The Sequence Stratigraphy of the Chinle Formation in the Dinosaur National Monument Region, Utah and Colorado, USA

A THESIS SUBMITTED TO THE FACULTY OF THE GRADUATE SCHOOL OF THE UNIVERSITY OF MINNESOTA BY

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ABSTRACT

Deposition of Late Triassic Chinle Formation fluvial and lacustrine sediments in the Dinosaur National Monument (DNM) region of Utah and Colorado is bounded by two major periods of landscape degradation (stratal truncation) and punctuated by a period of non-deposition (well-developed paleosol pedogenesis). Correlation of these unconformable regional features in outcrop and geophysical well-logs enable the identification of two genetically related Chinle depositional sequences.

The first depositional sequence includes the Gartra Member and the mottled member. These units are located within regional paleovalley networks that are attributed to Mid-Triassic Tr-3 sequence boundary erosion. The basal Gartra Member conglomeratic sandstone was deposited in topographically-constrained, low-sinuosity fluvial systems. As aggradation continued, the lateral constraint of the paleovalley on the fluvial system decreased and higher sinuosity fluvial deposits and minor floodplain deposits are preserved. Overlapping the paleovalley margins and covering the entire landscape are fine-grained mottled member floodplain deposits. Capping the mottled member is a well-developed paleosol that is a non-depositional sequence boundary and the end of the first depositional sequence.

The second depositional sequence includes the ocher member and the upper member. Ocher member siltstone deposition occurred in large, shallow, evaporative, non-marine influenced, lacustrine environments. Initial upper member sandstone and siltstone units were deposited in fluvial-deltaic environments. Later upper member deposits are weak-red lacustrine siltstones with few interbedded fine-grained sandstones. The J-0
erosional unconformity marks the top of preserved Chinle deposition and the bottom of the subsequent Jurassic eolian sandstones.

After identification and analysis of Chinle depositional environments and known allogenic influences (a gradually drying climate), an autogenic mechanism is proposed to be the primary influence on the observed facies changes. In a basin with pre-existing erosional topography, the annealing of this landscape could initiate autogenic retreat of the depositional systems leading to distinct facies changes in the distal basin. In the first Chinle depositional sequence (Gartra Member-mottled member), the annealing of the Tr-3 paleovalley network caused a decrease in fluvial energy and the autogenic retreat of deposition into the proximal Eagle Basin. In the distal basin, continued subsidence and a period non-deposition resulted in a well-developed paleosol horizon (mottled member paleosol) and the transgression of lacustrine facies (ocher member). As the sediment-charged fluvial mega-fan prograded out of the Eagle Basin, sedimentation in the distal basin gradually increased until thick fluvial-deltaic deposits (upper member sandstones) developed filling excess basin accommodation. Subsequent allogenic drying climatic conditions may explain a decrease in later distal basin sedimentation and the second observed lacustrine transgression (upper member siltstones).
ACKNOWLEDGMENTS

While it is my name that gets to be placed on the binding of this thesis, without the invaluable insight and support of many people, the completion of this project would have been unattainable.

For introducing me to the Colorado Plateau and sequence stratigraphy, I would like to thank Tim Demko. His inspirational passion for and knowledge of Utah geology, along with many epic road trips around the plateau, enabled this project to be both very successful and rewarding.

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CHAPTER I
INTRODUCTION

1.1 Purpose

The Upper Triassic Chinle Formation was deposited in two major depositional basins within the Colorado Plateau province of the southwestern USA. The southern basin is centered near the Four Corners region and is the larger and more extensively studied of the two. This study focuses on the smaller northern basin that is located in Northeast Utah and Northwest Colorado and contains Dinosaur National Monument (DNM). Within this basin, terrestrially-deposited fluvial and lacustrine successions preserve a unique record of the changing paleoenvironmental, tectonic and paleoclimatic conditions in western Pangaea. This project uses a sequence stratigraphic framework to better understand the architecture, evolution, and geographic extent of these ancient paleoenvironments on both local and regional scales. Specific research objectives are as follows:

1) Describe the sedimentology, depositional facies and large-scale architecture of the Chinle Formation in the DNM region of Utah and Colorado;

2) Use a sequence stratigraphic framework within the Chinle strata to identify major and minor depositional facies changes and their possible causes;

3) Interpret how different regional Chinle facies are genetically related;

4) Interpret and model the depositional history of the northern Chinle basin;

5) Clarify the northern Chinle basin stratigraphic nomenclature.
Figure 1.2-1  Field map depicting measured section and well-log locations. FG=Flaming Gorge, DF=Dry Fork Canyon, MN=Mine Property, RC=191 Road Cut, RF= Red Fleet State Park, RN=Racetrack Nose, RT=Racetrack, SS=Sounds of Silence, BL=Boat Launch, ES 1 & 2=Elder Spring 1 & 2, BI 1 & 2= Bourdette Island 1 & 2, BD 1, 2 & 3=Bourdette Draw 1, 2 & 3, SBD=South Bourdette Draw, SC=Skull Creek, DD=Disappointment Draw, DL=Deer Lodge Park, ## =Geophysical Well-Logs.
The geographic focus of this study is the Dinosaur National Monument region of northeast Utah and northwest Colorado (Figure 1.2-1). In this region Upper Triassic strata were exposed during tectonic uplift and subsequent erosion related to the Late Mesozoic- Early Cenozoic Laramide Orogeny (Perry et al., 1992). Chinle outcrops used in this study are located to the south-east of the Laramide-type east-west trending Uinta Uplift. The most extensive exposures of Chinle are located in Utah along the edges of the Split Mountain Anticline within DNM (Figure 1.2-1; Sections RN, RT, SS, BL, ES1, ES2) and to the south in Bourdette Draw (B11, B12, BD1-3, SBD). Other measured sections within Utah are located in the Red Wash (RW) on the northern side of the anticline, just north of the town of Vernal (DF, MN, RC, RF), and further north in the Flaming Gorge National Recreation Area (FG). In Colorado, sections were measured at the western-most end of the monument (DL) and to the south (DD, SC). Scientific research permits were obtained from Dinosaur National Monument (Study #:DINO-00105; Permit #:DINO-2005-SCI-0004) and the Flaming Gorge NRA. Geophysical data was acquired for wells primarily to the south and to the east and west of DNM in both Utah and Colorado (Figure 1.2-1).

1.3 METHODS

1.3.1 Field Methods

During the summer of 2005 and spring of 2006, 21 stratigraphic sections were measured (Figure 1.2-1, Appendix A), described, and sampled. Outcrop measurements were taken using a 1.5 meter jacob’s staff and a Brunton compass. Measurements began at the Moenkopi-Chinle contact and continued to the Chinle-Nugget contact. Outcrops were chosen based on accessibility and unit exposure. In sections where exposure was poor, trenching was used to view unaltered outcrop. Unit descriptions include both
lithologic analysis and description of local unit relationships using a sequence
stratigraphic framework. Large and small scale photographs and photopans were also
used to supplement the unit descriptions and enable region-scale unit correlations (Figure
3.3-1, 3.3-2). Satellite photos of the outcrops, with unit scale resolution, were also viewed
using Google Earth.

1.3.2 Geophysical Well Log Analysis

31 geophysical well logs (Figure 1.2-1) were used to carry the Chinle stratigraphy
described in outcrop into the subsurface and expand the study area. Geophysical
correlations with Chinle members were accomplished by either comparing the interpreted
well log lithology to nearby measured sections or by finding well logs with lithologic
descriptions accompanying the geophysical tests. Gamma ray, neutron-density, sonic, and
acoustic geophysical logs were all used to identify Chinle units. To acquire this data,
regional well locations and API numbers were found on the internet using Utah and
Colorado’s geological survey sites. Digital well logs were then obtained through MJ
Systems, Inc. educational research grant program. Using MJ Systems LogSleuth
program, the logs were analyzed and unit tops were identified. Using the Landmark
Geographics Co. smartSECTION well log correlation program, well logs and digitized
measured sections could be entered and correlated (Figure 3.3-1, 3.3-2). Well logs were
hung on regional datums within and above the Chinle.
CHAPTER II
GEOLOGIC SETTING

2.1 TECTONICS

Continental aggregation, beginning in the Carboniferous and ending in the Triassic, culminated with the continental landmasses of Gondwana, Laurussia, and large portions of modern Asia and combing to form of the Pangaea Supercontinent (Figure 2.2-1) (Parrish, 1993). During this period, plate tectonic activity directly affected the nature of the Late Triassic landscape. During the Pennsylvanian and Early Permian, the Ouachita collisional orogeny on the southeastern continental margin caused the uplift of the metamorphic-cored Ancestral Rocky Mountains (Dickinson, 1981; Kluth and Coney, 1981). During the Late Permian and Early Mesozoic, the Sonoma orogeny occurred on the western Pangaea margin as an island arc collided with the supercontinent (Speed, 1979; Stewart, 1980). This collisional event created an Andean-type (Dickinson, 1981) or Central American-type (Busby-Spera, 1988) continental margin with a large volcanic arc and continental back-arc basin where California, western Nevada and Arizona are today. Also attributed to the Sonoma orogeny is the erosion and/or nondeposition of Middle Triassic strata (Speed, 1978). A major Middle Triassic continental upwarping may have created this stratigraphic gap and caused the reorganization of Early Triassic depositional basins (Dubiel, 1994). Within the Eagle basin, abnormally thick Pennsylvanian, Permian and Triassic deposits may be explained by increased tectonic subsidence or movement of Pennsylvanian evaporite deposits (Dubiel et al., 1992).

The concept of Dynamic topography has also been applied to Pangaea continental evolution. This concept is defined as the vertical displacement of the Earth's surface generated in response to flow within the mantle (Burgess et al., 1997). On
Triassic Pangaea, two dynamic topography models have been suggested (Figure 2.1-1a). The first model suggests that laterally extensive supercontinental crust acts as an insulator that causes the temperature of the mantle below the continent to rise. As the mantle temperature increases it begins to rise and form a mid-continental topographic high that eventually leads to continental rifting (Gurnis, 1988). The second model relates to the initiation and evolution of Pre-Farallon plate subduction on Pangaea’s west coast during the Triassic (Figure 2.1-1b) (Gurnis, 1992). Gurnis (1992) divides this model into 3 stages that begin with initial high-angle subduction of old, cold, dense, oceanic crust that pulls down on the continental edge creating a large back-arc basin, an interior forebulge, and a back-bulge basin (Figure 2.1-1b). As the mid-oceanic ridge nears the continental plate, the subducting oceanic crust becomes younger, warmer, and more buoyant which results in a shallower subduction angle and uplift in the continental basin. Burgess et al. (1997) used these models, along with a generalized tectonic history of the western United States, to model how dynamic topography could have impacted landscape aggradation and degradation in the western United States through time. Lawton (1994) predicted that a dynamic forebulge (Gurnis, 1992) could block fluvial basin drainage in the back-bulge basin resulting in the observed changing fluvial and lacustrine depositional facies observed in Triassic strata.

Subsequent Late Cretaceous-Cenozoic age tectonic activity attributed to the Laramide orogeny created the modern Rocky Mountains, the Uinta Mountains and caused the deformation, uplift, and exposure of Paleozoic and Mesozoic strata including the Chinle Formation (Hintze, 1988).
Figure 2.1-1 Dynamic topography; (a) Graphic showing the two types of dynamic topography acting upon the Pangaeansupercontinent (modified from Burgess et al., 1997). (b) Illustration of the affect of subduction related dynamic topography on the continental margin (modified from Gurnis, 1992).

2.2 PALEOGEOGRAPHY

During the Triassic, the Pangaeansupercontinent reached its maximum size (Figure 2.2-1). This landmass sat symmetrically over the paleoequator and stretched from 85° N to 90° S (Ziegler et al., 1983). The northern Chinle depositional basin was located near the western equatorial margin of Pangaea at a paleolatitude of approximately 15° N. (Van der Voo et al., 1976; Pindell and Dewey, 1982). During the Early Triassic this region was covered by a shallow to marginal marine environment (Dubiel, 1994). Middle
Triassic subduction-related cratonic flexure and uplift led to reorganization of the regions depositional basins (Dubiel, 1994) and countered the low but rising Triassic eustatic sea levels (Haq et al., 1987; Vail et al., 1977). During the Late Triassic, the nearest marine shoreline would have been in the back-arc basin over 500 km to the west of DNM region (Figure 2.2-2)(Dubiel, 1994). This distance would presumably have prevented any eustatic influence on the basin.

