Structural And Stratigraphic Development Of A Salt Diapir Shoulder, Gypsum Valley, Colorado

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STRUCTURAL AND STRATIGRAPHIC DEVELOPMENT OF A SALT-DIAPIR SHOULDER, GYPSUM VALLEY, COLORADO

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To Marguerite McFarland Bradford, my grandmother, who has been a steady compass in my life and inspired me to pursue my passion of science.
Acknowledgements

This thesis would not have been accomplished without the people who had faith in the process, most notably my mother and father, Peggy and Richard McFarland, for their love and support, and their ability to provide a framework of advice and positive feedback during my time in El Paso. As parents, they taught me early life lessons of perseverance and toughness, which have suited me well for the challenges of scientific research. Secondly, the support from my sisters, Kate and Kelly. My inspiration to become a geologist was based mainly on the early trips in my life visiting my Aunt Boo and Uncle Mike in Idaho and Montana, where I developed a love for the big sky and the mountains, and wouldn’t be in this position if not for their inspiration of life in the west.

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Abstract

A salt shoulder is a low-angle segment of the salt-sediment interface where the margin of a passive diapir steps abruptly inboard. The Late Triassic Chinle Formation overlies caprock on the northeastern margin of Gypsum Valley (GV) at the salt-sediment interface in the natural amphitheater of Bridge Canyon. Modern erosion of the Dolores River has created 3D outcrop of the salt shoulder, expressed by a 500 meter wide, sub-horizontal platform that dips abruptly into the subsurface to the northeast toward the Dry Creek minibasin. Within Bridge Canyon, seven facies associations have been recognized and interpreted within the Chinle Formation: 1) non-caprock bearing channel-fill sandstone and stratified conglomerate (FA1), 2) caprock-bearing channel-fill sandstone and stratified conglomerate (FA2), 3) tabular shale, siltstone (FA3), 4) unsorted conglomerate lense (FA4), 5) caprock-bearing heterolithic channel sandstone and conglomerate (FA5), 6) fossiliferous mudstones & sandstones (FA6), 7) paleosols (FA7). The erosional contact between the Chinle Formation and caprock indicates the development of caprock prior to Chinle deposition. The distribution of facies associations transition from relatively coarser grained facies, including sheet-prone sandstone bodies of FA1 & FA2 and subordinate fine grained facies of FA3 on the outboard margin to isolated ribbon sand bodies of FA2 and FA5 and more common fine-grained facies of FA3 toward the inboard margin of the salt wall.

The Chinle Formation thins from approximately 160 meters at the outboard margin of the salt wall to 53 meters at the inboard margin. Three wedge halokinetic sequences onlap and overlap the shoulder forming an antiformal geometry of strata that are separated by low angle halokinetic sequence boundaries (2-3°). The lowest of the three halokinetic sequences (WHS-1) thins from 12.5 meters to pinchout over a distance of 200 meters, and is distinguished from the
overlying halokinetic sequence (WHS-2) by an angular unconformity of 2°. The WHS-2 reaches a maximum thickness of 27.5 meters and onlaps onto the shoulder over a distance of 325 meters, and is distinguished from the overlying halokinetic sequence by an angular unconformity of 3°. The third wedge halokinetic sequence reaches a maximum thickness of 120 meters and thins to 53 meters where it onlaps and overlaps onto the shoulder. The three wedge halokinetic sequences of the Chinle Formations stack into a tapered-CHS.

The evolution of the GV shoulder is described through the following sequence: (1) development of a Moenkopi-aged caprock on the GV salt wall through meteoric groundwater influx, (2) regional beveling at the end of Moenkopi/beginning of Chinle deposition, documented as the Tr-3 unconformity, (3) decrease in salt rise during Chinle deposition leading to progressive onlap and eventual overlap over the shoulder, (4) rotation of WHS-1 and WHS-2 by subsidence of the Dry Creek Minibasin flanking the still active northeast margin of the GV salt wall, evident by halokinetic sequence boundaries, (5) subsequent overlap by WHS-3 due to discontinued of the northeast margin of the GV salt wall created a drape-fold monocline overlying the shoulder, and (6) continued salt rise continued at an inboard position creating topography explaining the presence of the FA4 debris flows observed at the inboard margin of the salt wall in WHS-3. This timing has an effect that is two-fold: the outboard margin of the shoulder discontinues drape folding strata, as the drape folding zone transitions to the inboard margin of the salt wall, where salt continues to rise creating a new zone of drape folding. Chinle through Navajo strata share common high angle normal fault planes that collapsed strata overlying the shoulder that are dipping into the salt wall completing the antiformal geometry of the strata overlying the shoulder.
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Chapter 1: Introduction

