Late Paleocene-Eocene Syn-Oroegenic Fluvial Sedimentation and Detrital Fission Track Thermochronology of Laramide Syn-Oroegenic Sediments, Denver Basin, CO.

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Abstract

This study investigates the Laramide foreland syn-orogenic sediments preserved in the Denver Basin, CO and interprets the Late Cretaceous-Eocene tectonic and climate history of Central Colorado. This thesis is divided into 3 chapters. Chapter 1 provides an introduction to the Denver Basin and the Front Range, interpreting the stratigraphy to provide support and background for the following chapters. Chapter 2 analyses the Late Paleocene to Eocene fluvial sediments in the Denver Basin and interprets the deposits to be a large distributive fluvial system or megafan which formed during the Paleocene-Eocene Thermal Maximum. Chapter 3 uses detrital apatite and zircon fission track thermochronology to analyze the Front Range and Denver Basin thermal history.
Chapter 1- Introduction and Background on Denver Basin and Front Range, Colorado

Introduction

This study investigates the Laramide foreland syn-orogenic sediments preserved in the Denver Basin, CO and interprets the Late Cretaceous-Eocene tectonic and climate history of Central Colorado. Laramide orogenesis generated thick-skinned block uplifts in the Late Cretaceous through Eocene from Canada to Northern New Mexico, up to 700-1500 km inland from the convergent margin (figure 1.1) (English and Johnston, 2004). The Front Range in Central Colorado, USA is the easternmost extent of Laramide deformation. Sediments were shed from the Front Range and deposited east into the Denver Basin from 70-55 Ma (figure 1.2) (Raynolds et al., 2007). This thesis aims to answer two questions about these Laramide syn-orogenic sediments: 1) what prompted deposition of Late-Paleocene – Eocene fluvial sediments? 2) What is the thermal history of the Front Range and the Denver Basin? This chapter describes and interprets the Denver Basin strata to provide background for sedimentological analysis of late Paleocene-Eocene sediments (chapter 2) and detrital fission-track analysis of Laramide syn-orogenic sediments (chapter 3).

Background

Denver Basin Stratigraphy

The Denver Basin preserves ~4.5 km of sediment overlying Precambrian basement (figure 1.3, 1.4). The lowest sedimentary formations were deposited in the Paleozoic to Mesozoic and a wide variety of lithologies. These formations are overlain by the ~2 km thick Pierre Shale, a deep water marine deposit that represents encroachment of the Cretaceous Western Interior Seaway (Raynolds, 2002). Episodic regressions of the seaway lead to the deposition of the near shore and shoreface Fox Hills Sandstone (~71-69 Ma) (Raynolds, 2002; Raynolds et al., 2007). The Fox Hills is ~70 m thick in the Denver Basin and on the western side of the Front Range, indicating the formation was
deposited prior to Laramide uplift. The top of the Fox Hills Sandstone provides a constraint (<70-69 Ma) on the onset of deformation (figure 1.5) (Raynolds, 2002; Kelley, 2002).

Conformably overlying the Fox Hills Sandstone is the Laramie Formation (~69-68 Ma) (Raynolds, 2002; Raynolds and Johnson, 2003), a package of fine-grained carboniferous shale and coal with limited fluvial sandstone bodies that indicate a swampy coastal plane depositional environment (Raynolds, 2002). Petrographic analysis of the Laramie Formation reveals a significant volcanic component that has been interpreted to represent volcanism associated with the onset of the Laramide orogenesis (Wilson, 2002).

Thickness patterns and sedimentary texture indicate Front Range uplift began during Laramie Formation deposition. The Laramie Formation is thickest (~150 m) and contains more abundant fluvial sandstone bodies on the western side of the Denver Basin (Raynolds, 2002). On the eastern edge of the basin, the Laramie Formation is only ~60 m thick and generally outcrops as overbank facies (Raynolds, 2002). Elevated topography west of the basin explains thicker and more coarse-grained deposition in the western Denver Basin.

An unconformity surface at the top of the Laramie Formation also indicates Front Range deformation began during or directly following Laramie deposition. The Laramie is proximally (west) conformable with the overlying sedimentary formations (Kluth and Nelson, 1988), but it is unconformable distally (east) (Raynolds, 2002), indicating distal Laramie formation was eroded while proximal deposition continued. This sedimentation pattern is observed in underfilled basins, where sedimentation does not fill the available accommodation space resulting in proximal deposition and distal erosion on the peripheral bulge (Figure 1.6) (Catuneanu, 2006; Miall, 1995). Uplift in the Front Range during Laramie Formation deposition caused subsidence, generating accommodation space for proximal deposition. Initial syn-orogenic sedimentation was insufficient to fill basin accommodation space, resulting in underfilled basin conditions, and the erosion of the distal Laramie Formation (Figure 1.7). This unconformity pattern, combined with thickness and sediment texture information indicate Front Range uplift began during the deposition of the Laramide Formation (~70-68 Ma) (Raynolds et al., 2007).
Overlying the generally fine-grained Laramie Formation is ~1 km of coarse-grained arkosic and andesitic syn-orogenic fluvial sediments. These sediments were divided into two sequences (D1 and D2) based on a ~9 Ma unconformity separating the sequences. The base of the D1 sequence (overlying Laramie Formation) contains a coarse-grained, fan-shaped fluvial deposit known as the Arapahoe Conglomerate and Wildcat Mountain Fan (figure 1.8)(Raynolds, 2002; 2004). The <10 m thick Arapahoe Conglomerate and overlying coarse-grained arkosic and andesitic fluvial sandstone deposits have been interpreted to represent a large distributive fluvial system or megafan, based on its large area, widespread distal distribution of coarse grained facies, and radial distribution verging from a paleovalley near the apex at Wildcat Mountain (Raynolds, 2004). The Wildcat Mountain megafan distributed clasts >5 cm diameter over the entire area of the fan at distances >70 km from the topographic front (Raynolds, 2002; 2004). Clasts in the Wildcat Mountain Fan are derived from Mesozoic and Paleozoic sedimentary cover and Precambrian basement rocks, with basement clasts composing ~75% of the population in some locations (Raynolds, 2002).

The presence of basement clasts in the Wildcat Mountain fan indicates portions of the Front Range had ~3 km of sedimentary cover removed by the time of Wildcat Mountain Fan deposition at ~68 Ma (Raynolds, 2002; Raynolds and Johnson, 2003; Raynolds et al., 2007). Therefore ~3 km of exhumation must have occurred after the deposition of the Fox Hills Sandstone (70-69 Ma) and the onset of deposition of the Arapahoe Conglomerate (Figure 1.9) (Raynolds, 2002).

The upper units of the D1 sequence overlying the Wildcat Mountain fan are composed of proximally coarse-grained fluvial deposits that grade distally into abundant fine grained overbank, lignite, and coal deposits (Raynolds, 2002). The D1 sequence was deposited 68 - 64 Ma, and contains proximal progressive unconformities representing deformation associated with continued Front Range uplift and basin subsidence throughout the most of the deposition of the D1 sequence (figure 1.10)(Kluth and Nelson, 1988; Raynolds, 2002). Andesitic grains observed throughout the D1 sequence indicate volcanic activity and complete denudation of volcanic deposits on the Front Range (Raynolds, 2002). Distal fine-grained deposits preserve abundant fossil flora, fauna, and volcanic ash beds that allow for detailed stratigraphic dating of the D1 sequence and
paleoclimate interpretation throughout the sequence (Raynolds, 2002 and references therein).