Intimately related to the character of Late Triassic deposition are the Ancestral Rocky Mountains which include the Uncompahgre Highlands and the Ancestral Front Range (Figure 2.2-2). These Pennsylvanian-Permian age, metamorphic-cored, low-relief uplifts separated the Chinle into the two major depositional basins and directly influenced regional sediment input and flow (Stewart et al., 1972). The Ancestral Rockies bounded the northern basin to the south and east and directed basin drainage to the west and northwest (Dubiel, 1992). These uplifts were eventually overlapped by Late Triassic and Jurassic deposition (Dubiel, 1994).

Figure 2.2-1 Late Triassic paleoenvironmental reconstruction of the Pangaea Supercontinent. Colored ovals represent the Chinle depositional basins. Modified from a Blakey image at http://jan.ucc.nau.edu/~rcb7/RCB.html.
2.3 PALEOClimate

Pangaea's size and symmetrical position over the paleoequator (Figure 2.2-1) dramatically influenced Triassic climatic conditions by disrupting global atmospheric circulation patterns (Dubiel et al., 1991). Robinson (1973) was the first to describe the Pangaean climate as monsoonal based on the wide distribution of Triassic red beds. Monsoonal circulation is driven by cross-equatorial flow due to temperature and pressure contrasts between winter and summer hemispheres (Dubiel et al., 1991). As a result of this circulation, there are seasonal precipitation extremes with very wet summer months and considerably dryer winters (Dubiel et al., 1991). Conditions that impact the magnitude of the monsoon cycles include the size of the continent, moisture and heat source characteristics, and the regional topography, all of which were ideal on the
Pangaean continent for maximum monsoonal circulation (Dubiel et al., 1991). These wet mega-monsoonal conditions are unique to the Triassic and are generally bounded by Pennsylvanian-Permian and Jurassic eolian deposition, which were deposited in arid to hyper-arid conditions (Peterson, 1994). Dubiel et al. (1991) supported the Late Triassic mega-monsoon hypothesis based on analysis of Chinle Formation sedimentology, paleontology and paleoclimatic models.

Lithofacies and ichnofossils analyzed by Dubiel et al. (1991) show that the Late Triassic depositional environments (rivers, floodplains, lakes, deltas...) were wet with fluctuating water tables. Fluctuating water table interpretations were based on paleosol mottling due to mineral alteration and translocation and variations in crayfish and lungfish burrow depths (Dubiel et al., 1992; Hasiotis et al., 1998). Lithofacies changes in the upper part of the Chinle (e.g. eolian sand sheet strata in the Eagle basins upper red siltstone member) were interpreted as evidence of a Late Triassic climatic drying trend (Dubiel et al., 1991). At the end of the Triassic, large eolian erg systems dominated the landscape. Possible explanations for this drying trend include; 1) a northern plate tectonic migration of the basin away from the paleoequatorial region; and 2) an end to ideal mega-monsoonal conditions due to the break up of the Pangaeansupercontinent (Dubiel, 1994).
Figure 2.4-1  Nomenclature and gross stratigraphic architecture of Late Paleozoic to Mid-Mesozoic strata in northeastern Utah and northwestern Colorado. Unit thicknesses are not to scale.
2.4.1  **Pennsylvanian**

*Weber Sandstone.* A Middle Pennsylvanian unit composed of fine-grained, thick bedded, eolian quartz sandstones that display large-scale cross-bedding (Hansen et al., 1983). Fryberger (1979) determined that the eolian sands were blown in by winds predominantly from the north.

2.4.2  **Permian**

*Park City Formation.* A Permian formation separated into upper and lower units. The lower unit is unconformably found above, and sometimes truncating, the Weber Sandstone and is composed of light greyish brown marine sandstone, dolomite and limestone. This unit can be locally fossiliferous and phosphatic (Hansen et al., 1983). The upper unit of the Park City Formation grades into thin, light grey, locally fossiliferous, marine limestone, siltstone, sandstone, and dolomite beds. These units are correlative to the Phosphoria Formation found in Wyoming, Idaho, Utah, and Nevada (Hansen et al., 1983).

2.4.3  **Lower and Middle Triassic**

*Moenkopi Formation.* Lower Triassic strata found within the northern basin typically overlie Permian marine strata except in the ancestral Rocky Mountains where they unconformably onlap Pennsylvanian strata (Dubiel, 1992). These Triassic units include the Moenkopi Formation in the Uinta and Piceance Basins and the Permian/Triassic State Bridge Formation in the Eagle Basin. Moenkopi strata are predominantly easily erodable reddish-brown siltstones and shales with some interbedded gypsum and rippled fine grained sandstone layers (Gregson et al., 2000). Deposition
occurred during shallow marine transgressions and regressions that covered a large portion of the western craton (Dubiel, 1994).

**Tr-3 Unconformity.** During Middle Triassic time, a period of landscape degradation and non-deposition, possibly a result of tectonic activity along the western continental margin, created the unconformity termed the Tr-3 by Pipiringos and O’Sullivan (1978). This is typically seen as a sharp erosional contact that separates Lower Triassic Moenkopi red siltstones from the Upper Triassic Chinle conglomeratic sandstones (Stewart et al., 1972). Tr-3 degradation incised large paleovalley networks into the Moenkopi in both of the Chinle depositional basins. Within the DNM region, bedding above and below the Tr-3 is concordant making the contact an erosional disconformity (Stewart et al., 1972).

**2.4.4 Late Triassic**

Stratal interpretations and unit nomenclature of the Upper Triassic strata in the DNM region have varied over the past 50 years and are to this day a topic of debate. A major contributor to this nomenclature variation is the geographic location of the study area near the borders of Colorado, Utah, and Wyoming. In Utah and Colorado, the term Chinle Formation is used for Upper Triassic strata however the member characteristics and nomenclature vary by region. North of the Uinta Mountains in Wyoming, Popo Agie Formation nomenclature is used (Dubiel, 1992). In the Salt Lake region, the Ankarah Formation is used. Lucas (1993) proposed creating a “Chinle Group” that included Colorado Basin-wide correlations of all Upper Triassic strata. This proposal however disregards the geographic utility of region specific nomenclature, arbitrarily changes the rank of well-defined lithostratigraphic units, and does not follow the guidelines of the
North American Stratigraphic Code (Dubiel, 1994). In this paper, the nomenclature used by Stewart et al. (1972) and summarized in Dubiel (1992) will be used. The subformational units recognized in this study include the formalized Gartra Member, and the informal mottled, ocher, upper, and red siltstone members.

**Gartra Member.** Early research in northern Utah and Colorado recognized the Chinle and the basal Shinarump Conglomerate (Thomas et al., 1945, Kinney, 1955). Thomas and Krueger (1946) were the first to name the Upper Triassic basal conglomerate the Gartra Grit Member within the Stanaker Formation based on lithologic interpretations that separated the unit from the Shinarump Member to the south. These new terms were not generally accepted and subsequent research continued to use the terms Shinarump Member and Chinle Formation. Regional stratigraphic interpretations however determined that there were two separate Chinle basins and that the basal conglomeratic units were not directly correlative (Stewart et al., 1957, Poole et al., 1964). Therefore, the nomenclature now used for the basal conglomerate in the southern basin is the Shinarump Member and in the northern basin the Gartra Member. Time equivalence was later implied in paleogeographic maps by Stewart et al. (1972). Shropshire et al. (1974) deviated from this nomenclature pattern and lumped the genetically related Gartra and the mottled member into a “lower unit” and the rest of the Chinle into an “upper unit”.

The Gartra Member is recognized as a light colored, quartz-rich, fossil log-bearing, conglomeratic sandstone and sandstone that has a fining upward grain-size trend. In outcrop, the Gartra can be multi-story, amalgamated, cross-bedded channel deposits that form large cliff faces. The thickest Gartra units sit unconformably within the deepest incised Moenkopi paleovalleys (Stewart et al., 1972). Stewart et al. (1972) created an
outline of the paleovalley network based on Gartra exposures and interpreted that in the DNM region it was 200-300 km across (N-S) with interspersed topographic highs within. XRD analyses show that upper Gartra deposits are rich in kaolinite, which indicate formation in an oxidizing environment with intense leaching processes (Schultz, 1963). Maximum grain size in the northern basin significantly decreases away from the Ancestral Rocky Mountains which indicates that they were the major sediment source for the basin (Stewart et al., 1972). Based on the Gartra’s coarse grain-size, cross-bedded sedimentary structures, and large-scale depositional architecture, it has been interpreted to be a low sinuosity fluvial valley-fill deposit (Dubiel, 1992). Dubiel (1994) describes how an aggrading fluvial system confined within a paleovalley could evolve from a low to a high-sinuosity system with an overall fining upward grain size trend. Paleocurrent analysis by Stewart et al. (1972) and Shopshire (1974) indicate that the Gartra fluvial system flowed away from the Ancestral Rocky uplifts towards the west and northwest. Stewart et al. (1972) describes the Gartra Member and the mottled member as being “intimately related” with a gradational and intertonguing contact.

**mottled member.** The contact between the mottled member and the Gartra is placed at the top of the light-colored sandstone and the base of the purple and white siltstone (Stewart et al., 1972; Dubiel, 1992). Popo Agie Formation nomenclature from areas north of the the study area refers to this member as the purple unit (High et al., 1967).

The mottled member is composed of silty-to-sandy sediment that has a purple and white mottled appearance. Mottled member lithology typically consists of massive to thinly-bedded siltstones with interbedded Gartra-like sandstone lenses (Stewart et al.,
The mottled coloration has been interpreted to have been caused by alteration and translocation of iron bearing minerals caused in part by fluctuating water tables (Dubiel, 1992). Schultz (1963) determined that this clay is also rich in kaolinite indicating a fluctuating water table. Within this unit lungfish burrows (Dubiel et al., 1987), crayfish burrows (Hasiotis et al., 1989) and other preserved plant and invertebrate ichnofossils (Dubiel, 1992) are common and reflect water table depth and paleoenvironmental conditions. Dubiel (1992) interprets the mottled member to be pedogenically-modified floodplain soils with interbedded lateral accretion channel deposits.

Stewart et al. (1972) also describe modified Moenkopi below the Tr-3 unconformity as “mottled strata”, which they interpret to be strata altered by soil processes, and this should not be confused with the mottled member.

**ocher member.** The ocher member is found above of the mottled member. This unit is exclusively found in the western portion of the northern Late Triassic depositional basin and not in Eagle basin or the southern depositional basin (Stewart et al., 1972). The Chinle ocher member is also lithologically identical to the Popo Agie unit of the same name in central Wyoming (Keller, 1953).

The ocher member appears yellow to reddish orange in color and is composed of fine grained siltstones and claystones. The ocher member represents a large lacustrine unit that is generally structureless or thinly laminated (Dubiel, 1992). Within the ocher member, beds rich in montmorillonitic clay (Schultz, 1963) and containing nodules of analcime and carbonate (Keller, 1952) can be found and have been interpreted to be the result of diagenetic alteration of volcanic ash. The presence of analcime indicates saline-alkaline lake or diagenetic conditions after the ash-fall (Keller, 1952). High et al. (1969)
proposed that a large saline-alkaline lake was centered in southern Wyoming that at times expanded into the Uinta Mountain region. These lacustrine transgressions did not, however, reach into the higher elevation Eagle Basin. Dubiel (1992) uses the presence of saline-alkaline lake facies found exclusively in the northern basin, which may be time correlative to fresh water lake environments in the southern basin, to suggest that the northern basin experienced more arid, or closed hydrologic, climatic conditions.

Stewart et al. (1972) determined that the unit overlying the ocher member varied by region. Overlying eastern ocher member deposits is the red siltstone member and overlying western deposits is the upper member.

**red siltstone member.** In the northern Chinle basin, red siltstone member deposition is centered within the Eagle basin in NW Colorado where it is over 300 m thick and onlaps the basement rock exposed in the ancestral Rockies (Stewart et al., 1972). Away from the Eagle basin, this unit gradually becomes thinner and has been interpreted to be above the ocher member interfingering with the sandier upper member east of Vernal (Stewart et al., 1972). Red siltstone member lithofacies are typically structureless red siltstone, sandstone, or mudstones that may display lateral accretion bedding, carbonate nodules, ichnofossils, and horizontally bedded, wavy laminations (Dubiel et al., 1991). In the Eagle basin, lower red siltstone member depositional environments include fluvial channels, floodplain paleosols, and minor lacustrine deposits. These wet environments gradually evolved into upper red siltstone member paleosols and eolian deposits toward the end of deposition (Dubiel, 1994).