A salt shoulder is a low-angle segment of the salt-sediment interface where the margin of a passive diapir steps abruptly inboard (Fig. 1a). Salt shoulders have been identified on seismic and well log datasets from salt basins in offshore Brazil (e.g. Campos and Santos Basins), the Pricaspian Basin, the North Sea, as well as deep-water Gulf of Mexico (Demercian et al., 1993). “Shoulder-roll trap” (Fig. 1b) refers to the antiformal geometry of strata overlying some salt shoulders that are potential hydrocarbon traps. Salt shoulders have been recognized in Precambrian outcrops of South Australia (Hearon et al., 2014). However, there are no studies to date that document the detailed sedimentologic, stratigraphic, and structural components of these features, which is needed to determine the timing and controls on their development. Previous studies have suggested that the relationship between sediment accumulation and rising salt is key to understand the formation of salt structures and their associated structural and stratigraphic architecture (Vendeville & Jackson, 1992; Jackson et al., 1994; Rowan et al., 2003; Giles & Lawton, 2002; Giles & Rowan, 2012). This study is the first detailed description of the structural and stratigraphic features of an outcropping salt shoulder and interpretation of the controls on shoulder development.

The Dolores River canyon of Southwest Colorado provides an exceptional, natural 3-D exposure of a salt shoulder on the northeast flank of the Gypsum Valley salt wall. The canyon exposures provide an opportunity to complete a thorough stratigraphic analysis of the Late Triassic Chinle Formation, where it onlaps the shoulder and overlies exposed caprock on the shoulder. The canyon also provides the possibility to re-create the three-dimensional geometry of the salt shoulder and the overlying stratigraphy from the top of the caprock surface to the base of the Wingate Formation. This study will aid in evaluation of other salt shoulder structures,
particularly in increasing the prediction potential of structural-stratigraphic trap geometries in salt shoulder structures, and increasing the understanding and predictability of reservoir distribution and architecture in fluvial strata affected by syn-depositional growth of passive salt diapirs.
Figure 1. Conceptual models that illustrate: (A) stratal geometries flanking and overlying a salt shoulder structure without a rollover anticline trap associated with it, and (B) stratal geometries flanking and overlying a salt shoulder with an associated antiformal trap bounded by post-depositional faults.
Chapter 2: Geologic Setting

The Paradox Basin (Fig. 2) is an asymmetric (190 km x 265 km) salt basin that developed along the southwestern flank of the Uncompahgre uplift in Utah and Colorado during the Pennsylvanian-Permian Ancestral Rocky Mountain orogeny (White and Jacobson, 1983; Huntoon et al., 2002; Barbeau, 2003). The Paradox Basin was one of a series of intracratonic flexural basins within the Ancestral Rocky Mountain foreland that developed as a result of loading of thick-skinned, basement-cored uplifts (Handschy and Dyer, 1987; Blakey and Knepp, 1989; Dickinson and Lawton, 2003; Trudgill, 2011). The northwest-southeast trending Uncompahgre Uplift, abuts the deepest part of the Paradox Basin, and is believed to have been the main proximal sediment source for the Pennsylvanian-Triassic sediments within the northeastern part of the Paradox Basin (Mack and Rasmussen, 1984). The basin subsided rapidly during Pennsylvanian time while being filled with ~2500 meters of cyclically deposited dolostone, evaporite, and shale of the Paradox Formation (Fig. 3) in the foredeep of the basin. Age-equivalent biothermal carbonates formed in the western distal or forebulge/backbulge margin of the basin (Hite, 1968; Peterson and Hite, 1969; Baars and Stevenson, 1981; Condon, 1997; Trudgill, 2010). In the center of the basin, the Paradox Formation is a thick (up to ~ 3 km) megasequence of 29 shale-dolomite-evaporite cyclothems (Peterson and Hite, 1969). The basin’s geographic boundaries are defined by the maximum depositional extent of Paradox Formation evaporitic facies (Condon, 1997).