The D2 sequence unconformably overlies the D1 sequence with ~9 million years of section absent between the sequences (Raynolds, 2002). The nature of this unconformity is unclear, but has been interpreted to represent a period of subdued topography and limited uplift (figure 1.11)(Reynolds, 2002 and references therein). The lack of a paleosol horizon within the upper D1 indicates the unconformity surface was likely erosional.

The D2 sequence differs significantly from the D1 sequence in lithology and sedimentary texture (Reynolds, 2002). In contrast to mostly andesitic sediments and abundant fine grained overbank facies of the D1, the D2 sequence is almost completely arkosic and mostly composed of coarse-grained channel facies. Detailed sedimentological analysis of the D2 sequence in chapter 2 will provide insights into late Paleocene-Eocene climate and tectonic conditions.

Although the duration of D2 sedimentation is unclear, the uppermost D2 represents effective denudation of the Front Range and the formation of the Rocky Mountain Erosion Surface (RMES), a gently eastward sloping erosion surface formed in the Eocene (Leonard and Langford, 1996; Kelley and Chapin, 2004). The RMES is preserved on the tops of buttes and throughout the Front Range and formed before 37 Ma, when the surface was overlain by the Wall Mountain Tuff (figure 1.12) (Reynolds, 2002; Kelley and Chapin, 2004). Modern offset between the erosion surface preserved in the Front Range and in the Denver Basin is estimated to be 90 ± 60 m, indicating a lack of significant uplift in the mountains relative to the basin since the deposition of the D2 sequence (Leonard and Langford, 1996). Post-Eocene tilting, warping, and epiorogenic uplift generated the current topography through differential erosion (Leonard and Langford, 1996).

The full extent of sedimentary activity after the Laramide orogeny is unclear because Cenozoic epiorogenic uplift resulted in the erosion of most rocks overlying the RMES. An exception is the Wall Mountain Tuff (37 Ma), an up to 15 m thick rhyolitic ash flow which was deposited across the RMES and is likely the result of a single catastrophic eruption which occurred west of the Front Range (Raynolds, 2002). The
Castle Rock Conglomerate is an incised valley fluvial deposit which cut into the Wall Mountain Tuff and the upper D2 sequence (Raynolds, 2002). The up to 15 m thick Castle Rock Conglomerate was deposited in paleovalleys in the center of the basin (figure 1.3) contains large angular boulders from the Wall Mountain Tuff indicating multiple breakthrough flood events, likely caused by volcanic flows and mudflows (Raynolds, 2002; Morse 1979). Although other sediments may have been deposited above these formations, no evidence has been found to support significant base level change and sedimentation overlying the Castle Rock Conglomerate and Wall Mountain Tuff in the Denver Basin.
Figure 1.1 – Geologic map showing areas of pre-Laramide and Laramide magmatism. Block Uplifts and general extent of Laramide deformation without restored Cenozoic extension. Figure from English and Johnson (2004).
Figure 1.2 – Topographic map of western U.S. showing the extent of deformation associated with the Laramide orogenesis (red dots). The location of study area, Front Range and Denver Basin, is shown by the red box. Figure Modified from Bunge and Grand (2000).
Figure 1.3 – Geologic Map and Cross Section of the Denver Basin. Modified from figures courtesy of Dr. Ian Miller and the Denver Museum of Nature and Science.
Figure 1.4 – Stratigraphic column of Denver Basin. The Laramide Orogeny likely began during deposition of the Laramie Formation and continued through deposition of the D2 sequence. The top of the D2 sequence forms the Rocky Mountain Erosion Surface (RMES). Modified from figure courtesy of Ian Miller and Denver Museum of Nature and Science (2009).
Figure 1.5 – Pre-Laramide Stratigraphy. Fox Hills Sandstone is observed in equal thickness on to the East and West of the Front Rang indicating it was deposited prior to Laramide uplift. Laramide uplift likely began sometime during the deposition of the Laramie Formation.
Figure 1.6 – Schematic representation of underfilled basin conditions. Orogenic loading causes subsidence and flexural uplift. Sedimentation rates are less than subsidence rates generating a topographic high on the forebulge where erosion takes place forming an unconformity in the rock record. Figure from Catuneanu (2006).
Earliest Laramide Orogenesis and the deposition of Laramie Formation

Figure 1.7 – Deposition of Laramie formation and Early Laramide orogenesis. Distal unconformity at top of Laramie formation indicates underfilled basin conditions with distal erosion and proximal deposition. Laramie was likely sourced from the erosion of pre-Laramide stratigraphy, such as the Fox Hills Sandstone, Pierre Shale, and Paleozoic and Mesozoic Formations (Wilson, 2002).
Figure 1.8 – Isopach Map of Wildcat Mountain Fan (Arapahoe Conglomerate and overlying coarse grained fluvial sediments at the base of the D1 sequence). Figure from Raynolds (2004).
Figure 1.9 — Denver Basin during the deposition of the Wildcat Mountain Megafan. Presence of basement clasts within the Arapahoe Conglomerate indicates ~3 km of exhumation occurred before the onset of D1 deposition.
Denver Basin during the Deposition of D1 Sequence 67.5-64 Ma

Figure 1.10 – Denver Basin in the Late Cretaceous to Early Paleocene (67-64 Ma). Abundant volcanic activity (ash falls in D1) and andesitic detritus indicates volcanic eruption and complete denudation of volcanic rocks in the Front Range.
Denver Basin during the Unconformity separating the D1-D2 sequences.

Figure 1.11 – Denver Basin during the unconformity (~9 Ma) between the D1 and D2 sequences. Relatively flat surface with subdued topography likely existed during this time of non-deposition or erosion in the basin.
Figure 1.12 – Regional map of the Rocky Mountain Erosion Surface. Surface was likely formed in the Eocene and represents the relatively flat erosion surface which formed from the top of the D2 sequence and the Front Range. Figure from Kelley and Chapin (2004).
Chapter 2 – Paleocene-Eocene Laramide Syn-orogenic Fluvial Sedimentation

**Introduction**

Herein, the D2 sequence is interpreted to have been deposited by two pulses of sedimentation. The first pulse (~58-55 Ma) was caused by base level change, possibly due to uplift in the Front Range and resulted in the deposition of proximal coarse-grained facies and the formation of a widespread paleosol horizon. The second pulse of sedimentation resulted in the widespread distal deposition of coarse-grained facies and is interpreted to have been deposited by one or more large distributive fluvial systems or fluvial megafan(s). The progradation of the coarse facies coupled with a dramatic increase in sedimentation rates indicates the second pulse may have been caused by extreme climate conditions associated with the Paleocene-Eocene Thermal Maximum (~55 Ma). This interpretation is supported by stable isotope analysis which identifies the negative carbon isotope excursion associated with the Paleocene-Eocene Thermal Maximum throughout the second pulse of D2 sedimentation.

**Background**

**Distributive Fluvial Systems**

Distributive fluvial systems (DFS) are aggrading fluvial systems that deposit sediment in a radial, fan-shaped pattern (Weissmann et al., 2021; Hartley et al., 2010). DFS is a broad classification which encompasses systems producing alluvial fans, fluvial fans, and megafans, representing geomorphic features ranging in length from a <1 km to >700 km (Weissmann et al., 2010; Hartley et al., 2010). Most sediment deposition in modern continental fluvial systems occurs in DFS, and the vast majority (possibly >90%) of continental fluvial sediments preserved in the rock record were deposited by DFS (Weissmann et al, 2010).