**upper sandstone and conglomerate unit.** In the easternmost Uinta Mountain region, Stewart et al. (1972) describe the composition of the sandstone and conglomerate
member as gray, pink, and brown sandstone with some siltstone and conglomerate. Horizontal lamination, ripple laminae, current lineation, and crossbedding are recognized. Conglomerate units are typically composed of mudstone rip-up pebbles. This unit is laterally restricted and can be traced into floodplain deposits (Dubiel, 1992). Stewart et al. (1972) recognized this facies as the basal member of the red siltstone member and thought that it may be regionally unconformable. In the eastern part of DNM, Dubiel (1992) identified the sandstone and conglomerate member interfingering with the red siltstone member and the ocher member. Dubiel (1992) interpreted this unit to be a fluvial cut-and-fill deposit.

upper member. The lower contact of the upper member is between the highest structureless siltstone (ocher or red) and lowest massive sandstone (Stewart et al. 1972). Stewart et al. (1972) interpreted that to the west of the Green River the upper member is overlying the ocher member and to the east it is overlying red siltstone member. In some locations the red siltstone unit interfingers with the upper member (Stewart et al., 1972). The upper member is composed of basal red, grey and brown sandstone and red grey and brown siltstone and red, brown, and green upper mudstones (Stewart et al., 1972). Upper member clays are predominantly illite which can form in many environments where mild leaching of silicate materials occur (Schultz, 1963). While the percentage of each of these rock types in the upper member is regionally variable, there is an overall fining-upward trend throughout the unit.

The lower portion of the upper member is a fine to medium grained sandstone that exhibits evidence of channelized flow, rapid sedimentation, standing water, and wave reworking (Dubiel, 1992). The basal portion of the sandstone is generally massive;
however, up section climbing ripples, planar and trough cross-bedding, and mudcracks may occur with interbeds of red siltstone (Dubiel, 1992). To both the west and east of Vernal, UT, the percentage of sandstone in the upper member decreases (Stewart et al., 1972). Above the sandstone and interbedded siltstones are mudstones. The mudstones may be structureless or thinly laminated and can contain ripple laminae, mud cracks, or pellets of clayey siltstone (Stewart et al, 1972). Bioturbation can be found in both the sandstone and siltstone (Dubiel, 1992). Dubiel (1992) interpreted that the upper member was deposited in fluvial-delta settings that graded into lacustrine environments. The J-0 unconformity cuts down into the upper member and is overlain by Jurassic cross-bedded eolian sandstones and marks the top of Chinle strata (Pipiringos et al., 1978). Upper member strata are correlative to the upper carbonate member of the Popo Agie Formation (Dubiel, 1992).

Another hypothesis by Jenson (2005) proposes that the upper member is a marine tidal deposit and is the basal member of the Nugget Sandstone. This proposal changes the nomenclature of the upper member to the bell springs member of the Nugget Sandstone. This work is based on correlations to the Bell Springs Member in Wyoming; however, it presents no definitive evidence (isotope analysis or marine fossils) of a marine transgression. Jenson (2005) acknowledges that the Bell Springs member is Triassic age, based on dinosaur trackway evidence (Lockley et al., 1992) and agrees that J-0 unconformity may be found at the base of the eolian Nugget Sandstone.

2.4.5 Jurassic

In the DNM region, eolian, grey-to-yellow, cross-bedded Glen Canyon Group sandstones sit above the J-0 erosional unconformity (Pipiringos et al., 1978). Peterson
interpreted that the lowest Jurassic unit is the Wingate equivalent Nugget Sandstone. Dubiel (1992) interpreted that in the DNM region the Nugget Sandstone is 3-10 meters thick and is overlain by much thicker lighter colored sandstone that is correlative to the Navajo Sandstone. Unconformably (J-1 unconformity) above the Glen Canyon Group are dark-red shallow-marine shales, siltstones, and mudstones that make up the Caramel Formation. In DNM, the Caramel thins to east and is not present in outcrop at the eastern end of the monument. Overlying the Caramel Formation, and unconformably the Chinle in the Eagle Basin, is the Middle Jurassic Entrada Sandstone (Dubiel, 1992). The Entrada Sandstone is a thick cross-bedded eolian sandstone that exhibits evidence of brief marine transgressions and related sediment reworking (Gregson et al., 2000).
CHAPTER III
DATA AND RESULTS

3.1 Unit Descriptions – Lithology, Facies, and Architecture

In the DNM region four unique lithostratigraphic units are evident in the Chinle Formation. These include the basal Gartra Member conglomeratic sandstone and the mottled member siltstone, the ocher member siltstone, and the upper member sandstone and siltstone. Each of these packages preserve unique depositional environments and have generally distinct lithostratigraphic bounding surfaces.

3.1.1 Basal Unconformity

The Tr-3 unconformity in the northern depositional basin is recognized by stratal truncation and/or pedogenesis of Moenkopi Formation strata. Moenkopi and Chinle strata are parallel making the Tr-3 an erosional disconformity. When found at the base of thick Gartra conglomerate and sandstone deposits (Figure 3.1.1-1a, 3.1.2-1a), the contact is very sharp and pedogenesis of the underlying Moenkopi is minimal, usually consisting of < 1 meter of bleached, green-grey, horizontally laminated siltstone. The finer-grained Moenkopi also weathers back, creating recesses beneath Gartra sandstone ledges. Where Gartra deposits are thin to nonexistent (Figures 3.1.1-1b, 3.1.2-1b), pedogenesis of the Moenkopi formation below the basal contact becomes much more intense and can penetrate 2-3 meters below the contact. Increased pedogenesis can make identification of the Tr-3 contact difficult, especially when modified Moenkopi is underlying pedogenically-modified mottled member. Modified Moenkopi retains some of its red coloration, which is typically not found in the mottled member, however it is also mottled with purple, orange, yellow, and green-grey colors from mineral alteration and translocation.
Figure 3.1.1-1 The Tr-3 unconformity. (a) Slightly altered Moenkopi under a large Gartra channel in the Racetrack. (b) ~3.5 meters of intensely altered Moenkopi that is overlain by a thin Gartra (G3) layer in the Flaming Gorge area.

3.1.2 Gartra Member/mottled member

In the field area the Gartra Member is a package of light-colored, cliff-forming, fining-upward conglomeratic sandstone and sandstone (Figure 3.1.1-1a, 3.1.2-1a). Sedimentary structures within the Gartra include medium-to-large scale planar and trough cross-bedding within stacked, lenticular shaped, channel deposits. Higher order erosional unconformities are present within the Gartra and were formed as subsequent channels partially incised previously deposited channel deposits.

Gartra thickness (0-40 m) is dependent on the depth of the paleovalley incised into the Moenkopi prior to Chinle deposition. In the study region, the thickest Gartra outcrops are located in the western half of DNM and just north of Vernal, UT (Figure 1.2-1). At the northern end of Bourdette Draw, a ~30 meter thick, ~5 km long Gartra conglomeratic-sandstone outcrop can be seen onlapping Moenkopi strata to the south until the sandstone unit pinches out altogether leaving mottled member strata overlying modified Moenkopi (Appendix A: 24). Thinner and less-laterally extensive sandstone-
filled channels do reappear further to the south. Geophysical well-log data also indicates that there are additional large channel systems with thick Gartra units several kilometers to the south. In Colorado and at the Flaming Gorge outcrops, basal channelized sandstones are thin, laterally discontinuous, and usually found directly associated with the mottled member.

Grain size at the base of the thicker individual Gartra channel deposits is typically coarse to very-coarse sand with some pebble and cobble-sized clasts and occasional fossil logs that reach 1-2 m in length. The composition of these conglomeratic-sandstones and sandstones is predominantly quartz and quartzite. Up section, the Gartra channel deposits fine upwards into medium grained sandstones. Silty interbeds and yellow and purple mottling also become more common. This fining upward trend is seen in both outcrop and geophysical well-log data (Figure 3.2-1). The Gartra sandstones eventually grade into the overlying mottled member siltstone deposits.

The mottled member can be found throughout the entire study region conformably overlying the Gartra Member or unconformably overlying the Tr-3 unconformity and altered Moenkopi Formation. This unit is characterized by pedogenically modified, silty-to-sandy deposits that have purple and white mottled coloration (Figure 3.1.2-1b). Interbedded within the floodplain deposits are white, sandy, lateral accretion channel deposits that vary in thickness (1-3 m.) and lateral extent. Bioturbation has destroyed most sedimentary structures. Some ichnofossils that are preserved include crayfish burrows and rhizoliths. In some outcrops, a vertic slickenside fabric can be seen.
Along the Bourdette Draw and Elder Spring outcrops, a very mature, laterally extensive, 1 to 2 meter oxic paleosol horizon was identified near the top of the mottled member (Figure 3.1.2-1b; Appendix A: 19, 24, 28). Compared to typical mottled member deposits, this unit has a darker hue, higher concentrations of burrows, and is more resistant, causing it to weather out in relief from the surrounding siltstones. In Bourdette Draw, this paleosol horizon can be traced for kilometers along the entire outcrop (from SBD past BI) and appears to be continuous and parallel to underlying and overlying Chinle strata. Lithologically similar units are found within the Racetrack (Appendix A: 12) however poor unit exposure makes them impossible to trace laterally. At Skull Creek, a resistant cherty layer is found within the mottled member in a stratigraphically similar position (Appendix A: 30).

3.1.3 ocher member

The ocher member is found in all of the outcrops in the study area. Lithologically it is a massive-to-planar bedded siltstone that can contain carbonate nodule horizons (Figure, 3.1.3-1). Unit coloration is typically ocher (yellow), red-orange, or red and contains some mm-cm scale light green circular mottles. Carbonate nodules are often found in 5-15 cm thick lateral horizons. Unit thickness and color patterns are variable. In some outcrops, ocher colored strata dominate the unit and in others thinner ocher beds are bounded by red siltstone at the bottom and/or the top. *Scoyenia* beetle burrows and vertic slickenside fabric are apparent within the some ocher member deposits (Figure 3.1.3-1c). *Scoyenia* burrows are typically found in alluvial and lacustrine soils with high moisture (~40%), high humidity (approaching 100%), and in indurated substrates (Hasiotis, 2002).
Figure 3.1.2-1 Lower Chinle members; (a) Gartra Member cross-bedded, amalgamated channel deposits in the Racetrack. (b) Mottled member siltstone deposit in Bourdette Draw with thin sandy lateral accretion channel deposits and a horizontally continuous pedogenically modified paleosol horizon.

Figure 3.1.3-1 The ocher member; (a) Ocher member strata bounded by the mottled member paleosol and upper member sandstone. (b) Close up ocher siltstone with small round grey mottles. (c) Vertic slickenside pseudo-anticlines.
3.1.4 upper member

Above the ocher member, brown-to-red, fine-to-medium grained sandstones occur that gradually fine upward into weak-red and light green interbedded siltstones (Figure 3.1.4-1). While this unit is recognizable in all of the measured sections and well-logs, its composition and structure are somewhat heterogeneous. In some locations the base of this unit is a thick massive sandstone that grades into rippled, cross-bedded, and interbedded sandstones with red laminar siltstones up section (Figure 3.1.4-1a). In other locations there are no thick basal sandstones. Instead, rippled and cross-bedded sandstones with frequent silty interbeds are found near the base of the unit (Figure 3.1.4-1b). Ancorichnus beetle burrows and Pelecypodichnus clam burrows may be found within these units.

In some locations, above the interbedded sandstones and siltstones are horizontally laminated mudstone, siltstone and very fine sandstone beds. At this contact there is a color change from red to weak-red with additional light green beds. Beds below the Jurassic sand contact are also typically bleached light green. An exception to this color trend occurs in the Flaming Gorge where the entire upper siltstone is light green. In outcrop, the upper unit tends to weather back and be poorly exposed; however, the interbedded light green beds are more resistant and stand out. At the north end of Bourdette Draw, bedding units are clearly seen and mud cracks and soft sediment deformation were identified. Within the unit small (<2 cm) scattered white cherty (?) nodules were found. DNM paleontologist Scott Madsen did not believe that they were fossil material when he viewed the outcrop (Madsen, personal communication). Due to J-0 erosion, this unit is not always found in outcrop.
3.1.4 upper member; (a) ~8 m thick amalgamated, cross-bedded upper member sandstone deposit near Red Fleet State Park with Nugget Sandstone at the top-right. (b) <1.5 m upper member sandstone beds with silty interbeds in Bourdette Draw. Weak red upper member lacustrine siltstones are overlain by Nugget sandstones.

3.1.5 red siltstone member

The red siltstone member is definitively found outside of the study region within the Eagle Basin overlying lower Chinle Gartra and mottled member deposits. It is recognized as red and siltstone and fine sandstone. These units may contain burrows, bioturbation, carbonate horizons, and other evidence of pedogenesis. Near Rifle, CO, thick (<200 m) red siltstone member deposits were identified between Gartra conglomerates and the J-0 unconformity at the base of the Jurassic sandstones.