During the latest Pennsylvanian and Early Permian, an alluvial wedge (Cutler Group) on the southwest margin of the Uncompahgre uplift prograded southwestward (Mack and Rasmussen, 1984; Barbeau, 2003), locally generating enough sediment overburden to cause differential loading of the underlying Paradox evaporite, which is thought to have triggered
diapirism shortly after and possibly during salt deposition (Trudgill and Paz, 2009). Salt movement resulted in formation of a series of elongate salt walls roughly parallel in the northern Paradox Basin (Fig. 2) that passively grew over approximately 75 million years (Late Pennsylvanian through Triassic) (Ohlen & McIntyre, 1965; Doelling, 1988; Trudgill et al., 2004). Individual Triassic and Jurassic units have been shown to thin on the flanks of anticlines and thicken in synclines (Cater, 1970). The basal Triassic unit in the basin is the Moenkopi Formation (Fig. 3) of Early and Middle (?) Triassic age (Nuccio and Condon, 2000). Lower beds of the Moenkopi are of fluvial origin that shed eastward from a highland west of the Paradox Basin (Hunton, 1992; Huntoon et al., 1994). Younger members are a combination of sabkha, mudflat, fluvial, and one marine limestone unit (Nuccio and Condon, 2000). Erosion and possible non-deposition in the Middle Triassic led to the absence of the Moenkopi in parts of southeastern Utah and most of southwestern Colorado (Stewart et al., 1972, Nuccio and Condon, 2000). Uplift south of the study area in Late Triassic time led to the development of the northward-flowing fluvial system in the Upper Triassic Chinle Formation (Fig. 3) (Dubiel, 1989).

The contact at the top of the Chinle Formation and the base of the Glen Canyon Group was interpreted to be an unconformity as well as the Triassic-Jurassic boundary by Pipiringos and O’Sullivan (1978). The Glen Canyon Group (Fig. 3) is composed of the Wingate Formation, Kayenta Formation, and Navajo Formation (Nuccio and Condon, 2000). The Wingate and Navajo are massive eolian units, and the Kayenta is of fluvial origin (Nuccio and Condon, 2000). Unconformably overlying the Glen Canyon Group is the Middle Jurassic San Rafael Group (Fig. 3), which consists of the Carmel Formation, Entrada Formation, and the Summerville Formation (Nuccio and Condon, 2000). The Carmel Formation is a marine limestone, sandstone, and shale
(Nuccio and Condon, 2000). The Carmel is overlain by the Entrada Formation, a silty sandstone, likely deposited in or near shallow water in the San Rafael Swell that transitions eastward to an eolian deposit (Nuccio and Condon, 2000). The Summerville Formation (including the Wanakah Formation) conformably overlies the Entrada Formation and was deposited in a sabkha depositional environment (Nuccio and Condon, 2000). The Upper Jurassic Morrison Formation (Fig. 3) is the youngest Jurassic unit in the basin, deposited in a variety of depositional environments, ranging from eolian to fluvial and lacustrine (Nuccio and Condon, 2000).