DFS are distinguished from other continental fluvial systems (e.g., axial river) by their fan shaped, radial plan view depositional pattern which is convex upwards across the fan and concave upward down fan (Weissmann et al, 2010; Hartley et al., 2010). Radial deposits are formed because DFS rivers confined or held fixed in a valley at an
intersection point or apex, but unconfined downstream, where the river is free to bifurcate or migrate laterally, sometimes up to \( \sim 180^\circ \), across a low-relief alluvial basin (Leier et al., 2005; DeCelles and Cavazza, 1999; Weissmann et al, 2010; Hartley et al., 2010).

Fluvial channel size in DFS generally decreases down fan and sediment grain sizes fine distally (increased flood plane area/channel area ratio down fan) (Weissmann et al, 2010; Hartley et al., 2010). DFS terminate in a variety of ways, such as into an axial river, dune field, contributory fluvial system, Playa Lake, or wetland (Weissmann et al, 2010; Hartley et al., 2010). A range of alluvial processes (confined or unconfined flow) occur in DFS, which are found in most climate and tectonic settings, generating a large variety of fluvial sedimentary deposits (Weissmann et al, 2010; Hartley et al., 2010).

**Megafan Sedimentation**

Megafans are the largest classification of DFS, distinguished by their large area \( (\sim 10^3-10^5 \text{ km}^2) \), length (typically \( >30-100 \text{ km} \)), and low gradient \( (\sim 10^{-4} \text{ to } 10^{-6}) \) (Leier et al., 2005; Hartley et al., 2010). Generally in DFS, length is inversely proportional to gradient, such that longer DFS have a lower gradient (Satio and Oguchi, 2005; Hartley et al., 2010). Therefore, the size of the system (length / depositional area) and the resulting distribution of coarse grain material distal to the topographic front are the key identifiers of paleo-megafans in the rock record.

Megafans form in specific conditions which allow DFS to grow to large areas (Hartley et al., 2010). Controls on the length of DFS include horizontal accommodation space, discharge, and sediment supply (Hartley et al., 2010). Accommodation space can be generated through tectonic activity such as orogenesis or subsidence, and the largest megafans occur in unconfined foreland basins and cratonic settings due to the effectively infinite horizontal accommodation space (Hartley et al., 2010). Discharge determines flow distance and sediment transport capacity of a fluvial system and is controlled by climate (volume of precipitation) or tectonic factors (integration of drainage networks generating large catchment area) (Horton and DeCelles, 2001; Hartley et al., 2010). Flashy discharge causes episodic base level changes, which prompt incision into proximal fluvial deposits, shifting deposition basinward away from the topographic front increasing the length of the DFS (Hartley et al., 2010; Leier et al., 2005). A lack of sediment generation in the system’s source caused by either tectonic (e.g. lack of
topography) or climate (e.g. low discharge, low weathering/erosion) factors will restrict the size of the fan (Hartley et al., 2010). Megafans occur in most climatic and tectonic settings, but it has been argued that megafans are more likely to form in seasonal or monsoonal, sub-tropical climates where flashy discharge increases the size of the system (Leier et al., 2005).

**Paleocene-Eocene Thermal Maximum**

The Paleocene-Eocene Thermal Maximum (PETM) is a dramatic global warming event which occurred at the Paleocene-Eocene boundary (~55.5 Ma) and lasted for ~100 Ky (Zachos et al., 2005). The event is hypothesized to have been caused by an increase in atmospheric greenhouse gases caused by the dissociation of oceanic methane hydrates (Dickens et al., 1997). Isotopically depleted Methane gas released into the atmosphere caused a (~3%) negative carbon isotope excursion (CIE) (Zachos et al., 2005). This CIE is observed in the rock record as a rapid negative shift in $\delta^{13}C$, which persists for the duration of the PETM (~100 Ky), and then returns to pre-PETM $\delta^{13}C$ values over (~60 Ky) (Zachos et al., 2005).

Global Temperature generally increased during the PETM, but changes in precipitation patterns as a result of this warming event are unclear. Models and geologic evidence suggest intensified precipitation during the PETM on a global scale (e.g. Sloan et al., 1999, Robert and Kennett, 1994). Some regional studies suggest arid to semi-arid conditions (e.g. Schmitz and Pujalte, 2003), while other suggest humid conditions (Adatte et al., 2000). Kraus and Riggens (2007) conducted a detailed study of PETM stratigraphic sections in Northern Wyoming concluding processional wet/dry cycles occurred throughout the PETM, with arid, low precipitation conditions at the onset of the PETM, generally becoming more humid towards the end of the PETM.

**Methods**

The D2 sequence was studied through field reconnaissance and a literature review. Large exposures in the Denver Basin are limited (e.g. Dawson Butte) and therefore interpretations are based on numerous smaller outcrops from throughout around the basin. Near-complete sections were studied at Dawson's Butte (39°28'21.53"N,
104°56'08.46"W), Daniels Park (39°00'54.8"N, 104°16'07.58"W), and the Calhan Paint Mines (39°28'21.53"N, 104°55'23.15"W). These field locations were selected by reviewing detailed surface facies maps (Mayberry and Lindvall, 1972; 1977), published field studies (Raynolds, 2002, Morse 1979), and in consultation with the staff (primarily Dr. Ian Miller) at the Denver Museum of Nature and Science.

The goal of examining outcrops of the D2 sequence was to determine the depositional environment by examining sedimentary characteristics. This project aims to build upon the observations of previous workers (e.g., Morse, 1979, Solister and Tsudny, 1978; Raynolds 2002) by examining spatial changes in the sedimentary facies and the distribution of coarse grained channel deposits. Where possible, paleocurrent measurements were made from trough-cross beds in the D2 sequence. 3-5 measurements were made at 3 field locations and mean directions were determined. These results were combined with more extensive paleocurrent measurements from 13 locations collected by Morse (1979). Morse (1979) analyzed the uppermost D1 and D2 sequence, but did not distinguish between the two unconformity bounded sequences. Each field location in Morse (1979) was analyzed based on maps and descriptions and included only if the surface geology of the location was determined to be completely D2.

**Description and Interpretation of D2 sequence**

This section offers a detailed description of the D2 sediments and an interpretation of its depositional environment. Herein, the D2 sequence is interpreted to have been deposited by one or more fluvial megafan(s) in response to extreme climate conditions associated with the Paleocene-Eocene Thermal Maximum (PETM). Firstly, the D2 deposits are described and interpreted to have resulted from a rapidly avulsing braided river system. Secondly, the isopach, paleocurrent, and sedimentary texture are analyzed and the sequence is interpreted to have been deposited by one or more fluvial megafans. Thirdly, the timing of the deposition of the D2 sequence is investigated to reveal the braided fluvial deposits prograded away from the topographic front during D2 deposition. Finally, the progradation events are associated with the Paleocene-Eocene Thermal Maximum, indicating the climate event caused a dramatic increase in
sedimentation rates, forming one or more fluvial megafan(s) and depositing coarse grained sediments abnormally far away from the topographic front (~60 km).

**D2 Facies Associations and Paleosols**

Two distinct facies associations are observed throughout the D2 sequence: channel and overbank facies associations. Channel facies association dominates the proximal D2 stratigraphy (up to ~20 km from Front Range) with the overbank facies becoming more abundant in the distal (greater than 20 km) stratigraphy.