3.1.6 J-0 Unconformity

In the DNM region, this unconformity is found at the base of the cross-bedded Nugget Sandstone. When the Jurassic sandstones overly upper member siltstones and mudstones, the contact is easily identified by color, grain-size, and a higher degree of Chinle weathering which leaves a sandstone ledge (Figure 3.1.5-1b). When J-0 associated erosion has removed the finer grained upper member strata, similar Jurassic sandstones and upper member sandstones are in contact making identification of the unconformity more difficult (Figure 3.1.4-1a). Grain size, bedding thickness, sedimentary structures, and geophysical characteristics can be used to differentiate them.
3.2 GEOPHYSICAL WELL-LOG ANALYSIS

By comparing geophysical well-log data to associated lithologic well-logs and nearby measured sections, Chinle units were identified and correlated into the subsurface. Gamma ray logs were most helpful in determining lithology, while neutron-density, sonic, acoustic, and sample logs were also used to provide secondary checks. Gamma ray geophysical tools measure the radioactivity, on an API unit scale, of strata along the length of the bore hole. Siltstones and mudstones typically contain more radioactive material (typically clay with radiogenic K\(^{40}\)) than sandstones and conglomerates therefore have a higher gamma ray value. Wells with both sample and gamma ray logs were used check this assumption and identify Chinle unit signatures. The following paragraphs will describe common well-log signatures that were used to identify Chinle deposits in the sub-surface.

The gamma ray signature of the Moenkopi Formation (Figure 3.2-1) has relatively high-to-medium level API readings that alternate in a serrated pattern indicating alternating beds of siltstone and fine grained sands. The most indicative Moenkopi feature however is the overlying, thick, Gartra Member with characteristically low API values.

The thick basal valley-fill Gartra channels appear as blocky stacked units that are separated by thin high gamma ray spikes (Figure 3.2-1). In other locations, the stacked channel deposits are not as prevalent and thicker siltstone units are found within the Gartra sands. Towards the top of the unit, grain size fines upward into alternating siltstones and sandstones that represent the mottled member floodplain deposits with interbedded channels. Near or at the top of the mottled member, a high gamma ray high is
commonly found that may mark the change from the floodplain deposits to the overlying muddier lacustrine deposits.

Above the mottled member spike, the base of the next unit is a relatively high gamma ray, low neutron-density, fine grained deposit that correlates to the ocher member siltstones (Figure 3.2-1). This unit gradually coarsens upwards, with few major gamma ray spikes in either direction, into fine grained sandstone. Neutron-density surveys sometimes show a peak within this section that is approximately 7 meters thick.

The upper member is identified by a large rise and plateau in the neutron density, sonic and acoustic logs and blocky, relatively low API readings with thin but frequent interbedded high gamma ray spikes (Figure 3.2-1). These units are interpreted to be fine-grained, fluvial-deltaic sandstone units with silty interbeds. Up section, grain-size decreases and fine grained siltstones deposits are interbedded more frequently with thin sandstone deposits. In the study area, the finer grained upper units may be thicker units (<10 m) or thin interbeds in the east.

The J-0 contact (Figure 3.2-1) is identified by massive, low gamma, eolian sandstone unit that continues for hundreds of meters. When eolian sandstones are overlying fluvial-deltaic sandstones, analysis of the secondary geophysical tests can help identify the contact because upper member Chinle units generally have high and low neutron-density spikes whereas the overlying sandstone is relatively more consistent.
<table>
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<th>Comp. Neutron</th>
<th>Dep. Environment</th>
<th>Unit</th>
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<td></td>
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<td>eolian erg</td>
<td>Nugget Frm.</td>
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<td>upper mbr.</td>
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<td>high sinuosity fluvial</td>
<td>mottled mbr.</td>
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<td></td>
<td></td>
<td>shallow marine</td>
<td>Moenkopi Frm.</td>
</tr>
</tbody>
</table>

Figure 3.2-1  Geophysical well log showing Chinle Formation strata and environmental interpretations.
Figure 3.3-1: Fence diagram of study region, UT & CO
(see Figure 1.2-1 for exact locations)

Note thicker Gartra and ocher member units in the west and thicker preserved upper member units in the east. Upper Chinle unit differences may relate to the sections proximity to the sediment source.
Figure 3.3-2: Fence diagram of outcrops on the Utah side of DNM (see Figure 1.2-1 for exact locations)

- Nugget Formation
  - upper member (deltaic & lacustrine)
  - ochre member (lacustrine)
  - mottled member (floodplain)
  & Gartra Member (fluvial)

- Moenkopi Formation
  - Tr. 3 unconformity

Interpretation:
- Interpreted direction of local paleovalley fluvial flow.
- Gartra Member onlaps out on paleovalley margin

Note: The major decrease of Gartra paleovalley conglomeratic sandstones to the south. Also note the changing upper member characteristics.

Scale: ~15 miles E-W
~10 miles N-S
4.1 Introduction

The concepts and applications of sequence stratigraphy have revolutionized the field of sedimentary geology. In contrast to a lithostratigraphic approach, where similar units are correlated based on physical properties, sequence stratigraphy integrates aspects of both sedimentology and stratigraphy to better understand how genetically related stratigraphic units, facies tracts, and depositional elements relate to one another in both time and space (Figure 4.4-1) (Catuneau, 2006). This discipline builds upon Walther's law, which recognizes that the same depositional successions that are present in a vertical plane are also present in the horizontal plane, barring a break of sedimentation (Posamentier et al, 1999). This law is used by sequence stratigraphers to interpret and predict depositional trends that would occur in a basin environment and constrain and identify the influences on that depositional system (eustasy/limnostasy, tectonics, climate, sediment source). For instance, in Figure 4.1-1, if any of the basins boundary conditions were altered, a shift in all three facies would be seen. If sea level rose, the facies would transgress landward. If tectonics increased sediment flux, the facies would prograde basinward. Each environment is genetically related in time and space.

Figure 4.1-1 Three unique depositional environments with unique lithologies that are temporally and spatially correlative. The cross section shows a prograding fluvial-deltaic depositional package.
4.2 Continental Sequence Stratigraphy

Historically, sequence stratigraphy has been most successfully applied to nearshore marine sedimentary deposits affected by eustasy (Posamentier et al., 1999). There are however models that have been developed to help predict how continental depositional systems will react to allogenic tectonic and climatic changes.

For continental lacustrine systems, Carroll and Bohacs (1999) analyzed the depositional facies of ancient and modern lake environments. They found that both potential accommodation, related to tectonism, and the influx of sediment + water, related to climate, determine lake occurrence, distribution, and character. Based on the balance between these criteria, the lacustrine depositional environments can be divided into overfilled, balance-filled, or underfilled basins (Carroll et al., 1999)(Figure 4.2-1). These lacustrine basin-types each preserve unique depositional facies (Figure 4.2-2) that may change as tectonic and climatic conditions change.

Figure 4.2-1 Lacustrine basin classification model (Carol and Bohacs, 1999)
For continental alluvial systems, Lane (1955) developed a model (Figure 4.2-3) that roughly outlines the reaction (incision or aggradation) alluvial systems experience as boundary conditions (sediment supply and size, discharge, and slope) change through time. This is a good conceptual model however it does not account for other factors that influence alluvial depositional systems (i.e. vegetation).

Figure 4.2-3  Alluvial deposition model outlining the general impact of changing boundary conditions (Lane, 1955).
Catuneanu (2006) also proposed a continental fluvial system sequence stratigraphic model that defines low- and high- *accommodation systems tracts (AST)* instead of the traditional eustatic base-level driven *lowstand-, transgressive-, and highstand-systems tracts*. Like traditional eustatically-influenced depositional sequences, continental fluvial depositional sequences are composed of genetically related strata that are bounded by unconformities. Unlike marine influenced models, low- and high- AST’s are the building blocks of the depositional sequences and they are solely influenced by tectonism, sediment source, and climate (Catuneanu, 2006). In the AST model, low-AST (LAST) deposition begins after a local depositional hiatus or period of degradation. These LAST initially experience low rates of increased accommodation and fill previously incised valleys with coarse-grained, amalgamated, low-sinuosity channel deposits. These coarse deposits grade into pedogenically modified LAST floodplain and paleosol deposits. As fluvial deposition continues, the rate of accommodation creation increases and the LAST’s grade into high-AST’s (HAST) with thicker fine-grained, high-sinuosity fluvial system deposits with fewer paleosols and channel bodies. This cycle renews itself when the landscape is steepened, typically by tectonic activity, creating fluvial disequilibrium and another period of landscape degradation.

One assumption behind the AST model is that following a period of tectonically driven degradation, the initial increase in accommodation will be slow and will increase through time allowing increased preservation of thicker, finer-grained units. The observed decrease in fluvial energy occurs because 1) locally the fluvial system is no longer constrained by the paleovalley topography and is able to migrate across the landscape allowing finer-grained floodplain deposition; and 2) on a regional scale, a
basin that receives ample sediment, while eroding its sediment source, will fill its accommodation and decrease the slope of the landscape as it works to reach fluvial equilibrium.

One problem with this model is that it predicts that HAST units will be thicker than LAST units as accommodation increases. In theory this is correct, however this also assumes that the sediment transport mechanism (fluvial system) will be able to adjust to an increased lateral floodplain extent (constrained valley to entire basin). If sediment flux does not change and the area of effective deposition greatly increases, an increase in unit thickness may be difficult.

<table>
<thead>
<tr>
<th>Systems tract Features</th>
<th>Low accommodation systems tract (LAST)</th>
<th>High accommodation systems tract (HAST)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depositional trend</td>
<td>early progradational</td>
<td>Aggradational</td>
</tr>
<tr>
<td>Depositional energy</td>
<td>early increase, then decline</td>
<td>decline through time</td>
</tr>
<tr>
<td>Grading</td>
<td>coarsening-upward at base</td>
<td>fining-upward</td>
</tr>
<tr>
<td>Grain size</td>
<td>Coarser</td>
<td>Finer</td>
</tr>
<tr>
<td>Geometry</td>
<td>Irregular, discontinuous</td>
<td>tabular or wedge-shaped</td>
</tr>
<tr>
<td>Sand:mud ratio</td>
<td>High</td>
<td>Low</td>
</tr>
<tr>
<td>Reservoir architecture</td>
<td>amalgamated channel fills</td>
<td>isolated ribbon sandstones</td>
</tr>
<tr>
<td>Floodplain facies</td>
<td>Sparse</td>
<td>Abundant</td>
</tr>
<tr>
<td>Thickness</td>
<td>tends to be thinner</td>
<td>tends to be thicker</td>
</tr>
<tr>
<td>Coal seams</td>
<td>minor or absent</td>
<td>well developed</td>
</tr>
<tr>
<td>Paleosols</td>
<td>well developed</td>
<td>poorly developed</td>
</tr>
</tbody>
</table>

Figure 4.2-4  Depositional characteristics of low- and high- AST (Catuneanu, 2006).

Well-developed paleosols are also chronostratigraphically significant within the Chinle. Paleosols form in strata that have extended periods of surface exposure with no significant active deposition or erosion (i.e. the margins of the fluvial systems). Paleosol development may occur during periods of both regional landscape aggradation and degradation as long as the paleosol surface is not dramatically influenced. The period for
potential pedogenesis begins after deposition or exposure of the parent material and ends as the paleosol surface is buried by subsequent deposition. Mechanisms for pedogenesis include biogenic and/or climatic influences. The degree and depth of pedogenesis is a first-order indicator as to how long a surface was exposed. In the study area, evidence of mature pedogenesis occurs in both the Moenkopi and Gartra/mottled member strata as described in section 3.1.2.

Additional discussion and application of current continental sequence stratigraphic concepts are found in Beer's (2005) research of the Chinle Formation in South Central Utah.

4.3 Interpretation

In this section, depositional facies are identified, the stratal architecture is described, and a sequence stratigraphic framework is applied to the Chinle Formation. Two depositional sequences and 3 unconformities are identified within the DNM region Upper Triassic strata. The lower depositional sequence includes the Gartra and mottled members and the upper sequence contains the ocher member and the upper member sandstone and siltstone deposits.

Tr-3 erosional unconformity

During the Middle Triassic a major base level shift in the Colorado Plateau region resulted in a regional depositional hiatus and degradation of Early Triassic Moenkopi strata. This created the unconformable contact between Lower and Upper Triassic strata that is regionally identified between truncated and altered Moenkopi strata and onlapping and overlapping Chinle strata as described in section 3.1.1. The depth of incision into the Moenkopi was variable. Individual paleovalleys may be kilometers across and range from
0-40 m deep. Combined, these incisions created a SE to NW oriented paleovalley network 200-300 kilometers across (Stewart et al., 1972). The degree of Moenkopi pedogenesis is directly associated with its location on the paleovalley slope. At the base of the paleovalley, altered strata that would have been present were eroded by fluvial entrenchment. What remains is a thin layer of bleached Moenkopi, which may have been altered after Gartra deposition due to groundwater gleying processes. On the paleovalley margins however, paleosols may be well developed. The higher degree of pedogenesis occurred due to longer periods of surface exposure, possibly beginning during the Tr-3 degradation and ending with Chinle strata overlapping the paleovalley margins.