Salt-withdrawal mini-basins developed adjacent to salt walls and subsequently influenced variation in sediment thickness and facies, especially adjacent to the flanks of the steep-sided salt walls (Doelling, 1988; Trudgill et al., 2004). Passive diapirism is interpreted to be the major mechanism for the salt structure growth throughout this time followed by later shortening during the Paleogene Laramide orogenic event during development of Sevier Foreland Basins (Grout and Verbeek, 1997; Trudgill et al., 2004). These northwest-trending salt structures are confined to the Paradox fold and fault belt and range in structural style from buried salt pillows to exposed salt walls (Doelling et al., 2002; Trudgill et al., 2004). The salt walls are interpreted to have been strongly controlled in orientation and thickness by deep-seated northwestward tending basement faults (Trudgill et al., 2004).
Figure 2. Location map of the salt anticline region within the Paradox Basin. Salt walls: Castle Valley; Fisher Valley; Gypsum Valley; Lisbon Valley-Dolores; Moab-Spanish Valley; Paradox Valley; Schafer Dome; Salt Valley; Sinbad Valley. Extent of Paradox evaporite is from Condon (1997); distribution of salt walls is after Shoemaker et al., 1958. Study area is outlined in red, located on the northeast margin of Gypsum Valley.
Figure 3. Stratigraphic column of units in the northern Paradox Basin in the Gypsum Valley area. Colors used match those on geologic maps and cross sections of the Gypsum Valley salt wall. Depositional environment, depositional events and controls are after Stokes and Phoenix (1948); Doelling and Ross, (1998); and Trudgill (2011). Key unconformities annotated are the Tr-3 and J-0 unconformities.
Starting in the Late Jurassic or Early Cretaceous (Morrison Formation), all of the diapirs and other salt structures in the Paradox Basin were buried with the overburden thickness reaching several thousands of feet (1.6 km or more) by the end of the Cretaceous (Rasmussen, 2015). Burial continued through the Paleogene (until c. 35 Ma), about 5 million years after the Laramide Orogeny (80-40 Ma) had ended (Rasmussen, 2015). During the Laramide Orogeny, regional west-east shortening folded all of the diapirs and anticlines (Rasmussen, 2015). Normal faults that commonly strike parallel to the salt wall structures are thought to be associated with salt dissolution (Trudgill, 2011). Widespread incision seen in the Rocky Mountains and Colorado Plateau during the Neogene exhumed diapir evaporites, and the relatively wetter climate of the Pleistocene led to increased erosion, dissolution, and subsequent collapse of the salt anticline roof rocks (Rasmussen, 2015), Nuccio and Condon (1996) estimate that more than 3 km of overburden were eroded during this time.

2.2 Gypsum Valley Salt Wall

Gypsum Valley (48 kilometer x 3.2 km) is a breached and collapsed, NW-SE trending salt wall in southwest Colorado (Cater, 1970). Paradox Formation evaporite is exposed on the valley floor (Fig. 4), but no halite is present. Only a residuum of gypsum caprock remains, thus giving the valley its name. Permian, Triassic, and Jurassic strata are exposed along the flanks of Gypsum Valley and dip away from the core of salt (Cater, 1970). Both ends of the anticline transition into flat-lying strata (Fig. 4) (Cater, 1970). Generally, older formations dip more steeply, and successively younger formations are deposited across the upturned and truncated edges of older formations (Cater, 1970). Gypsum Valley is flanked to the northeast by Dry Creek minibasin and to the southwest by Disappointment minibasin. Dry Creek minibasin plunges to the southeast and flattens to the northwest (Cater, 1970). Dissolution of the diapir
resulted in the collapse of its roof, where modern caprock forms the floor of a northwest-trending valley (Cater, 1970). Beginning at about 25 Ma and continuing until the present, uplift and erosion have stripped approximately 13,350 feet of strata from the area, exposing Upper Pennsylvanian rocks at the surface (Nuccio and Condon, 2000). Collapse of the axial part of the anticline likely occurred in multiple stages: starting with the crests of the anticlines were down dropped as grabens followed by Miocene uplift of the Colorado Plateau, initiating major stream incision and concomitant breaching of salt cores (Cater, 1970).
Figure 4. Geologic map of Gypsum Valley, Colorado based on Stokes and Phoenix (1948). Units and colors match those on the stratigraphic column (Figure 3). The study area (Figure 5) is outlined in red in both map and cross section view.
Gypsum Valley (Fig. 4) is geographically separated into three recognized components marked by where the Dolores River cuts across the salt wall. Little Gypsum Valley occupies the area to the northwest of the Dolores River and Big Gypsum Valley occupies the area to the southwest of the Dolores River, and Silveys Pocket occupies the area furthest northwest, draining a local portion of the valley (Stokes and Phoenix, 1948). In Big Gypsum Valley, vertical to overturned strata as old as Pennsylvanian are exposed, largely in positions where erosion has penetrated the overlying Morison Formation (Stokes and Phoenix, 1948). In little Gypsum Valley, extensive outcrops of the Morrison Formation are present along the valley floor (Stokes and Phoenix, 1948). Northwest of the Dolores River, Gypsum Valley curves northward where collapsed blocks of the Salt Wash Member of the Morrison Formation are found in contact with the Paradox Formation (Cater, 1970). This relationship indicates that due to collapse of the salt wall, strata that were once overlying the anticline have been displaced up to 3,000 feet below their pre-collapse or regional positions (Cater, 1970). Along the northeast side of Little Gypsum Valley, the Chinle Formation unconformably overlies the Paradox Formation at Bridge Canyon (Fig. 4).

