the east to the foothills of the Rocky Mountains in the west (Brown, 1962). A basin-scale quantitative analysis of the megafloral record shows that vegetation was dominated by dicotyledonous angiosperm species, i.e. one major group of flowering plants, for 79 % (Peppe, 2010). The brackish water tongues of the Cannonball Member, exposed along the Little Missouri River (Fig. 3.1B), are composed of dark black-brown mudstones that were likely deposited in lagoons during terrigenous inflow of clay and silt, derived from rivers draining a low-gradient coastal plain (Van Alstine, 1974). The overall fine grain sizes, brackish faunal associations (i.e. bivalves, crabs, ostracods, and benthic foraminifera), and sedimentary facies associations (i.e. salt marsh, tidal flat, tidal channels, lagoon, mainland beaches, shoreface barriers, and shelf) within the Slope and Cannonball Members have been linked to the comparable coastal setting of the northern Netherlands (Cvancara, 1972; Van Alstine, 1974; Fenner, 1974; Cvancara, 1976). Nevertheless, Lindholm (1984) further specified that “the Cannonball body of water” in North-Dakota may have been one big lagoon, instead of a sea, and that different directions of sediment supply over the lagoon resulted in varying depositional environments and faunal compositions. Low continental shelf-gradients likely continued for several hundreds of kilometres, following a shallow inland continental seaway until open sea/ocean waters were reached in the south, and/or in the north (Cvancara, 1986).

Given the palaeo-environmental reconstructions, longitudinal profiles during deposition of the lower Fort Union Formation were of low-gradient, with the gradient of the coastal plain being higher than the gradient of the shelf. At the Caribbean and/or Arctic shelf break, the gradient probably steepened along a continental slope. Such long continental margin profiles may not exist today. Nevertheless, the shape of the shelf that would have been added to river profiles within the Lebo Shale Member if sea-level fell, may be comparable to the Malcolm River (Alaska, USA) as described by Miall (1991). Here a low-angle continental shelf would be added to a steeper-angle, concave coastal plain during falling sea-level (Miall, 1991). Such a situation may provide increased accommodation, i.e. space for aggradation, if the difference between the coastal plain gradient and the shelf gradient was significant (Miall, 1991; Schumm, 1993; Blum & Törnqvist, 2000). Opposed to the sequence model of Wright & Marriott (1993), a base-level fall more likely resulted in fluvial aggradation rather than fluvial incision within the Lebo Shale Member in north-eastern Montana. Moreover, with a falling base-level causing headward incision, deep fluvial valleys and mature inter-valley soils would be expected to be present proximal to the coast, but such features have not been
documented within the time-equivalent fluvial and deltaic strata of the Slope Member in western Dakota (Belt et al., 1984; Hartman, 1993; Warwick et al., 2004; Peppe et al., 2009).

Alternatively, a base-level rise in this system caused flattening of the fluvial graded profile with the new profile intersecting the old at the terrace intersection: the point between net aggradation at the downstream side and net incision at the upstream side (Pons, 1957; Van Dijk, 1991; Tornqvist, 1993; Berendsen & Stouthamer, 2001; Westerhoff et al., 2003). Transgression might have been initially delayed at the delta mouth due to compensation by sediment supply (Berendsen & Stouthamer, 2001), but continued sea-level rise caused upstream migration of the terrace intersection and, hence, fluvial aggradation propagating in the same direction (e.g. illustrations in Pons, 1957; Westerhoff et al., 2003). In the lowland palaeo-environment of the Western Interior Lebo Shale, it seems unlikely that the terrace intersection was ever situated downstream of the study area. Nevertheless, if prior to sea-level rise this was the case, the rise may have exerted control on river incision.