**Lower Chinle Formation Members - Dinosaur National Monument, UT**

![Figure 4.3-1 Cartoon of lower Chinle members onlapping and overtopping the Tr-3 paleovalley margins. The interpretation is based off extensive outcrop exposure in Bourdette Draw (Appendix A: 24). The paleovalley system runs generally E-W and the cross-section has a S-N bearing.](image)

**Lower Chinle Strata: Gartra Member and mottled member**

Lower Chinle units include the Gartra Member and the mottled member and are found within and overtopping the incised Tr-3 paleovalley network (Figure 4.3-1). Lower Chinle units evolve upsection from low-sinuosity, coarse-grained, amalgamated, Gartra fluvial deposits into silty mottled member floodplain deposits, as described in 3.1.2. To better define the character of the Gartra fluvial sandstone deposits, I have separated them into Gartra 1 (G1), Gartra 2 (G2), and Gartra 3 (G3) units.
The basal G1 unit is only found at the base of the deepest paleovalleys and represents the earliest Chinle deposition. The G1 is composed of cross-bedded conglomeratic sandstones with cobbles at the base of the channel deposits and no fine-grain material (Figure 3.1.2-1). G1 depositional environments are interpreted to be high-energy, low-sinuosity fluvial systems that were constrained to the paleovalley network. During the wet monsoonal seasons, water and sediment would be directed into the paleovalley network, the swollen river systems would partially rework previously deposited fluvial sediment, and the fine-grained material would be transported out of the basin preserving only coarse-grained deposits at the base of the valley.

G2 deposits are found above and partially truncating both G1 deposits and Moenkopi strata (Figure 4.3.1). Where the edge of G2 deposition is observed, it is onlapping the paleovalley wall (Appendix A: 24). In comparison to the G1 conglomeratic sandstones, the G2 sandstone is finer grained and contains some thin, silty, interbedded floodplain deposits, mottling, and pedogenesis. The deposition and preservation of finer grained sediment indicates that the G2 low-sinuosity fluvial systems experienced a decrease in energy and sediment reworking and an increase in sinuosity.

G3 high-sinuosity, fluvial channel deposits are thinner (1-2 meter), laterally restricted, and found interbedded within mottled member floodplain deposits. The mottled member is found overlapping G2 units and any remaining exposed Moenkopi surfaces (Figure 4.3.1; Appendix A: 24). Compared to G1 and G2 deposits, the mottled member unit is relatively thin, however it is found throughout the study area onlapping and overlapping the paleovalley margins. The mottled member is composed of pedogenically modified, mottled (typically purple), sandy siltstone floodplain deposits.
At the top of the mottled member is a mature paleosol unit. This well-developed paleosol marks the second sequence boundary and the end of the first depositional sequence.

In Carroll and Bohacs (1999) model, the lower Chinle depositional environments would have represented a water and sediment charged fluvial system able to consistently transport water and sediment out of the basin.

**mottled member non-depositional unconformity**

The second sequence boundary is represented in outcrop by a well-developed paleosol found at the top of the mottled member (Section 3.1.2). This paleosol exhibits mature pedogenesis indicating an extended period of surficial exposure with little to no aggradation or degradation. This paleosol horizon is clearly seen in the Utah DNM outcrops and in Bourdette Draw (Appendix A: 24); however, poor outcrop exposure make broader regional correlations of this soil horizon difficult.

**Upper Chinle Strata: ocher member and upper member**

In the distal western half of the northern Chinle basin, ocher member strata (Section 3.1.3) are found overlying the mottled member paleosols. The ocher member is a yellow to red colored, massive to thin-bedded siltstone. Upsection grain-size appears to gradually coarsen upward indicating an increasing sediment flux into the basin or progradation during basin fill. Within the unit are horizons of carbonate and analcime nodules, shrink-swell slickensides, and *scoyen"ia* beetle burrows. The presence of these features indicates that the ocher member was deposited in an underfilled lacustrine basin (Carroll and Bohacs, 1999) with fluctuating water levels and periodic saline-alkaline water conditions. The deposition of these units would have occurred in both lacustrine and marginal-lacustrine settings.
Upper member sandstones and conglomerates overlie, and in some locations partially truncate, the ocher member siltstones. The composition of the initial upper member varies between thick amalgamated cross-bedded sandstones, thinner sandstones with interbedded siltstones, and rarely, basal, laterally restricted, rip-up clast conglomerates (Section 3.1.4). Given the upper members interpreted depositional environment, a prograding fluvial-delta system, unit characteristics likely depend on the deltas proximity to the fluvial sediment source and lacustrine shoreline. Further up section, overall grain-size decreases and interbedded sandstone and siltstone beds become more prevalent. In Carroll and Bohacs (1999) model, the deposition of large fluvial-lacustrine deltas marks the transition from an underfilled basin to a balance-filled basin.

Above the fluvial-deltaic deposits are upper member lacustrine siltstones (Sections 3.1.4). The upper member lacustrine deposits are composed of horizontally laminated, weak-red and light green grey mudstones, siltstones and very-fine sheet delta sandstones. Also present are soft-sediment deformations and mudcrack features. This unit is interpreted to be a shallow, fresh-water, balance-filled lacustrine basin (Carroll and Bohacs, 1999). The upper member siltstones are the final preserved Chinle deposit in the study area; however, the siltstones are also eroded away in some locations. All subsequent deposits have been eroded by the J-0 erosional unconformity.

Within the Eagle Basin, upper Chinle strata may be much thicker (< 300m) and is composed entirely of red siltstone member paleosol, floodplain, and eolian deposits.

**J-0 erosional unconformity**

The J-0 erosional unconformity marks a period of Upper Triassic stratal degradation (Section 3.1.6). Above the erosional surface are Jurassic eolian sandstones.
5.1 Boundary Conditions and Depositional Model

When trying to determine the depositional history of a basin, in the spirit of Ockham’s razor, the simplest model with the fewest boundary conditions is a logical starting point. To explain the changing Chinle depositional facies in the northern basin, a model was created using the simplest boundary conditions. These boundary conditions are listed below and are graphed in Figure 5.1-1;

1. There was constant rate of basin subsidence throughout Late Triassic time.
2. The regional climate became gradually dryer throughout the Late Triassic (Dubiel et al., 1991).
3. Sediment flux entering the distal basin decreased as the climate became dryer and the Ancestral Rockies were onlapped and overlapped by Chinle Strata (Dubiel et al., 1991).
4. A forebulge, a topographic high related to subduction (Section 2.1), existed on the western margin of the depositional basin and experienced little to no subsidence in relation to the basin (Lawton, 1994).

Boundary conditions 2-4 are based on observations made by previous researchers. Due to the lack of a clear understanding of the Late Triassic tectonic conditions, a constant rate of basin subsidence (boundary condition 1) was used to provide the simplest depositional scenario.

Based on previous sequence stratigraphic models (Caroll and Bohacs, 1999; Lane, 1955) and the stated boundary conditions, a hypothetical succession of depositional
environments can be predicted for the basin. Initially, the seasonally high volumes of mega-monsoonal precipitation would enable the transport of high volumes of sediment from the Ancestral Rockies into the basin. The high volume of sediment would fill all accommodation created by the regional subsidence and the excess sediment would be transported past the forebulge and out of the basin. In Caroll and Bohacs (1999) model, this would be a fluvial to overfilled-basin environment. Through time however, as sediment flux through the basin decreases with precipitation, sedimentation would no longer be able to completely fill accommodation created by subsidence and excess accommodation would be created. As accommodation increased, a balance-filled or underfilled lacustrine basin would form behind the forebulge/sill. Figure 5.1-1 illustrates these simplified boundary conditions and the predicted effects on the depositional systems. This depositional pattern however it is not what is seen in the study area strata.

Figure 5.1-1  Model boundary conditions for the northern Chinle basin deposition.
Figure 5.1-2  Interpreted boundary conditions for the Late Triassic Chinle deposition in the DNM region.

Table 5.1-1  Definitions used in depositional boundary condition models.

Accommodation:  Relative amount of space available for sedimentation.
Excess accommodation:  Accommodation > sediment flux to study area
Excess sediment flux:  Sediment flux to study area > accommodation
Precipitation:  Relative volume of precipitation falling in the depositional basin.
Sediment flux:  Relative volume of sediment reaching the DNM study area from the Ancestral Rocky Mountain sediment source.
Subsidence:  Rate of subsidence of the depositional basin.

Figure 5.1-2 represents the interpreted Late Triassic boundary conditions for the DNM region that are based on the field work completed for this study. In the field, changes in the sedimentary facies were identified that were not predicted in the general boundary condition model. These anomalies indicate shifts in sediment flux and/or
accommodation in the DNM region. The mechanism I propose to explain these facies shifts is autogenic annealing of topography. Annealing of topography is the process of smoothing uneven erosional topography through subsequent sediment aggradation. This autogenic process may explain both local and basin-scale facies changes.

The concept of facies changes resulting from annealing topography is unique from other continental sequence stratigraphic models (e.g., Carol and Bohacs (1992) and Catuneanu (2006)) in that it does not rely solely upon allogenic tectonic or climatic change as the forcing mechanisms. Autogenic stratigraphic changes for basins with constant allogenic boundary conditions are numerically modeled and discussed in Muto et al. (2007).

5.2 Depositional History

The following sections describe the depositional history of the Chinle Formation in the DNM region and the interpreted mechanisms behind the changing facies.

5.2.1 Landscape Degradation 1

During the mid-Triassic, the western Pangaeaan margin experienced uplift, probably related to the subduction along the western margin of Pangaea, which caused the incision of large paleovalley networks into Early Triassic Moenkopi strata (Appendix B: 02). These paleovalley networks created relatively direct conduits for water and sediment transport through the Chinle basin into the western back-arc basin. In outcrop, this period of degradation is recorded as the Tr-3 unconformity. Another product of the oceanic plate subduction was the development of a dynamic forebulge (Section 2.1) located between the back-arc basin and the continental interior.
5.2.2 Landscape Aggradation

At some point between the Mid-Triassic and the Late Triassic, basin uplift ceased and subsidence began. As a result of this change, accommodation was now being created and the fluvial systems changed from degradational to aggradational. Initial accommodation and aggradation would have developed at the base of the laterally constrained paleovalley network. Within this network, high volumes of water and sediment would be seasonally focused in high-energy, low-sinuosity fluvial systems (Appendix B: 03). These high-energy seasonal flooding events were able to rework previously deposited fluvial material leaving only coarse-grained Gartra conglomeratic sandstones behind. Partial-erosion of the paleovalley walls also occurred however induration horizons developed on the exposed Moenkopi paleovalley surfaces may have inhibited this degradation (Nanson et al., 1995). The constant creation of new accommodation and the decrease in fluvial energy may have also inhibited the erosion of the paleovalley network. As aggradation in the paleovalley and partial-erosion of the valley margins continued, the increased distance between the paleovalley walls would have increased the area of the floodplain. This would result in a decrease in the degree of lateral constraint, an increase in the net accommodation, and a decrease in the effective fluvial energy. Within these upper Gartra (G2) deposits, the decrease in fluvial energy and increase in sinuosity resulted in a decrease in overall grain-size and increase in the preservation of sandstone mottling and silty interfluve deposits. As aggradation continued, paleovalley margins were breached and the fluvial system became laterally unconfined.
In the DNM region, the loss of lateral confinement caused the fluvial system to change into a high-sinuosity, low-energy fluvial system able to migrate, avulse, and deposit sediment across the entire landscape. These lower energy fluvial environments experienced a decrease in vertical aggradation rate and were preserved as thin-bedded mottled member floodplain siltstones with small interbedded channel sandstones (Appendix B: 04). Another local result of this fluvial system change is an apparent increase in paleosol development due to extended periods of surficial soil exposure and pedogenesis.

On a regional, basin-wide scale, the filling of the paleovalley network dramatically increased the basin’s available net accommodation laterally and decreased the efficiency of sediment transport through the basin (Figure 5.1-2). This triggered the autogenic retreat of the basins depositional systems and effectively created a sediment trap next to the sediment source in the Eagle basin. Within the Eagle Basin, red siltstone member low-energy, high-sinuosity fluvial systems developed into a sediment-rich fluvial mega-fan that would gradually prograde across the northern depositional basin. Conversely, while the mega-fan was forming in the proximal basin, the distal basin became sediment starved and experienced a depositional hiatus that allowed the mottled member well-developed paleosols to form.