*Channel Facies Association*

Two distinct facies exits in the channel facies association: 1) massive pebble conglomerate to massive coarse-grained sandstone, and 2) trough-cross bedded, coarse-grained sandstone (Morse, 1979). The massive facies generally fines-upward over 1 m increments with grain sizes ranging from cobbles to coarse sand, and preserves rip-up clasts, petrified logs, and large (>10 cm diameter) granitic clasts (figure 1.10). Trough-cross-bedded facies is generally composed of cross-stratified coarse sand that contains basal ripple marks in sections where this facies overlies overbank facies associations (figure 1.11).

The two facies are observed in complex cross-cutting relationships with facies changes vertically and laterally. Coarse channel lag observed in the massive facies, and truncations to the cross-bedded facies allow for the identification discrete channels in outcrop, which are generally ~1 m deep (figure 1.12).

*Description of Overbank Facies Association*

The overbank facies association is composed of two facies, fine-grained flood plain and crevasse splay deposits. The fine grained flood plain deposits are best preserved distally as olive-brown sand and silt interbedded with channel facies association (Reynolds, 2002). Overbank sediments are preserved as mud lenses, discontinuous beds, and rip-up clasts within the channel facies association (figure 1.11 and 1.12). Paleosols formed in the overbank facies, with a distinctive, thick paleosol at the base of the D2 sequence (Reynolds, 2002; Farnham and Kraus, 2002). Crevasse splay deposits are
distinguished from the channel facies association by smaller grain size (medium to fine sand) and thin (~5 cm) horizontal bedding (figure 1.13).

**D2 Paleosols**

Paleosols are fossil soils formed from the breakdown of minerals at the surface (Farnham and Kraus, 2002). Various paleosol horizons are preserved throughout the D2 sequence, with a widespread, well-developed paleosol surface found at the base of the D2 sequence (Farnham and Kraus, 2002). Paleosols are generally found in the overbank facies association, but proximally occur in the channel facies association (figure 1.14). D2 paleosols vary from thin (<1 m) orange or gray clay rich zones to thick (>10 m) bright red, purple, orange, and yellow zones with abundant root casts, slickenslides, carbonate nodules, and fossil plant material (Farnham and Kraus, 2002).

A distinctive, widespread paleosol at the base of the D2 sequence is aggradational and was formed during the deposition of the D2 sequence rather than during the 9 My year unconformity separating the D1 and D2 (Farnham and Kraus, 2002). The paleosol is thickest in the distal portions of the D2 sequence, thinning and ultimately disappearing to the west (Raynolds, 2002).

**Interpretation**

The D2 sediments are braided stream and flood plain deposits (Morse, 1979) indicative of a rapidly avulsing fluvial system. Complex vertical and horizontal channel stacking patterns and cross-cutting facies relationships indicate aggradational channel abandonment. Abundant fine-grained rip-up clasts and scant preservation of overbank facies indicate avulsions resulted in incision into the fine grained overbank deposits.

The basal paleosol was formed during the initial deposition of the D2 sequence (Farnham and Kraus, 2002). The paleosol is thin and weakly developed in the proximal part of the basin indicating less weathering and likely higher sedimentation rates close to the topographic front during the formation of this horizon. Slower sedimentation rates in the distal portion of the basin generated a more strongly developed basal paleosol.

**Distribution of D2**
**Description**

The complete depositional thickness and distribution of the D2 sequence is unclear because Neogene epiorogenic uplift stimulated the erosion of much of the upper D2 sequence. Proximally the thickness D2 was determined by projecting the RMES preserved on buttes (figure 1.15) (Raynolds, 2002). The isopach map (Reynolds, 2002) combined with paleocurrent data (Morse, 1979) indicates a generally fan shaped depositional pattern, with the greatest thickness observed just north of Colorado Springs with transverse paleocurrent (figure 1.15).

The D2 sequence distributes coarse grained sediments throughout the sequence over an area >4000 km² and more than 60 km from the topographic front. Although the sediment grain size decreases away from the Front Range, pebble conglomerates and gravels are abundant at localities ~60 km into the basin, giving the D2 sequence a dramatically coarser texture than the D1 sequence.

**Interpretation**

The D2 sequence is a widespread coarse-grained braided stream deposit which was deposited by one or more fluvial megafan(s). Although paleocurrent data does not show a clear radial distribution, they do indicate paleoflow was generally transverse to the range. Transverse fluvial deposits are formed by distributive fluvial systems (Weissmann, 2010), and therefore the D2 sequence was deposited by one or more DSF. The large area and coarse sedimentary texture (abundant coarse-grained facies observed >60 km from topographic front) indicate D2 sequence was deposited by a fluvial megafan which allowed for the widespread transverse and tangential distribution of coarse-grained facies.

**Constraints on Timing of D2 Deposition**

**Description**

The D2 sequence was deposited in the Late Paleocene to Early Eocene (58-55 Ma) based on a variety of age constraints (Obradovich, 2002). Fossil pollen and leaves were analyzed from a variety of localities around the basin (figure 1.16) (Nichols and
Flemmings, 2002; Solister and Tschudy, 1978: Johnson et al., 2003). Fossil pollen assemblages from above the paleosol in the distal portion of the basin and from within the D2 in the Kiowa core yield ages of Early Eocene (Nichols and Flemmings, 2002; Solister and Tschudy, 1978). Samples from surface localities in the proximal D2 and samples from within the Castle Pines core were devoid of Early Eocene flora, containing only pollen or late Paleocene age (P6) (Nichols and Flemmings, 2002; Solister and Tsudny, 1978).

Four fossil leaf localities were analyzed in the Denver Basin, but only one in the distal portion of the basin could be dated to be early Eocene (Johnson et al, 2003). Fossil leaf localities in the proximal part of the basin near castle pines well were not adequately preserved to provide age constraints (Johnson et al., 2003).

Magnetostratigraphic analysis of the Kiowa and Castle Pines cored wells found reversed polarity in the Kiowa well and a normal, reversed, normal pattern in the Castle pines well (Hicks et al., 2002). Abundant early Eocene pollen in the Kiowa core (Nichols and Flemmings, 2002) indicates the strata was deposited during the C24r interval, spanning the late Paleocene – middle Eocene (Hicks et al., 2002). The absence of Eocene pollen but presence of late Paleocene pollen in the castle pines core indicates section may correlate to the C26n, C25r, and C25n intervals, implying depositional ages of ~58-56ma.

*Interpretation*

Age constraints on the D2 suggest the sequence was deposited from the Late Paleocene to Early Eocene. Base Level change, possibly associated with uplift in the Front Range, in the Late Paleocene (~58 Ma) prompted the onset of D2 sedimentation over the D1 unconformity surface. Most initial D2 sedimentation occurred proximally (braided rivers and alluvial fans) and very slow distal deposition generated the thick, well-developed, aggradational basal paleosol. Distal progradation of the D2 sequence through the early Eocene led to the deposition of younger early Eocene sediments over the widespread paleosol horizon.

*Stable Isotope Analysis*
Fricke (unpublished data) conducted stable isotope analysis of sections of the D2 sequence, identifying a negative isotope excursion in the distal D2 sequence. Samples were collected from the Kiowa cored well identify a carbon isotope excursion (CIE) interpreted to be associated with the Paleocene-Eocene Thermal Maximum (PETM). In the lower part of the paleosol sequence, $\delta^{13}C$ values generally range from $\sim$ -23 to -25 $\%_o$ (Fricke, unpublished). These $\delta^{13}C$ values are typical for non-PETM sediments from other intermontane basins such as the Bighorn basin in Wyoming (Kraus and Riggens, 2007)(Frick, unpublished).