**Base-level changes during long-eccentricity cycles**

Looking specifically at eccentricity driven climate scenarios, 405-kyr cycles may have affected global (eustatic) sea-level changes in greenhouse worlds (Wendler et al., 2014) and hence may have controlled downstream base-level changes. Recent hydrological mass-balance calculations indicate that ca 30-m global-sea level changes in the mid-Cretaceous greenhouse world, with negligible glacio-eustasy, may result from orbital-paced changes in global groundwater storage: i.e. aquifer-eustasy (Wendler et al., 2016). Eustatic sea-level fall would be in response to net storage of water in continental aquifers during an intensified hydrological cycle and enhanced fluvial run-off, while a relaxed hydrological cycle causes aquifer discharge exceeding aquifer charge inducing eustatic sea-level rise (Wendler & Wendler, 2016; Sames et al., 2016). This scenario envisages eustatic sea-level lowstands during 405-kyr long-eccentricity maxima as a consequence of an intensified hydrological cycle during eccentricity modulated high-amplitude precession forcing and eustatic sea-level highstand during 405-kyr long-eccentricity minima as a consequence of a relaxed hydrological cycle during low-amplitude precession-forcing. The aquifer-eustatic sea-level would be falling during transitions from long-eccentricity minima to maxima and rising viceversa. A rising sea level may exerted a control on incision only when the terrace intersection was downstream with respect to the study site during the rise.

**Upstream climate**

A plausible alternative for downstream base-level rise is that regional climate changes occurring upstream and paced by long-eccentricity control the AIS formation in the Lebo Shale Member. Upstream regional climate, predominantly through the effect of precipitation and temperature changes, exert a significant influence on fluvial environments by its control on vegetation, weathering, erosion, discharge and sediment supply (e.g. Knox, 1972; Summerfield, 1991; Vandenberghe, 2003). Several studies have shown how river channels adjusted their patterns and styles in response to different changes in hydrological regimes related to Quaternary climate change (e.g. Schumm, 1968a; Baker, 1977; Baker & Penteado-Orellana, 1977; Bogaart & Van Balen, 2000; Bridgland, 2000; Goodbred Jr, 2003; Tandon et al., 2006; Roy et al., 2012; Blum et al., 2013; Vandenberghe, 2015). Baker & Penteado-Orellano (1977) studied the response of the Colorado River in central Texas (USA) to late Quaternary climate change and suggested that incision is triggered by enhanced occurrence of rare high-magnitude flooding events in an arid glacial climate. Nevertheless, such a fluvial response to arid climate contradicts with the Paleocene-Eocene Thermal
Maximum (PETM) of the Bighorn Basin (USA), where an arid climate regime is responsible for fluvial sheet-sandstone aggradation instead of incision (Foreman, 2014). The observation that the valley-related facies of AIS-2 occur toward or during a stable climate without major extremes, i.e. long eccentricity minima (Fig. S3.1), another process than high-magnitude peak discharges in an arid climate might have caused the incision. Given the three to four 100-kyr CCS prevailing between the valley-fills of the three AIS, the AIS-1 and AIS-3 incisions do likely share the same phase relation with long-eccentricity as the AIS-2 incision.

Unfortunately, it remains elusive at this stage whether the fluvial incisions started towards or within long-eccentricity minima due to resolution constraints in the age model. Nonetheless, the prolonged reduction of the precession-amplitude associated with long-eccentricity minima could have resulted in a persistent relaxation of the hydrological cycle over approximately 200 kyr long intervals. The presence of higher frequency precession-induced climate variations that occur superimposed on this stable background climate may have determined the onset of incision.

Fluvial incision could only have been triggered if discharge significantly exceeded sediment supply (e.g. Blum & Törnqvist, 2000), although both absolute amounts of discharge and sediment supply are expected to diminish in a climate with less extremes (i.e. low amplitude precession cycles). A constant supply of moisture to the source areas of the drainage network might have prevailed, promoting the development of a dense vegetation cover upstream. This upstream vegetation reduces erosion and thus sediment supply to the drainage system (e.g. Schumm, 1968b). On the other hand, discharge might have remained stable, especially if not significantly reduced by high evapotranspiration (Bogaart & Van Balen, 2000). High seasonal rainfall events during precession minima might have especially caused peak discharge events that cannot be buffered by evapotranspiration in the densely vegetated areas. Therefore, the AIS incisions in the Lebo Shale Member might have been triggered in particular at times of precession minima during long-eccentricity minima.