5.2.3 Depositional Hiatus

During this depositional hiatus, subsidence in the sediment starved distal basin created excess accommodation. Within this increasing accommodation, the ocher member lacustrine system developed behind the forebulge, which acted as a sill (Appendix B: 05). Detailed facies analysis (sections 2.4.3 and 3.1.3) indicate that the
lacustrine system was shallow, evaporative and at least periodically saline, all of which are indicative of underfilled lacustrine basin conditions and a closed hydrologic system with net evapotranspiration. As subsidence continued without an increase in sediment flux to the distal basin, ocher member strata transgressed landward towards the Eagle Basin and the prograding fluvial mega-fan.

During deposition of the ocher member in the DNM region, the fluvial mega-fan, consisting low-energy, high-sinuosity, red siltstone member floodplain deposits, prograded out of the Eagle Basin. When this sediment charged fluvial system reached the excess accommodation in the lacustrine environment, upper member lacustrine-deltaic deposits began to fill the excess accommodation (Appendix B: 06). These initial upper member deposits were thick, cross-bedded, deltaic deposits. As the lacustrine accommodation filled, thinner fluvial-deltaic sandstones and siltstones followed as westward progradation of the mega-fan continued. Progradation of the upper member strata would have continued until all of the excess accommodation behind the forebulge/sill was filled and the fluvial system reached equilibrium.

Following upper member fluvial-deltaic deposition, a second lacustrine transgression of the basin occurred. The combination of continued subsidence and a decrease sediment flux to the distal portions of the basin, due to a dryer climate, again enable the creation of excess accommodation and the formation of a second lacustrine system behind the forebulge (Appendix B:07). This upper member lacustrine system transgressed landward into the DNM region; however, the transgression did not reach the Eagle basin. The lack of evaporite deposits and major carbonate beds indicate that this lacustrine system was a fresh water, balance-filled basin with an open hydrologic system.
5.2.4 Landscape Degradation 2

Within the DNM study region, all strata above the upper member have been truncated by the J-0 unconformity, which is overlain by Jurassic eolian deposits (Appendix B: 08).

5.3 Alternative Explanations

By using a model with simplified boundary conditions, a gradually drying climate and constant basin subsidence, the observed Chinle facies changes can be explained by the autogenic annealing of topography. However, the same facies changes may have also developed due to unknown fluctuations in either of these allogenic influences. While the Late Triassic climatic conditions have been well documented (section 2.3) if a more accurate depositional model is to be created, further work on the basins tectonic history must be done.

Also, another possible influence in the development of the upper member lacustrine system could be increased accommodation caused by increased isostatic flexure and compaction resulting from rapid upper member sandstone/fluvial mega-fan sediment loading. A rapid increase in sediment load could cause compaction of previously deposited lacustrine mudstones and increase the rate of subsidence. An increase in accommodation could result in the landward transgression of the upper member lacustrine environments. If this mechanism is to be used, further work on actual degree of basin change due to loading must be completed.

5.4 Regional Chinle Nomenclature

In most cases, the previously defined nomenclature and unit descriptions (Stewart et al., 1972; Dubiel, 1992, 1994) for the Chinle Formation are clear. There are however
couple of names, contacts, and definitions that could be made more clear based on analysis completed in this study.

First, I would place the top of the mottled member at the well-developed paleosol horizon. This surface marks a period of regional non-deposition and a clear sequence boundary.

The boundary between the ocher member and overlying upper member should be identified at the point where there is a clear change from predominantly siltstone to sandstone deposition. In gamma ray well logs this contact appears gradual; however in outcrop and other well log surveys the contact is usually readily apparent.

Determining where red siltstone deposition ends and ocher and upper member deposition begins is also a difficult contact to identify due to their genetically related origins, similar lithology, and no definitive historical definition. I propose that whenever ocher member lacustrine facies are present, which are found throughout the distal basin, the overlying strata should be identified as the upper member. Wherever major lacustrine deposition is not found and lower Chinle strata are covered by fine-grained red floodplain deposits, primarily found in the proximal Eagle Basin, red siltstone member terminology should be used for the entire upper Chinle depositional sequence.
CHAPTER VI
CONCLUSIONS

By placing the Chinle Formation within a sequence stratigraphic model using simple but reasonable allogenic boundary conditions, the evolution of the basins depositional facies can be explained primarily using an autogenic mechanism. By annealing the pre-Chinle erosional paleovalley topography, the Chinle depositional systems alter the boundary conditions (sediment flux) acting upon the distal basin and force autogenic regional changes in sedimentation.

Initial changes are observed in the upsection increases in fluvial sinuosity and decreases in grain-size of Gartra Member units within the paleovalley. As the paleovalley margins were breached by mottled member floodplain deposits, the depositional systems were no longer laterally constrained and sedimentation occurred across the entire landscape. This change in sedimentation triggered an autogenic retreat of the depositional systems into the proximal basin, which left the distal basin sediment starved. Non-deposition in the distal basin allowed the well-developed mottled member paleosol to form. The combination of non-deposition and basin subsidence subsequently lead to excess accommodation and the ocher member lacustrine transgression. During this lacustrine transgression, a sediment-charged fluvial mega-fan prograded out of the proximal basin gradually filling the basins accommodation with upper member fluvial-deltaic deposits. A second, possibly climate related, lacustrine transgression followed depositing upper member weak red siltstones. Any subsequent Chinle deposition in the study area has been eroded by the J-0 unconformity.
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APPENDIX A

Measured Sections and Outcrop Photos
Figure 01: Field map depicting measured section and well-log locations. FG=Flaming Gorge, DF=Dry Fork Canyon, MN=Mine Property, RC=191 Road Cut, RF=Red Fleet State Park, RN=Racetrack Nose, RT=Racetrack, SS=Sounds of Silence, BL=Boat Launch, ES 1 & 2=Elder Spring 1 & 2, BI 1 & 2=Bourdette Island 1 & 2, BD 1, 2 & 3=Bourdette Draw 1, 2 & 3, SBD=South Bourdette Draw, SC=Skull Creek, DD=Disappointment Draw, DL=Deer Lodge Park, ##=Geophysical Well-Logs.
Figure 02
Measured Section Key

Header Key
Figure 00-----------------------------------------------Figure ##
Red Fleet State Park, Utah--------------------------------Outcrop name, state
6/03/05, 40.59791 N 109.44325 W----------------------Date measured, UTM location

Symbology Used in Measured Sections

**Sedimentology**
- Planar Bedding
- Current Ripples
- Trough Cross Stratification
- Rip-up Clasts
- Unconformity
- No exposure

**Paleontology**
- Crayfish Burrows
- Invertebrate Burrows
- Complete Bioturbation
- Root Traces
- Petrified Wood

**Paleopedology**
- Color Mottles
- Gleyed Horizon
- Cracks
- Slickensides
- Carbonate Nodules

*induration* Well-Developed Induration
Figure 03: Flaming Gorge (FG) Sheep Creek campground: (a) Chinle strata overlying Moenkopi strata that exhibit well-developed pedogenesis (d). Moenkopi alteration includes mineral and clay translocation, mottling, and bioturbation; (b) upper member sandstone with pelecypodichnus burrows; (c) upper member lacustrine siltstone that has a unique green-grey coloration.
Figure 04
Flaming Gorge NRA, Sheep Creek Campground, Utah
6/01/05, 40.92551N 109.70459W

<table>
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<tr>
<th>Lithostratigraphic Nomenclature</th>
<th>Chronostratigraphic Surfaces</th>
<th>Environment of Deposition</th>
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<td></td>
<td></td>
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<td>fluvial-deltaic</td>
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<td>Chuska Formation</td>
<td></td>
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<tr>
<td>ocher member</td>
<td></td>
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<td>shallow marine</td>
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Grain size ($\phi$)  
-6 < -4 < -2 < 0 < 2 < 4
Figure 05a
Dry Fork Canyon, NW of Vernal, Utah
6/05/05, 40.55829N 109.64486W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition
lacustrine

upper member

fluvial-deltaic

ocher member

lacustrine, lacustrine margin, and floodplain

mottled member and Gartra (G3)
mottled member paleosol?

floodplain and high sinousity fluvial

Gartra (G2)

low sinousity fluvial

Gartra (G1)

low sinosity fluvial

Moenkopi Formation

Tr-3 sequence boundary

shallow marine

Grain size (φ)
-6 < -4 < -2 < 0 < 2 > 4

cb < pb < f < s < sl
Figure 05b
Dry Fork Canyon, NW of Vernal, Utah
6/05/05, 40.55829N 109.64486W

Lithostratigraphic Chronostratigraphic Environment
Nomenclature Surfaces of Deposition

Nugget Formation 1-0 sequence boundary
- eolian ergs
- upper member
- lacustrine

Grain size (ø) -6 <4 <2 0 >2 >4

col psl ar cs fs silt
Figure 06
191 Mine Slope, North of Vernal, Utah
5/22/05, 40.59137N 109.48005W

Lithostratigraphic Nomenclature  Chronostratigraphic Surfaces  Environment of Deposition

Nugget Formation
- sequence boundary
- upper member
- eolian erg

Chinle Formation
- ochre member
- lacustrine and lacustrine margin

mottled member and Gartra (G3)
- mottled member
- paleosol?
- floodplain and high sinuosity fluvial

Gartra (G2)
- low sinuosity fluvial

Moenkopi Formation
- sequence boundary
- shallow marine

Grain size (φ)
-6  64  -4  cef  -2  phi  0  s  2  t  >4  silt

10m  20m  30m  40m  50m  60m

64
Figure 07: 191 Road Cut (RC) and Red Fleet State Park (RF): 191 roadcut is near the bottom of the photo and the Red Fleet measured section is seen in the distance (red line). Note the differences in Gartra thickness relating to depth of the incised paleovalley (~20m thick, ~5m thin).
Figure 08
191 Road Cut, Vernal, Utah
5/26/05, 40.58723N 109.46880W

Lithostratigraphic Chronostratigraphic
Nomenclature Surfaces Environment of Deposition

Nugget Formation sequence boundary eolian erg
upper member fluvial-deltaic

ocher member lacustrine and lacustrine margin

mottled member and Gartra (G3) mottled member paleosol?
flowplain and high sinuosity fluvial

Gartra (G2)
low sinuosity fluvial

Gartra (G1j)
low sinuosity fluvial

Moenkopi Formation sequence boundary shallow marine
Grain size (a) -6 <4 <2 0 2 4 silt clays}

---

90m
80m
70m
60m
50m
40m
30m
20m
10m
0m
Figure 09
Red Fleet State Park, Utah
6/03/05, 40.59791N 109.44325W

Lithostratigraphic Nomenclature  Chronostratigraphic Surfaces  Environment of Deposition

Nugget Formation

- sequence boundary
- eolian erg

- upper member
- fluvi-deltaic

Chinle Formation

- ocher member
- lacustrine and lacustrine margin

mottled member and Gartra (G3)
mottled member paleosol?
floodplain and high sinuousity fluvial

Gartra (G2)

low sinuousity fluvial

Moenkopi Formation

- sequence boundary
- shallow marine

Grain size (μ)
-16 -4 -2 0 2 >4

- silt
- sand
- gravel
- cobble
- pebble
- clay

90m
80m
70m
60m
50m
40m
30m
20m
10m
0m
Figure 10: Racetrack Nose (RN): (a) outcrop with thick Gartra and upper member sandstones; (b) ∼2 m. long fossilized log at the base of a Gartra channel; (c) siltstone beneath the Gartra channel exhibiting soft sediment deformation; (d) contact between ocher member lacustrine siltstone and upper member and fluvial-deltaic sandstone deposit; (e) pelecypodichnus trace fossils found near the top of the ocher member siltstone.
Figure 11a
Racetrack Nose, Dinosaur National Monument, Utah
6/10/05, 40.48330N 109.32846W

Lithostratigraphic
Nomenclature
Chronostratigraphic
Surfaces
Environment
of Deposition

upper member

fluvial-deltaic

ocher member

lacustrine and
floodplain

mottled member
and Gartra (G3)

mottled member
unconformity?

floodplain and
high sinuosity fluvial

Gartra (G2)

low sinouosity fluvial

sequence boundary

Moenkopi Formation

shallow marine

Grain size (a)

continued...
Figure 11b
Racetrack Nose, Dinosaur National Monument, Utah
6/10/05, 40.48330N 109.32846W

Nugget Formation sequence boundary

Chinle Fm. upper member

Grain size (Ø) -6 <4 <2 0 2 >4

eolian ergs lacustrine fluvial-deltaic

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition
Figure 12: Racetrack (RT): (a) upper member transition from thick red fluvial-deltaic sandstone to interbedded sandstone and siltstone to weak red lacustrine siltstone; (b) Gartra conglomeratic sandstone fluvial channel deposit; (c) mottled member unconformity. A well developed indurated paleosol; (d) ocher member slickenside fabric; (e) ocher member scoyenia beetle burrows.
Figure 13a
Racetrack (RT), Dinosaur National Monument, Utah
5/20/05, 40.44802N 109.29540W