The top of the paleosol interval marks a rapid shift to significantly more negative $\delta^{13}C$ values which continues throughout the overlying 200 ft of core. These negative values are similar to PETM sediments observed in the Bighorn Basin (Kraus and Riggens, 2007). When combined with the other age constraints (pollen, magnetostratigraphy), one can conclude the negative isotope excursion is associated with the $\sim$100-150 ky PETM.

**Discussion**

Analysis of the D2 sequence revealed the sediments are braided fluvial sandstone deposits which were deposited by one or more fluvial megafans. Deposition occurred in two pulses (figure 1.18). The first pulse occurred from $\sim$58-55.5 Ma forming proximal coarse-grained alluvial and fluvial fan deposits and the widespread D2 basal paleosol with sedimentation rates on the order of $\sim$100 ft / My proximally. Isotope data (Frickie and Picard, 2003-2004; Snell, 2007) suggests the second pulse was deposited during the PETM and is characterized by a dramatic increase in sedimentation rates ($\sim$150 ft of fluvial deposits in $\sim$150 ky), progradation of the coarse grained facies, and the formation of a fluvial megafan. This shift in sedimentation patterns during the PETM suggests the second pulse was caused by PETM climate change which resulted in flashy fluvial discharge and/or increased erosion rates in the Front Range.

One possible cause of the PETM pulse of D2 sedimentation suggests climate change (global warming and increased seasonality) resulting in flashy discharge which increased sedimentation rates and caused progradation of the fluvial system. Analysis of PETM sediments in the Bighorn Basin, WY concluded the regional PETM climate was
characterized by wet/dry cycles (Kraus and Riggins, 2007), which may have been similar to seasonal or monsoonal climates observed today in the Himalaya foreland basin. The formation of a fluvial megafan during a period of increased seasonality supports the hypothesis that megafans are more likely to form in subtropical latitudes with seasonal or monsoonal climates (Leier et al., 2005). Flashy discharge in seasonal/monsoonal climates promotes abundant and frequent major avulsion events which allow the size of DFS to grow large, sometimes into fluvial megafans (Leier et al., 2005).

An alternative explanation would be that increased erosion rates in the Front Range during the PETM caused orogenic unloading which prompted progradation of the D2 sequence. Orogenic unloading generally occurs at the end of orogenesis, and is caused by the net removal of material from the mountain range resulting in isostatic rebound in the mountains and in the proximal basin and flexural subsidence in the distal basin (figure 1.19) (Miall, 1996; Catuneanu, 2006). This causes the fluvial system to erode or bypass the proximal sedimentary deposits and sedimentation occurs in the distal basin where flexural subsidence generated accommodation space (Miall, 1996; Catuneanu, 2006). The D2 deposits are the last Laramide syn-orogenic sediments therefore one would expect some orogenic unloading to have occurred during D2 deposition regardless of the climate, but the dramatic increase in sedimentation rates associated with the PETM pulse suggests this process was amplified by PETM climate changes. Both increased erosion in the Front Range and flashy discharge caused by PETM climate change likely contributed to the second pulse of D2 sedimentation.

Fluvial megafan formation also occurred at the base of PETM sediments in the Claret Conglomerate in a Pyrenees foreland basin, Spain (Schmitz and Pujalte, 2007). The existence of megafan deposits in other PETM sediments suggests flashy discharge conditions were widespread, possibly even global. However a widespread coarse-grained fluvial deposit was not observed in the Bighorn Basin PETM sediments suggesting megafan formation did not occur in all foreland basins during the PETM.

Climate varies with latitude and Leier et al. (2005) noted modern megafans exist between 15° and 35°, in latitudes more prone to have seasonal/monsoonal climates. The PETM may have caused a shift in this latitudinal seasonal climate belt, causing the formation of megafans in the Pyrenees and Front Range foreland basin. Other
continental foreland basins which preserved PETM sediments and resided in the latitudinal seasonal climate belt during the PETM, may also contain widespread coarse-grained fluvial deposits.

**Conclusion**

The D2 sequence was deposited by a fluvial megafan during the Paleocene-Eocene Thermal Maximum. Base level change in the Denver Basin in the late Paleocene (~58 Ma) prompted D2 sedimentation, characterized by proximal coarse grained deposition, distal fine grained deposition, and the formation of a widespread paleosol horizon consistent with small to medium size DFS. The PETM climate event caused the formation of a fluvial megafan by increasing erosion rates and causing flashy fluvial discharge. PETM megafan formation in Pyrenees and Front Range foreland basins suggests a shift in the latitudinal seasonal climate belt during the PETM, indicating widespread coarse grained fluvial deposits may exist in PETM sections in other foreland basins.
Figure 1.10 – Pictures of the massive coarse grained facies of the channel facies association. (Left top) petrified log preserved in a coarse sand channel deposit. (Left bottom) 10 cm diameter granitic clast in coarse sand to granule channel deposit. (Right) massive pebble conglomerate overlying the basal paleosol at Calhan paint mines. Pebble conglomerate is found in abundance in channel deposits ~60 km from the topographic front.
Figure 1.11 - Outcrop of D2 at Daniels Park. Shows through-cross bedded facies (cross beds in green), massive facies, fine-grained overbank facies, rip up clasts, and crevasse splay deposit.
Figure 1.12 – (top left) small ~25 cm fining upward sections within a pebble conglomerate (top right) coarse channel lag on the base of a channel deposit. (bottom) paleo-channel deposits outlined in red, fine grained rip up clasts weather out to forming holes in rock face.
Figure 1.13 – Thinly bedded fine and medium sand with shaley interbeds, interpreted to represent crevasse splay deposit. Crevasse splay deposits are overlain by conglomerate channel fill deposit.
Figure 1.14 – Paleosol horizons in the D2 sequence. (top left) Paleosol horizon forming in the proximal D2 sequence, formed in coarse sand channel fill deposit (Dawson Butte). (top right) well developed basal paleosol found at the base of the D2 sequence (Calhan Paint Mines). (bottom) large root casts in a paleosol horizon which formed on a channel fill deposit.
Figure 1.15 – Isopach and paleocurrent from the D2 sequence. Paleocurrent indicates the D2 Rivers flowed to the east. The D2 was deposited by transverst distributive fluvial systems. Isopach indicates fan shaped deposition of the D2. Greatest thickness in the southern portion may be due to uplift in the pikes peak region of the Front Range (Raynolds, 2002). Isopach from Raynolds, 2002. Paleocurrent from Morse, 1988.
Figure 1.16 - Fossil pollen and leaf age constraints on the D2 Sequence. Fossil pollen data indicates the proximal sediments are late-Paleocene in age and the distal sediments are early-Eocene.
Figure 1.18 – Depositional model for the D2 sequence. Onset of D2 sedimentation (~58 ma) was prompted by uplift in the Front Range. Throughout the Late Paleocene sedimentation rates were low in the distal portion of the basin producing the widespread thick basal paleosol horizon. Climate change associated with the PETM prompted progradation of the coarse facies by dramatically increasing sedimentation rates.
Figure 1.19 – Figure demonstrating orogenic unloading. Removal of material in the mountain range leads to isostatic rebound causing erosion or bypass of proximal sedimentary deposits and deposition into the foresag. Figure from Catuneanu (2006).
Chapter 3 – Fission-Track Analysis of the Denver Basin

Introduction

Detrital apatite and zircon fission-track analysis was conducted on Laramide synorogenic sediments in the Denver Basin with the goal of constraining the Late Cretaceous to Eocene exhumation of the Front Range. Samples were collected of granitic clasts and sandstones with the goal of detecting Laramide exhumation cooling and determining the effects of volcanic inputs into the basin on fission track data. These results are combined with apatite and zircon fission track analysis of samples collected from the Castle Pines and Kiowa well (Kelley, 2002) to provide a basin wide thermal history.