The impact of vegetation on fluvial incision during global sea-level rise agrees well with the fluvial terraces in northwestern Europe that were created by river incision during Late Glacial global sea-level rise (Bridgland, 2000; Vandenberghe, 2015) and has been confirmed by numerical modelling (Bogaart & Van Balen, 2000). In this case fluvial incision is indirectly linked to the upstream development of vegetation, with relative low evapotranspiration, causing a strong reduction in sediment supply compared to discharge.

The sediments eroded by the AIS incisions in the Lebo Shale Member would have been deposited downstream, especially when entering the low-energy flow regime of the Cannonball lagoon (Lindholm, 1984). Sediment deposition along the downstream-flattening long-profile may have resulted in local delta progradation overruling the aquifer-eustatic sea-level rise component (Wendler & Wendler, 2016). It might thus be that long-eccentricity upstream climate control on fluvial incision causes local-scale progradation and, hence, locally an opposite sea-level motion, i.e. lowering, than the rise predicted by aquifer-eustasy.

The termination of fluvial incision could have taken place when at a certain valley-floor depth the equilibrium fluvial graded profile was reached at which discharge and sediment supply were in balance. As an alternative for or in addition to the lowering valley floor, the continuing aquifer-eustatic sea-level rise during the long-eccentricity minima could also have caused the termination of incision and likely enabled the switch to aggradation. Namely, the backwater-induced rising groundwater table inundating the valleys causes a marked reduction of the stream power in the river channel. As a consequence, enhanced sediment supply over discharge stopped incision and
enabled the constitution of the fine sediment load bayou system responsible for the mud-prone valley-infill aggradation.

**Conceptual stratigraphic model**

Figure 3.9 shows stratigraphic architectural models, depicting an AIS formed over one 405-kyr long-eccentricity cycle. Concerning the formation of a CCS, the current model adopts the short-eccentricity-scale model of Noorbergen et al. (2018, option 2).

**Stage I – Peat aggradation**

Low amplitude precession cycles during a long-eccentricity minimum constitute for a stable climate without major seasonal extremes. This causes a relaxed global hydrological cycle. Relatively low hinterland erosion causes sediment supply to be approximately equal to discharge resulting in a fluvial graded profile equilibrium. In combination with an aquifer-eustatic sea-level highstand, the high base-level provides suitable conditions for prolonged accumulation of widespread peat across densely vegetated mires. Drainage is diffusion-dominated and occasionally along low sediment load, small, laterally migrating channels.

**Stage II – Clastic aggradation**

High amplitude precession cycles during a long-eccentricity maximum constitute for an unstable climate with major seasonal extremes during the precession minima. These stages cause an intensified hydrological cycle. High hinterland erosion during the precession minima causes sediment supply to significantly exceed discharge resulting in clastic aggradation. During wet season(s) sediments are distributed by splays and laterally migrating channels while during dry season(s) lowered groundwater-levels induce sediment oxidation on the overbanks. The high sediment supply along the graded fluvial profile causes deltaic progradation. In combination with an aquifer-eustatic sea-level lowstand sea-level further regresses resulting in a base-level drop.

**Stage III – Incision and inter-valley soil formation**

Lowered amplitude precession cycles toward or within a long-eccentricity minimum causes climate to stabilize because of a reduction in seasonal contrasts. This causes relaxation of the global hydrological cycle. Aquifer-eustatic sea-level rise results in base-level rise and longitudinal profile flattening but probably the terrace intersection never migrate downstream of the study area in this low-gradient setting. Therefore, the base-level rise is unlikely to control incision. Long-term constant moisture supply to the hinterland source areas cause an expanding vegetation cover upstream with increasing, but still relatively low, evapotranspiration. Sediment supply reduces due to upstream reduction of erosion by the vegetation cover and is now in deficit compared to discharge. The channel incises and creates a valley. Repeated incisions during high discharges, particularly during the higher seasonality precession minima, confines the stream flow along the valleys. Groundwater levels lower towards the river and leached inter-valley soil profiles develop. The eroded sediments are deposited coastward as a consequence of reduced energy conditions. As a result, this sediment supply may locally overrule aquifer-eustatic sea-level rise causing progradation at those locations.