Lithostratigraphic Nomenclature
Chemostratigraphic Surfaces
Environment of Deposition

upper member
fluvial-deltaic

ocher member
lacustrine and lacustrine margin

mottled member and Gartra (G2)
sequence boundary mottled member
unconformity floodplain and high sinuosity fluvial

Chinle Formation
Gartra (G2)
low sinuosity fluvial

Gartra (G1)
low sinuosity fluvial

Moenkopi Formation
Tr-3 sequence boundary
shallow marine
Grain size (σ)
Figure 13b
The Racetrack, Dinosaur National Monument, Utah
5/20/05, 40.44802N 109.29540W

Lithostratigraphic Chronostratigraphic Environment of Deposition Nomenclature Surfaces of Deposition

Nugget Formation eolian erg
J-0 sequence boundary
upper member
lacustrine
fluvial deltaic

Grain size (\(\sigma\))

\(-6\) \(<-4\) \(<-2\) \(0\) \(2\) \(4\) silt
Nugget Fm.
upper mbr. lacustrine
upper member fluvial-deltaic
ocher member

Figure 14: Upper member fluvial-deltaic sandstone and lacustrine siltstone at the Sounds of Silence (SS). Thick fluvial-deltaic sandstones (b) pinch out to the east (left) into thin interbedded sandstone and siltstones (a). Upper member weak red lacustrine beds overlie upper member sandstones.
Figure 15
Sounds of Silence Trail, Dinosaur National Monument, Utah
5/23/05, 40.44441N 109.27984W

Lithostratigraphic Nomenclature
Nugget Formation J-O sequence boundary eolian ergs

Chronostratigraphic Surfaces of Deposition

Environment of Deposition
lacustrine

Surfaces of Deposition
lacustrine and lacustrine margin

upper member fluvial-deltaic

ocher member

mottled member and Gartra (G3) mottled member unconformity? floodplain and high sinousity fluvial

Chinle Formation

Gartra (G2) low sinousity fluvial

Gartra (G1) low sinousity fluvial

Moenkopi Formation Tr-3 sequence boundary shallow marine

Grain size (\( \phi \))

\(-6 \ < -4 \ < -2 \ < 0 \ < 2 \ < 4 \)
Figure 16: Boat Launch (BL): (a) Chinle section on western side of the Green River; (b) Gartra Member fossil log (~1 ft. diameter); (c) upper member sandstone with silty interbeds; (d) upper member laminar bedding and climbing ripples; (e) contacts between the upper member fluvial-deltaic sandstone, upper member lacustrine siltstone, and Nugget Formation eolian sandstone.
Figure 17a
Split Mountain Boat Launch, Dinosaur National Monument, Utah
7/06/05, 40.44246N 109.24956W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition

upper member
fluvial-deltaic

ocher member
lacustrine and lacustrine margin

Chinle Formation
mottled member and Gartra (G3)
mottled member unconformity
floodplain and high sinosity fluvial

Gartra (G2)
low sinosity fluvial

Gartra (G1)
low sinosity fluvial

Moenkopi Formation
Tr-3 sequence boundary
shallow marine

Grain size (a) -5 <4 phi <2 phi 0 2 phi >4 phi silt

continued...
Figure 17b
Split Mountain Boat Launch, Dinosaur National Monument, Utah
7/06/05, 40.44246N 109.24956W

Lithostratigraphic 
Nomenclature
Chronostratigraphic 
Surfaces
Environment 
of Deposition

Nugget Formation
eolian erg
J-0 sequence boundary
lacustrine
fluvial-deltaic

Grain size (\(\sigma\))
\(-6 \rightarrow \ll 4 \rightarrow \ll 2 \rightarrow 0 \rightarrow \ll 2 \rightarrow \ll 4 \rightarrow \ll silt\)
Figure 18
Red Wash, Dinosaur National Monument, Utah
6/16/05, 40.54941N 109.12039W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition

Nugget Formation
- eolian ergs
  upper member

ocher member

Chinle Formation
- fluvial-deltaic
  mottled member
  and Gartra (G3)

 Gartra (G2)

Moenkopi Formation
- shallow marine
  Tr-3 sequence boundary
  Moenkopi Formation

Grain size (φ)
-6 < -4 < -2 < 0 > 2 > 4

Surfaces:
- J-O sequence boundary
- sequence boundary
- floodplain and high sinosity fluvial
- Moenkopi Formation
- Tr-3 sequence boundary
- shallow marine
Figure 19: Elder Spring (ES 1 & 2): (a) Google Earth image of western ES 2 outcrop; (b) cross-bedded Gartra (G1) conglomeratic sandstone; (c) well-developed mottled member paleosol with mature pedogenesis (mineral translocation and mottling, vertical white crayfish burrows, induration); (d) thin upper member sandstones with silty interbedding. The transition between upper mbr. fluvial-deltaic and lacustrine deposition can be identified by the color transition from reddish-orange to weak red; (e) scouenia beetle burrows in upper member. Pen for scale.
Figure 20
Elder Spring 1, Dinosaur National Monument, Utah
5/20/05, 40.42302N 109.18595W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition

--------- J-0 sequence boundary

lacustrine

upper member

fluvial-deltaic

Chinle Formation

ocher member

lacustrine and lacustrine margin

--------- sequence boundary

mottled member and Gartra (G3)
mottled member unconformity

mottled member and Gartra (G2)

Gartra (G2)

low sinousity fluvial

--------- sequence boundary

mottled member and Gartra (G1)

Gartra (G1)

low sinousity fluvial

Moenkopi Formation

Tr-3 sequence boundary

shallow marine

Grain size (φ) -6 -4 -2 0 2 4

col. silt cl. f. s."
Figure 21
Elder Spring 2, Dinosaur National Monument, Utah
6/29/05, 40.42880N 109.19041W

Lithostratigraphic Nomenclature

Nugget Formation


eolian erg

J-O sequence boundary

Upper member

fluvial-deltaic

ocher member

lacustrine and lacustrine margin

Chinle Formation

sequence boundary

mottled member and Gartra (G3)
mottled member unconformity

Gartra (G2)

low sinuousity fluvial

Gartra (G1)

low sinuousity fluvial

Moenkopi Formation

Tr-3 sequence boundary

shallow marine

Grain size (a)

-6 <2a1 <-2 peb <0 cs 2 fs >4 slt

90m

80m

70m

60m

50m

40m

30m

20m

10m

82
Figure 22a
North Bourdette Draw Island, Dinosaur National Monument, Utah
6/24/05, 40.40802N 109.17428W

<table>
<thead>
<tr>
<th>Lithostratigraphic Nomenclature</th>
<th>Chronostratigraphic Surfaces</th>
<th>Environment of Deposition</th>
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<tbody>
<tr>
<td>upper member</td>
<td>fluvial-deltaic</td>
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<tr>
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</tr>
<tr>
<td>ochre member</td>
<td>lacustrine and floodplain</td>
<td></td>
</tr>
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<td></td>
<td></td>
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<tr>
<td>Chinle Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>sequence boundary</td>
<td></td>
<td></td>
</tr>
<tr>
<td>mottled member and Gartra (G3)</td>
<td>mottled member</td>
<td></td>
</tr>
<tr>
<td>unconformity</td>
<td>high sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Gartra (G2)</td>
<td>low sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Moenkopi Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>sequence boundary</td>
<td>shallow marine</td>
<td></td>
</tr>
</tbody>
</table>

Grain size (φ): -6 >-4 >-2 0 2 4 shf

continued...
Figure 22b
North Bourdette Draw Island, Dinosaur National Monument, Utah
6/24/05, 40.40802N 109.17428W

Lithostratigraphic Chronostratigraphic Environment
Nomenclature Surfaces of Deposition

Nugget Formation
- sequence boundary
  eolian ergs

Chinle Formation
  upper member
  lacustrine
  fluvial-deltaic

Grain size (ø) -6 <-4 <-2 <0 2 >4
cal  silt
Figure 23
North Bourdette Island 2, Dinosaur National Monument, Utah
6/26/05, 40.40719N 109.16762W

<table>
<thead>
<tr>
<th>Lithostratigraphic Nomenclature</th>
<th>Chronostratigraphic Surfaces</th>
<th>Environment of Deposition</th>
</tr>
</thead>
<tbody>
<tr>
<td>ocher member</td>
<td>sequence boundary</td>
<td>lacustrine &amp; lacustrine margin</td>
</tr>
<tr>
<td>mottled member</td>
<td>mottled member unconformity</td>
<td>induration</td>
</tr>
<tr>
<td>Gartra (G3)</td>
<td>soil surface</td>
<td>high sinosity fluvial</td>
</tr>
<tr>
<td>mottled member</td>
<td>floodplain</td>
<td></td>
</tr>
<tr>
<td>Gartra (G2 or G3)</td>
<td>low sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Chinle Formation</td>
<td>low sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Gartra (G2)</td>
<td>low sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Gartra (G1)</td>
<td>low sinosity fluvial</td>
<td></td>
</tr>
<tr>
<td>Moenkopi Formation</td>
<td>Tr-3 sequence boundary</td>
<td>shallow marine</td>
</tr>
</tbody>
</table>

Grain size (φ) -6 <-4 <-2 <-1 0 2 4 >4 silt
Figure 24a: North Bourdette Draw (upper image ~.8 miles): The Chinle Formation is found between the red Moenkopi strata and the thick, cliff-forming, Jurassic Nugget Sandstones at the top of the outcrop. The lower image shows a paleovalley margin where Gartra fluvial deposits onlap and mottled member floodplain deposits overlap Moenkopi strata. See AI 24b for a close up of a Moenkopi-mottled member contact. Red lines represent NBD measured section locations.
Figure 24b: Chinle mottled member strata truncating, onlapping, and overlapping pedogenically modified Moenkopi strata found below the Tr-3 unconformity. In the upper image, high-sinuosity G3 fluvial channel deposits are truncating and onlapping pedogenically modified Moenkopi strata. Across the top of both the G3 and Moenkopi units are silty mottled member floodplain deposits. At the top of the mottled member is an indurated, well-developed, heavily bioturbated paleosol horizon. This horizon represents an extended period of surface exposure with little to no aggradation or degradation.
Figure 25
North Bourdette Draw 1, Utah
5/24/05, 40.39441N 109.18873W

<table>
<thead>
<tr>
<th>Lithostratigraphic Nomenclature</th>
<th>Chronostratigraphic Surfaces</th>
<th>Environment of Deposition</th>
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</thead>
<tbody>
<tr>
<td>Nugget Formation</td>
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<td>eolian erg</td>
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<tr>
<td>upper member</td>
<td></td>
<td>lacustrine margin</td>
</tr>
<tr>
<td>Chuska Formation</td>
<td></td>
<td>lacustrine</td>
</tr>
<tr>
<td>ocher member</td>
<td></td>
<td>lacustrine margin</td>
</tr>
<tr>
<td>Mottled member</td>
<td>sequence boundary</td>
<td>soil surface</td>
</tr>
<tr>
<td>Mottled member</td>
<td>unconformity</td>
<td>Garfield (G3)</td>
</tr>
<tr>
<td>Mottled member</td>
<td>Tr-3 sequence boundary</td>
<td>floodplain</td>
</tr>
<tr>
<td>Mottled member</td>
<td></td>
<td>shallow marine</td>
</tr>
</tbody>
</table>

Grain size ($\phi$) -6 <0 <2 0 2 3 >5
Figure 26
North Bourdette Draw 2, Utah
6/25/05, 40.39052N 109.19067W

Lithostratigraphic Chronostratigraphic Environment
Nomenclature Surfaces of Deposition

- ocher member
- mottled member
- sequence boundary
- mottled member
- unconformity
- soil surface
- lacustrine & lacustrine margin
- induration
- floodplain
- Garbra (G3)
- Moenkopi Formation
- Tr-3 sequence boundary
- shallow marine
- high sinousity fluvial

Grain size (\(\phi\))
- \(-6\) to \(<-4\)
- \(<-2\)
- \(0\)
- \(2\)
- \(>4\)

(0) -6
<4
cel
<2
pl
0
fl
<2
sls
>4
sill
Figure 27
North Bourdette Draw 3, Utah
6/25/05, 40.38719N 109.19150W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition

ocher member
mottled member
mottled member
mottled member
Gartra (G2 or G3)