Background

Fission-track Analysis

Fission-track (FT) thermochronology determines a minerals thermal history by measuring accumulations of ionization damage (fission-tracks) caused by spontaneous nuclear fission in the decay of $^{238}$U (Bernet and Spiegel, 2004; Reiners and Brandon, 2006). The density of fission-tracks in a crystal lattice (number of fission-tracks / area) is dependent on the uranium content of the mineral and the time over which the tracks accumulated (Bernet and Spiegel, 2004; Reiners and Brandon, 2006). Therefore the optically determined track density is compared to the mineral’s uranium content to determine FT age (Bernet and Spiegel, 2004; Reiners and Brandon, 2006). At elevated temperatures, the crystal lattice is repaired (annealed), and fission-tracks are erased.

Annealing of the crystal lattice occurs over a range of temperatures called the partial annealing zone (PAZ) (Reiners and Brandon, 2006). This range of temperatures depends on chemical composition and mineralogy of a crystal, meaning different minerals and minerals of different chemical compositions will have unique PAZs (Reiners and Brandon, 2006). For example, most apatite compositions have a PAZ between ~70-140°C (Reiners and Brandon, 2006). At temperatures above 140 °C, the rate of annealing is much greater than the rate of track generation such that the observed FT density will be effectively zero. At temperatures below 70°C, the rate of annealing is effectively zero and all fission-tracks are preserved. For the temperatures 70-140 °C, the
rate of annealing approximates the rate of generation and some of the fission-tracks are annealed (Reiners and Brandon, 2006). Since FT ages are determined using track density, minerals at temperatures greater than the PAZ will have a FT age of zero, minerals at temperatures below the PAZ will have an age reflective of the minerals last cooling event, and minerals in the PAZ will have an age greater than zero but (figure 2.1) (Reiners and Brandon, 2006).

FT analysis of a mineral yields a single measured time, known as the effective closure temperature (Reiners and Brandon, 2006). As a mineral is cooled, annealing occurs throughout the PAZ, and the effective closure temperature measured in FT analysis is a function of how long the mineral was in the PAZ (figure 2.1) (Reiners and Brandon, 2006). Faster cooling rates yield a higher closure temperature and slower cooling rates yield a lower closure temperature (Reiners and Brandon, 2006).

Fission-track analysis is used in orogenic belts to determine exhumation rates (Reiners and Brandon, 2006). Tectonic uplift and associated denudation provides a mechanism by which rocks from below the PAZ can be exhumed and cooled (Reiners and Brandon, 2006). FT ages from surface rock in a mountain belt represent the time since that rock was at temperatures greater than the partial annealing zone and can be used to determine exhumation rates (Reiners and Brandon, 2006). The analysis of bedrock in mountains is limited in that it can only determine ages for rocks currently exposed at the surface, whereas analysis of detrital samples provides information about rocks eroded from the (Reiners and Brandon, 2006).

Detrital Fission-track Thermochronology

Detrital fission-track thermochronology examines mineral grains preserved in sedimentary rocks to determine exhumation rates and sediment provenance throughout the history of the orogen (Reiners and Brandon, 2006). Detrital thermochronology offers several advantages over bedrock thermochronology (Reiners and Brandon, 2006). Sedimentary rocks contain minerals eroded from a mountain throughout its development and provide a long-term exhumation history of the orogen (Reiners and Brandon, 2006). Detrital thermochronology can also be used to determine sediment provenance throughout the history of the orogeny, and detrital samples often contain grains from
many rocks throughout the mountain range, allowing for the sampling of a larger area (Reiners and Brandon, 2006).

Given a simple basin thermal history in which samples have not been reheated to annealing temperatures, detrital FT ages reflect cooling and exhumation in the source rock (Reiners and Brandon, 2006). This technique is most effective in basins which have well constrained depositional ages throughout the stratigraphy, as it provides a base line to which the FT ages are compared (Garver et al, 1999).

One major difference between detrital and bedrock thermochronology is detrital samples contain grains with a range of ages (Bernet and Garver, 2005). Detrital samples are generally derived from sources throughout the mountain range and differential erosion rates throughout orogen produce a range of FT ages. This range of ages generally shows a distinctive cluster behavior, consistent with the local and regional geology (Garver et al, 1999; Bernet and Garver, 2005). Generally about 100 mineral grains per detrital sample are necessary to produce a histogram which that can be analyzed statistically to identify distinct peaks with a reasonable certainty (Bernet and Garver, 2005).

An analysis of a range of stratigraphic intervals produces a long-term erosion history and documents how the erosion rate changed over time. Lag time is inversely related to erosion rate, such that more rapid erosion will yield a smaller lag time, and slower erosion will yield a large lag time (figure 2.3). Erosion rates vary in tectonically active mountain belts generally from .05 to 10 mm/yr (Brandon and Reiners, 2006).

Detrital thermochronology can be complicated by various geologic events. In regions with active volcanism, input of volcanic grains into basins produces young volcanic age peaks, which have no relation to exhumation in the orogenic belt. Thermal event a major advantage of detrital thermochronology is the ability to observe FT ages throughout the history of the orogeny by sampling throughout the stratigraphy (Bernet and Garver, 2005). Exhumation rate trends can be detected by determining lag time, the FT age of a sample minus the depositional age, which represents the time since a mineral was exhumed past base the PAZ to the time when it was deposited in the basin (Bernet and Garver, 2005).
The time sediment is mobile (time from when rock parcel is eroded to when it is deposited in the basin) is generally less than 1 m.y. and insignificant for the purposes of FT analysis (Reiners and Brandon, 2006 and references therein). However detrital minerals can be stored and recycled producing a lag time which is longer than if the sediments had not been reworked. With the assumption sediment reworking is minimal, lag time, combined with the depth to the base of the PAZ, approximates erosion rates in the source terrane (Bernet and Garver, 2005). However erosion causes the upward advection of heat, increasing the geothermal gradient and changes in the exhumation rates can complicate estimations of geothermal gradients (Rahl et al, 2007; Reiners and Brandon, 2006). Numerical models are necessary in complicated systems, especially with fast erosion rates, to make accurate measurements of exhumation rates (Rahl et al., 2007).

Thermal events in the basin can cause partial annealing of mineral grains, reducing the FT age. Annealing can be caused by burial to a depth within the PAZ or by hydrothermal heating caused by volcanism or upward fluid migration. An accurate estimation of paleo-geothermal gradients, both in the basin and in the mountain belt, and sample burial depth in the basin is essential to determining the potential effects of annealing events.

**Front Range Bedrock Thermochronology**

Bedrock thermochronology provides a useful baseline to which detrital FT results can be compared. Front Range apatite FT thermochronology is complicated by various thermal events, namely volcanism associated with the Rio Grande Rift, (west of Front Range) which have partially and completely reset apatite grains in areas of the Front Range (Kelley and Chapin, 2004). The Oligocene Cripple Creek volcanic center, just west of Pikes Peak, caused the complete annealing of apatite grains in the granitic basement rock (Kelley and Chapin, 2004). Volcanic activity caused apatite grains, up to 1 km from the intrusion, to be reset to an age of \(\approx 27 \pm 2\) Ma (Kelley and Chapin, 2004).