**Stage IV – Channel stabilization and incipient peat formation**

Low amplitude precession cycles during a long-eccentricity minimum constitute for a stable climate without major seasonal extremes. This causes a relaxed global hydrological cycle. The vegetation
cover upstream is at its maximum extent causing very low sediment supply downstream. Discharge exceeds sediment supply at times of reduced evapotranspiration, mostly during slightly enhanced seasonality in the low-amplitude precession minima. At the same time, base-level rises to its maximum point during an aquifer-eustatic sea-level highstand. The backwater effects of the marine transgression constitute for high groundwater levels upstream. The valleys inundate and the stream power of the river channel reduces promoting aggradation when particularly muddy sediments are sporadically supplied. When the indurated inter-valley soils are inundated there is energy dispersal over the extensive platforms which give rise to the development of a low-energy muddy bayou system. The stable wet climate conditions in combination with the high groundwater levels cause high organic production and the accumulation of these organics produces peat bodies (Stage V, equivalent to Stage I).

Figure 3.9. Conceptual stratigraphic model illustrating the role of long-eccentricity-scale climate change in building aggradation-incision sequences (AIS) in the Lebo Shale Member in north-eastern Montana (left panel) and in building down-gradient lower Fort Union stratigraphy in the Williston Basin (middle panel). Phase relations with long- (purple) and short- (green) eccentricity and precession, including time-uncertainty, are shown on the left. A detailed description of each stage is provided in the text.
Conclusions

Age control and time-stratigraphic correlations in a ca 15 km long stratigraphic fence panel of the fluvial Lebo Shale Member in north-eastern Montana reveal the presence of ten 100-kyr-scale coal-clastic successions (CCS) and three 400-kyr-scale aggradation-incision sequences (AIS). Magnetostratigraphic correlation shows that polarity chron C28r, with a duration of 224 kyr (Dinarès-Turell et al., 2014), spans the upper part of AIS-2 and lower part of AIS-3 (Fig. 3.8). This implies that the second fluvial incision phase, i.e. the incision of AIS-2 corresponds to the transition from long-eccentricity maximum to minimum or to the minimum itself (Fig. S3.1). Separated by three to four short-eccentricity-scale CCS (Noorbergen et al., 2018), the AIS-1 and AIS-3 incisions seem to occur during the same long-eccentricity configurations.

The origin of the AIS follows from a critical evaluation of possible tectonic, base-level, and upstream climate control on their formation. Foreland basin flexural tectonism, and associated cyclic creation of positive or negative space for sedimentation, acted on ca 5-Myr timescales in the Western Interior Basin between the Campanian and Paleocene. Some million years of orogenic quiescence during the early Paleocene of north-eastern Montana is suggesting, however, that tectonic forcing in this time interval played a minor role. At times of deposition, coastal plain gradients were likely higher than shelf gradients making it unlikely that base-level fall triggered headward fluvial incision.

In the stable climate background of long-eccentricity minima fluvial incision could have been repeatedly triggered when discharge was enhanced relative to sediment supply when the latter being reduced by an expanding vegetation cover upstream. The alternative downstream control on incision by longitudinal profile flattening during aquifer-eustatic base-level rise is not likely because of the downstream location of the study area. The incision may have sustained until a fluvial equilibrium was autogenically established and/or the stream power was reduced because the valley floor was inundated by backwater-induced groundwater table rise. Continued base-level rise and reduced stream power possibly caused aggradation of mostly fine-grained sediments. The deposition of the fine-grained sediment load in a meandering bayou would then have produced the low-angle inclined mud-rich palaeovalley-fills of the Lebo Shale Member. The model of Figure 3.9 provides a reference for recognizing fluvial aggradation-incision sequences regulated by long-eccentricity forcing.
Figure S3.1. Relation between short- and long-eccentricity in polarity chron C28r in marine sections (Zumaia, site 1262, site 1209) is linked to the aggradation-incision sequences in the Lebo Shale. Short-term eccentricity minima are indicated by green arrows based on lithostratigraphy of Zumaia, Spain (A), core photos and Fe counts for site 1209 (B) and site 1262 (C). D shows the timing of incision based on the Lebo Shale panel. Reversal C29n-C28r occurs in top V-coal followed by few meters of clastic overbank sediment aggradation. Incision occurred therefore somewhere after C29n-C28r. Valley-fill aggradation starts later but these deposits of Facies Association B only occur in C28r.