Tr-3 sequence boundary
mottled member
unconformity
mottled member

Moenkopi Formation

Tr-3 sequence boundary

soil surface
induration
floodplain
high sinosity fluvial
shallow marine

Grain size (φ) -6 <4 <2 0 2 >4
Figure 28: South Bourdette Draw (SBD): (a) Chinle section with basal mottled member floodplain and G3 channel deposition overlying modified Moenkopi strata; (b) mottled member siltstone exhibiting vertic shrink-and-swell and slickenside textures; (c) pedogenically well-developed, indurated, paleosol horizon at the top of the mottled member. This horizon may be traced (N-S) along the Bourdette Draw outcrops (~8 miles).
Nugget Formation

Upper member

Chinle Formation

Ocher member

Mottled member

Mottled member and Gartra (G3)

Moenkopi Formation

Tr-3 sequence boundary

Pedogenically modified Moenkopi

Shallow marine

Grain size (φ)

-6 <4 ccl <2 <0.5 0 2 f< 4 silt
Figure 30: Skull Creek (SC): (a) outcrop near measured section. Thin Gartra (G3) unit below mottled member; (b) and (c) indurated, cherty, paleosol horizon at top of mottled member. The horizon may be correlative to well-developed mottled member paleosols in Utah (Bourdette Draw, Racetrack...).
Figure 31
Skull Creek, Colorado
6/17/05, 40.29150N 108.70251W

Lithostratigraphic Chronostratigraphic Environment Nomenclature Surfaces of Deposition

Nugget Formation sequence boundary eolian ergs
upper member fluvial-deltaic

Chinle Formation
	ochrome member lacustrine, lacustrine margin, and floodplain

Mammite member mottled member and Gartra (G3) sequence boundary mottled member unconformity fluvial-deltaic and high sinuosity fluvial

Gartra (G2) mottled member and Gartra (G3) sequence boundary mottled member unconformity fluvial-deltaic and high sinuosity fluvial

Moenkopi Formation sequence boundary shallow marine

Grain size (\(\phi\))

\[
\begin{align*}
-6 & \quad <4 \\
<2 & \quad 0 \\
2 & \quad >4 \\
\end{align*}
\]
Figure 32
Disappointment Draw, Colorado
7/06/05, 40.38317N 108.55809W

Lithostratigraphic Chronostratigraphic Environment
Nomenclature Surfaces of Deposition

Nugget Formation

--- J-O sequence boundary ---

Chinle Formation

upper member

ocher member

Moenkopi Formation

mottled member and Gartra (G3)
mottled member paleosol
M3-3 sequence boundary
shallow marine

Grain size (\(\phi\))
\([-6 < -4 < -2 < 0 \leq 2 \leq > 4\)]

lacustrine margin & fluvial-deltaic
fluvial-deltaic
lacustrine margin & floodplain
lacustrine and floodplain
floodplain and high sinuousity fluvial
Figure 33: Deer Lodge Park (DL) outcrop across Yampa River from measured section. Note ~10m weak red siltstone between 2 upper member sandstone packages.
Figure 34
Deer Lodge Park, Dinosaur National Monument, Colorado
6/07/05, 40.44985N 108.54311W

Lithostratigraphic Nomenclature
Chronostratigraphic Surfaces
Environment of Deposition

Nugget Formation
J-0 sequence boundary
eolian ergs
lacustrine

Chinle Formation
fluvial-deltaic

Moenkopi Formation
shallow marine

Grain size (ø)
### Figure 35

#### Measured Section Locations

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<tr>
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<th>Date</th>
<th>Measured Unit Thickness - meters</th>
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<td>Total Gartra 3-mm</td>
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<td>191 Mine</td>
<td>N. Vernal</td>
<td>40.59137N, 109.48005W</td>
<td>5/22/2006</td>
<td>63  15  11  28  9  -</td>
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<tr>
<td>RC</td>
<td>Hwy 191 Road Cut</td>
<td>N. Vernal</td>
<td>40.58723N, 109.46880W</td>
<td>5/18, 26/06</td>
<td>63  15  7  30  11  -</td>
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<tr>
<td>RT</td>
<td>Race Track</td>
<td>DNM Racetrack</td>
<td>40.44802N, 109.29540W</td>
<td>5/20-21/06</td>
<td>109 35  10  32  17  8</td>
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<tr>
<td>RW</td>
<td>Red Wash</td>
<td>N. DNM, UT</td>
<td>40.54941N, 109.12039W</td>
<td>6/16/2006</td>
<td>52  16  4  21  11  -</td>
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<tr>
<td>ES1</td>
<td>Elder Spring 1</td>
<td>DNM Racetrack</td>
<td>40.42302N, 109.18595W</td>
<td>5/19-20/06</td>
<td>85  15  13  34  12  11</td>
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<td>Bourdette Draw</td>
<td>Bourdette Draw</td>
<td>40.40719N, 109.16762W</td>
<td>6/20/2006</td>
<td>34* 30  4* - - -</td>
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<td>BD1</td>
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<td>Bourdette Draw</td>
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<td>Bourdette Draw</td>
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<td>Bourdette Draw</td>
<td>40.38719N, 109.19150W</td>
<td>6/25/2006</td>
<td>13* 7  5* - - -</td>
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<tr>
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<td>Skull Creek</td>
<td>Colorado Sections</td>
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<td>6/17/2006</td>
<td>69  - 10  25  34  -</td>
</tr>
</tbody>
</table>

**Unit abbreviations:**
- G3-mm = Gartra 3 and mottled member
- om = ochre member
- ums = upper member sandstone
- uml = upper member lacustrine

* incomplete measured section
## Figure 36a

### Well-Log Locations

<table>
<thead>
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<th>#</th>
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Unit abbreviations: G3-mm=Gartra 3 and mottled member; om=ocher member; ums=upper member sandstone; uml=upper member lacustrine.
### Figure 36b

#### Well-Log Locations

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Unit abbreviations: **G3-mm**=Gartra 3 and mottled member; **om**=ocher member; **ums**=upper member sandstone; **uml**=upper member lacustrine.
APPENDIX B

Depositional Environments and Boundary Conditions
Figure 01: Key to Paleoenvironmental Interpretations

KEY: Figures represent a period of Late Triassic deposition. Upper figure is a map view of the Northern Chinle basin with interpreted depositional environments. Also shown is the study area and locations of the cross-sections (A-A'; B-B'). Cross-section A-A' represents predicted basin deposition. B-B' represents study area deposition. Graph represents relative depositional conditions in the study area (Chapter 5.1).
Tr-3 Unconformity: During the Mid-Triassic, mega-monsoonal climatic conditions and tectonic uplift caused major landscape degradation and incision of a large paleovalley network into the Moenkopi Formation. The excess sediment flux would have been able to bypass the forebulge and enter the back-arc basin. Pedogenesis of Moenkopi strata would have begun on the valley margins.
Gartra Member: The change from basin uplift to subsidence allows for the creation of accommodation at the bottom of the paleovalleys. Accommodation is filled by the sediment flux through the valleys. The seasonal influx of precipitation causes previously deposited sediment to be reworked leaving only coarse grained units. The reworking of sediment decreases as the valley fills and the floodplain widens.
mottled member and red siltstone member: As the paleovalleys were breached, the fluvial systems were no longer constrained, which increased accommodation and prevented the new fluvial systems from transporting sediment to the distal basin. In the proximal basin a sediment-charged fluvial mega-fan developed (red siltstone mbr.). The distal basin became sediment starved and mottled member well-developed paleosols formed.
ocher member and red siltstone member: As the fluvial mega-fan continued to prograde out of the Eagle Basin, the distal basin remained sediment starved. This, along with continued subsidence, enabled excess accommodation to form behind the forebulge. In this space an underfilled lacustrine system developed and trangressed the basin landward. This lacustrine environment periodically dried out.
upper member: When the mega-fan reached the sediment starved lacustrine system, large fluvial-deltaic deposits filled the excess accommodation. As the mega-fan continued to prograde and fill accommodation, deposition graded into floodplain units.
upper member lacustrine and red siltstone: As the climate continued to become drier, sediment transport to the distal basin decreased creating excess accommodation and a balance-filled lacustrine environment. Red siltstone member deposition in the proximal basin continued.
J-0 Unconformity and Nugget Formation: Within the study region, an erosional unconformity is found incising into the upper member, sometimes completely eroding away upper member lacustrine deposition. Above the J-0 are thick eolian sandstones.
APPENDIX C

Annealing Topography and Autogenic Facies Change
Most continental sequence stratigraphic models (e.g., Carol and Bohacs, 1992; Catuneanu, 2006) use fluctuations in allogetic forcing mechanisms (tectonics and climate) to explain changes in depositional facies. We propose that the process of annealing (flattening topography through aggradation) erosional paleovalley topography in an environment with steady allogetic, or autogenic (Muto et al., 2007), forcing mechanisms may also cause facies changes.

By annealing paleovalley topography, depositional systems and their boundary conditions (area) evolve and force regional changes in sedimentation without the influence of any tectonic or climatic change. Below is a model that uses simple autogenic boundary conditions to describe a series of facies changes that could develop on a large-scale eroded paleovalley landscape.

**Model Boundary Conditions:**

1. Regional pre-existing erosional paleovalley topography;
2. Constant rate of basin subsidence during deposition;
3. Constant climatic conditions;
4. Constant sediment flux entering the basin.

In order to develop the erosional topography required for this model, the regional base level must be lowered, possibly due to tectonic uplift of the basin, creating a degradational system. In a degradational system, there is vertical incision and lateral fluvial reworking and valley widening that results in eroded sediment being carried downstream. While still an overall degradational system, some fluvial deposition does occur and coarser grained material will be deposited in the proximal basin and finer grained material will be transported into and through the distal basin. In order to begin aggradation, boundary conditions must change (i.e. a shift from uplift to basin
subsidence) and it is from here that steady autogenic forcing mechanisms and the modeled deposition begins.

In an aggradational system, basin deposition either progrades from the proximal basin and/or transgresses from the distal basin. Within the paleovalley structure, sedimentation begins at its base and gradually onlaps the paleovalley walls. During the aggradation process, any landscape surface above the paleovalley floodplain remains degradational or non-depositional until it is onlapped or overlapped. With time and pedogenesis, these exposed surfaces may become indurated, or hardened, and be able to resist erosion. Early paleovalley fluvial systems (Appendix C: 1a) are laterally confined and will eventually transport nearly all of the basins water and sediment. These concentrated fluvial systems are able rework previously deposited sediment leaving locally homogenous deposits that become finer towards the distal basin. As aggradation within the paleovalley progresses (Appendix C: 1b), the distance between the paleovalley margins increases, due to valley geometry and fluvial erosion, which in turn increases the area of the floodplain. As floodplain area increases, the fluvial system becomes less constrained and its ability to rework previous deposition is diminished. This results in increased preservation of fine-grained sedimentation and floodplain pedogenesis.

As the pre-existing erosional topography is completely annealed (Appendix C: 1c), the area of the floodplain area dramatically increases which considerably decreases the basins “capture ratio” (see equation below). The capture ratio represents a basins ability to transport sediment through the basin. This effectively broader, flatter depositional basin forces fluvial systems to rapidly deposit their sediment load in the first available accommodation near the sediment source instead of transporting it through the
basin. This change in sedimentation triggers large-scale autogenic retreat (Muto et al., 2007) of the active depositional systems into the proximal basin. In the proximal basin, when excess accommodation is filled a prograding, sediment-charged, fluvial mega-fan develops.

As the fluvial mega-fan developed within the proximal basin, the distal basin is sediment starved and well-developed paleosol horizons may form on the exposed surfaces. This lack of sedimentation, along with continued subsidence, allows excess accommodation space to develop in the distal basin until the sediment-charged fluvial mega-fan, which originated in the proximal basin, progrades across the entire basin filling all excess accommodation as it advances.

\[
\text{Capture Ratio} = \frac{Q_s}{(\sigma \cdot A)};
\]
\[
Q_s = \text{sediment supply (m}^3/\text{a),}
\]
\[
\sigma = \text{subsidence rate (m/a),}
\]
\[
A = \text{area of depositional basin (m}^3)\]

(Kim et al., 2006)

Further topics that could be addressed in subsequent research include:

Is seasonal mega-monsoonal precipitation required to develop this facies evolution?

Is paleovalley induration required to preserve the paleovalley structure?

What scale of basin and paleovalley are required to develop these facies changes?

How does the rate of paleovalley aggradation impact sedimentation and facies characteristics?
Figure 01: Decreasing Lateral Fluvial Confinement and Related Autogenic Facies Change

a. Well-confined, low-sinuosity fluvial environment. Conglomeratic sandstone deposition

b. Moderately-confined, low-sinuosity fluvial environment. Sandstone deposition w/ small silty interbeds.

c. Unconfined, high-sinuosity fluvial/floodplain environment. Floodplain siltstones w/ small channel sandstones.