Many areas of the Front Range contain grains which were not thermally reset yielding apatite FT ages ranging from 64-600 Ma, representing rocks from both below and above the base of the Late Cretaceous apatite PAZ (Kelley and Chapin, 2004).
Preservation of the PAZ allows for an estimation of total uplift and denudation in the Front Range (Kelley and Chapin, 2004). Total surface uplift was estimated to be 2.6 km based on the Fox Hills Sandstone, which was located at approximately sea level in the Denver Basin prior to Laramide deformation, and now resides at a much greater elevation due to regional uplift (chapter 1)(Kelley and Chapin, 2004). The apatite PAZ is preserved on Pikes Peak and indicates uplift since late Cretaceous was around 5.9 km (Kelley, 2002). Therefore, between 3.9 and 4.9 km of rock were exhumed in the Front Range by Laramide deformation (Kelley and Chapin, 2004).

**Methods**

This section outlines the methods and procedures used in this study to generate apatite and zircon fission-track ages. Some mineral separation procedures were conducted at Washington and Lee University and FT analysis was conducted at the Grenoble Fission-track Lab by Dr. Matthias Bernet.

Samples were collected from a variety of locations around the Denver Basin and include both sandstone samples and granitic clasts (figure 2.4). Samples were crushed and sieved to collect grain sizes between 25 and 100 μm and sent to the Grenoble Fission-track Lab where apatite and zircon grains were separated using standard magnetic and heavy-liquid separation techniques.

Zircon and apatite grains were mounted in TAP Teflon and epoxy respectively, polished to reveal the interior of the grains, and attached to a muscovite mica detector. Zircons were etched for 18 hours in nitric acid and apatite grains were etched in potassium oxide for 30 seconds to reveal fission-tracks. Grains were irradiated in a nuclear reactor to induce fission decay in the grains and in the mica detector. After irradiation micas were removed and etched for 10 seconds in to reveal induced fission-tracks. All samples were counted on an optical microscope at 1250x dry (100x objective, 1.25 tube factor, 10 oculars) using a zeta of 347.22 ± 3.40 (±1 SE. Binomial peak-fit ages were determined for each sample.

Local geothermal gradient and thermal conductivity was determined from data collected from the Kiowa and Castle Pines wells (Kelley and Blackwell, 2002). The base of the apatite PAZ (~140°C) was likely at a depth of 3 km, roughly at the top of the
granitic basement rocks (Kelley and Blackwell, 2002; Kelley, 2002). The base of the zircon PAZ (~240°C) likely occurred at a depth of ~8 km (Kelley and Blackwell, 2002; Kelley, 2002).

Results

Of the 19 samples collected, 2 contained apatite grains and 10 contained ample zircon grains for fission-track analysis. Zircon FT results indicate significant volcanic input at all stratigraphic intervals and possibly post-depositional partial annealing in the southern part of the basin. Apatite FT results indicate annealing may have occurred after deposition throughout the basin.

Zircon Fission-track Results

Datable zircon grains range in age from 34.1-1637 Ma (figure 2.5). Most of the grains collected (~75%) were metamict and therefore unable to be dated. Ideally about 100 grains are needed per sample, but none of the samples yielded close to 100 datable grains. Bedrock AFT thermochronology indicates the zircon PAZ was not exhumed in the Front Range (Kelley and Chapin, 2004) and therefore ZFT ages cannot be related to Laramide exhumation.

Detrital zircon FT cooling ages fall into 6 populations: Paleocene-Eocene (P1), Late Cretaceous (P2), Early Cretaceous (P3), Late Triassic (P4), Paleozoic (P5), and Precambrian (P6) (figure 2.5). Results from this study fall into similar age populations as the results from zircon FT analysis on the Kiowa and Castle Pines Wells (Kelley, 2002)(Figure 2.6). P6 ages represent Precambrian cooling of the granitic basement rocks. P5 – P3 ages are from pre-Laramide cooling events, such as volcanic eruptions or the Ancestral Rockies orogenesis. P2 ages are indicative of recycled volcanic grains from ash falls preserved in the Pierre Shale and Fox Hills Sandstone. P2 ages in many samples are the same as depositional ages, indicating grains were introduced by syn-orogenic volcanic activity. Throughout the Laramide Orogeny, there was significant regional volcanic activity (Reynolds, 2002), and volcanic grain populations are consistent with the results of petrographic analysis (Wilson, 2002).

P1 cooling ages are generally found in the D2 sequence (depositional age 58-55 Ma). Like in P2, most ages appear to be caused by volcanic grain inputs, but one sample
(831) contains grains which have an FT age which is significantly less than the depositional age, indicating grains were reheated to annealing temperatures (180°C) after deposition in the basin (figure 2.7).

Apatite Fission-track Results

Results from the apatite FT analysis are displayed in figure 2.8. Two samples from location 5 (figure 2.4) were analyzed yielding ages ranging from 16.5-638.4 Ma. Both samples indicate post-depositional annealing as a high percentage of grains have cooling ages which are less than depositional ages. Sample 829-a is a single granitic clast with a depositional age of ~55 Ma and an apatite FT age of 29.4 ± 3.7 Ma, indicating the clast was heated to temperatures within or above the apatite PAZ after deposition in the basin.

Apatite FT data from Kiowa and Castle Pines Cores (Kelley, 2002) was combined with dating of the cores (Hicks, 2002) to reveal potential post-depositional annealing in apatite samples throughout the cores (figure 2.9). Sample sizes are very small for the core data and therefore large statistical uncertainties exist in these data. Some samples (Kiowa 209, 259) (figure 2.10) exhibit strong post depositional annealing peaks and contain a marginally significant number of grains (~10-15), while others (Castle Pines 319, 446, 611) have peaks based on less than 5 grains.

Discussion

Apatite and zircon fission-track results do not give information explaining the Laramide exhumation of the Front Range but indicate limited post-depositional annealing may have occurred in the Denver Basin. Abundant volcanic grains and small sample sizes combine to produce results which do not give information about the regional thermal history. The stratigraphy indicates rocks from below the late Cretaceous PAZ (depth ~3 km, top of granitic basement) were exhumed and deposited in the basin between ~70-68 Ma (Chapter 1), and therefore apatite FT data should show a strong cooling event around 70-68 Ma. Apatite FT results (this study) indicate grains were annealed in the Eocene and Oligocene, suggesting post-depositional annealing assuming a D2 depositional age of ~58-55 Ma. Data from the Kiowa and Castle Pines cores
(Kelley, 2002) shows evidence of post-depositional annealing, but abundant volcanic grains and small sample sizes leave only room for speculation.

Apatite surface samples (this study) indicate post depositional annealing, but there is reason to suspect the sediments may have been reworked, introducing young grains because FT ages are indicative of Cripple Creek volcanism, and clasts of Wall Mountain Tuff were observed in assumed D2 sediments. Grains may have been reset by the Cripple Creek Intrusion in the Front Range, eroded from the mountains, and deposited in paleovalleys cut into the syn-orogenic sediments. Apatite surface samples were collected from sample location 5 (figure 2.4), directly west of completely and partially reset granitic basement rocks, in a likely drainage path. The FT age of sample 829-a (granitic clast) 29.4 ± 3.7 Ma is comparable to FT ages (27 ± 2 Ma) observed by Kelley and Chapin (2004) in granitic basement rocks completely reset by the Cripple Creek Intrusion. Additionally at sample location 5, clasts of Wall Mountain Tuff (~37 Ma) were found imbedded in rocks thought to be in the D2 sequence due to their arkosic lithology and close proximity to the basal paleosol. The presence of younger Wall Mountain Tuff clasts and granitic clasts with a thermal signature matching basement rocks reset by the Cripple Creek Intrusion suggests the original D2 deposits were eroded and replaced by a nearly identical, but younger sedimentary deposit. Reworking introduces younger sediments and would explain the young FT ages observed in the apatite samples. If the deposits were not reworked, the FT data suggest post depositional annealing occurred, with temperatures greater than ~60-80° C in the southeastern part of the basin.

One zircon sample (831, location 6) had an FT age which was less than the depositional age. Many of the zircon grains in the Front Range source rocks were partially or completely annealed by the Oligocene Cripple Creek Intrusion. If the sediments were not reworked, it would suggest temperatures of greater than 175° C in the southern part of the basin.

Zircon FT data and petrographic analysis (Wilson, 2002) finds young volcanic grains throughout the syn-orogenic stratigraphy, making the identification of the ~70-68 Ma exhumation event difficult. Volcaniclastic rocks have zero lag time because their depositional age is the same as their cooling age. Uncertainties associated with the depositional age and the greater uncertainties associated with fission-track analysis make
the separation of a zero lag time and a 1-2 My lag time very difficult. Furthermore, the reworking of older sediments which contain volcanic grains can create false lag times which are not indicative of orogenic exhumation.

Apatite data from the cores (Kelley, 2002) have grain populations which may have been partially annealed after deposition. Partial annealing does not completely reset the grain, but rather reduces the FT age, such that the age is between the previous cooling event and the later reheating event. Partial annealing can be detected when a FT age is less than the depositional age, and many of the samples collected from the core have FT ages which are less than depositional age (figure 2.10). However, not all samples collected from the cores indicate they were annealed. Sample Kiowa 366 has nearly identical ZFT peak age, AFT peak age, and depositional age (~66 Ma) (figure 2.11), suggesting the sample was not partially annealed, and that both apatite and zircon grains have a volcanic origin. Kowa 366 is deeper in the core than samples which appear to have been annealed. This “annealing upward” pattern is unusual for post-depositional annealing as deeper samples should be heated to greater temperatures than higher samples.

The young FT ages observed in the core could be due to small sample sizes, which if larger would indicate a volcanic event rather than basin annealing. Some samples (Kiowa 209, 259) have relatively strong partially reset peaks (10-20 grains) while others are based on only 2 grains (Castle Pines 319, 446) (figure 2.10). Fission-track data is generally associated with large uncertainties and therefore large sample sizes (100 grains for detrital, 30-40 grains for bedrock samples) are necessary separate the peaks from the edges of age distributions.

**Conclusion**

Detrital apatite and zircon fission-track analysis confirms the existence of volcanic grains at all syn-orogenic stratigraphic intervals indicating volcanic activity occurred throughout the Laramide Orogenesis, but FT analysis fails to detect Laramide exhumation cooling or definitively determine if post-depositional annealing occurred in the Denver Basin. Flat slab subduction has been proposed as a mechanism for Laramide orogenesis, and is generally associated with a volcanic window (area of a lack of
volcanism) in the flat slab region (English and Johnston, 2004). Volcanism throughout Laramide orogenesis challenges the flat slab hypothesis by indicating no volcanic window existed.

Abundant volcanic grains in the syn-orogenic stratigraphy make identification of Laramide exhumation cooling difficult. Detrital granitic clasts collected in this study, which cannot contain volcanic grain populations, contain few or no apatite grains. Apatite poor source rock and abundant volcanic grain input combine to mask the detrital Laramide exhumation cooling.

The one granitic clast which yielded abundant apatite grains had a thermal signature indicative of annealing due to the Cripple Creek Intrusion. This suggests synorogenic sediments were reworked in the Oligocene or that the southern portion of the basin was annealed after deposition. Further analysis of detrital samples is necessary to definitively conclude if post depositional annealing or sediment reworking occurred in the Denver Basin syn-orogenic sediments.
Figure 2.1 – Representative figure of the partial annealing zone (PAZ) (Left) and chart showing effective closure temperatures for various thermochronometers (Right). As depth increases temperature increases. Fission-track age decreases throughout the PAZ until the base, where the FT age is effectively zero. Cooling the sample through the PAZ quickly will yield a lower effective closure temperature while faster cooling rates yields a higher effective closure temperature. Left figure from Kelley (2004), Right from Reiners and Brandon (2006)
Figure 2.2 – Figure demonstrating the distribution of FT ages commonly observed in detrital samples. Peaks represent differential exhumation rates throughout the mountain belt. Generally in the core of the orogeny, exhumation rates are fast and young grains are exhumed, while on the flanks exhumation is slow and older grains are exhumed. Modified from Bernet et al. (2005).
Figure 2.3 – Figure demonstrating the application of Lag-time. Lag time is the fission-track age minus the depositional age and represents exhumation rates in the source terrain. Lag time plot can show changes in the exhumation rate throughout the orogeny. Increasing lag time represents slowing exhumation rates, decreasing lag time represents increasing exhumation rates and a constant lag time represents steady state exhumation. Modified from Bernet et al. (2005).
Figure 2.4 – Locations of FT samples collected in the Denver Basin
### Detrital Zircon FT Data

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**Figure 2.5** – Detrital zircon fission-track data. Data generally groups into six age peaks. One sample (red) has a FT age which is less than the depositional age indicating partial annealing may have occurred post deposition.

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### Detrital Zircon FT data from Kelley, 2002

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**Figure 2.6** — Zircon Fission-track data from the Kiowa and Castle Pines wells (Kelley, 2002; Hicks et al, 2002). Data generally mirrors ft results from this study and indicates volcanic input at all stratigraphic intervals.

**Figure 2.7** — Probability density plot for individual grain ages for sample 831 (depositional age 68 Ma). Bimodal distribution (red) with peak ages of 48.7 ±8.8 and 64.3 ±11.1. If depositional age is accurate, sample was annealed after deposition.
Table 2.8 – Apatite data from surface samples collected in the Denver Basin. Samples have FT ages which are less than depositional ages indicating post depositional annealing (red).

<table>
<thead>
<tr>
<th>Location</th>
<th>Samp. No.</th>
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<th>n</th>
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</table>

Figure 2.8 – Apatite data from surface samples collected in the Denver Basin. Samples have FT ages which are less than depositional ages indicating post depositional annealing (red).

Table 2.9 – Apatite Fission-track data from the Kiowa and Castle Pines wells (Kelley, 2002; Hicks et al, 2002). Red cells highlight where FT ages are less than depositional ages indicating post depositional annealing.
**Figure 2.10** – FT data which indicates post-depositional annealing in the Kiowa and Castle Pines cores (Kelley, 2002). Grain counts are generally very low for detrital fission-track analysis and therefore uncertainties associated with the age peaks are large. It is unclear as to whether the young ages are caused by low grain counts or actual post-depositional annealing. Charts from Kelley (2002), depositional ages from Hicks et al. (2002)
Figure 2.11 – Kiowa 366 m – depositional age 66 Ma – Relatively high grain counts compared with other core samples. Depositional age is the same as apatite and zircon FT age indicating a volcanic origin for the grains. Kiowa 366m is lower than samples which have age peaks indicative of partial annealing. The lack of partial annealing in this sample indicates the low FT ages in other samples may be due to the small sample sizes.
Acknowledgements

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