2.3 Salt Shoulder at Bridge Canyon, Little Gypsum Valley

The Upper Triassic Chinle Formation is exposed on each side of the Dolores River on the northeast margin of Gypsum Valley (Fig. 4) in a narrow canyon informally named “Bridge Canyon” (Fig. 5). Within this area, outcrop of the Chinle extends northeast for 1.5 km until it is no longer exposed at the surface. This area was selected for this study because it provides a unique 3-dimensional exposure of strata of the Chinle Formation onlapping and overlapping the exposed shoulder of the diapir (Fig. 5). The lower strata overlying the salt shoulder dip gently (4-6º) to the east, and transition to dips (23-27º) to the northeast on the margin of the shoulder,
and pass into subcrop to the northeast. The Chinle is unconformably overlain by the Latest Triassic to Jurassic Wingate Formation (Fig. 6), which also thins toward the diapir. The Chinle, Wingate, Kayenta, and Navajo formations that overly the Chinle are all offset by high angle (85-88º) normal faults that plunge to the southwest (Fig. 5).
Figure 5. Geologic map of Bridge Canyon in Little Gypsum Valley and panoramic view of Bridge Canyon area. (A) Base-layer map is a high-resolution photograph from ESRI World Imagery. (B) Photopanorama of the north canyon wall of exposed Chinle overlying caprock with measured sections for reference.
Figure 6. Three-dimensional reconstruction of Bridge Canyon. Individual point cloud images of Bridge Canyon from various vantage points were taken using the software Vulcan Maptek. Formations shown in reconstruction include in ascending order: GV Salt Wall gypsum and dolomite caprock in grey/tan, Chinle Formation in red, brick red that forms steep slopes with intermittent ledges, Wingate Formation in tan/red that forms vertical cliffs, and the Kayenta Formation in dark red that forms steep ledges. (A) Birds eye view facing the northern wall of Bridge Canyon showing the slope forming Chinle Formation onlapping and overlapping caprock. (B) Ground view looking down Bridge Canyon from outcropping gypsum caprock. (C) View looking east of the Chinle Formation onlapping caprock and the overlying cliff-forming Wingate & Kayenta Formations.
2.4 Chinle Formation Depositional History and Regional Stratigraphy

The Late Triassic Chinle Formation extends over most of the Colorado Plateau and is composed of a variety of continental sedimentary rocks including claystone, siltstone, sandstone, limestone, and conglomerate (Stewart et al., 1972; Blakey & Gubitosa, 1984; Dubiel, 1987; Lucas et al., 1997). It is defined by six formal lithostratigraphic members, which in ascending order are the: Shinarump, Monitor Butte, Moss Back, Petrified Forest, Owl Rock, and Church Rock as described by Stewart et al. (1972). However, the standard lithostratigraphic members are not evident and are difficult to correlate adjacent to salt structures (Blakey and Gubitosa, 1984; Hazel, 1991, 1994). The Chinle pinches out on the flanks of the Uncompahgre and Front Range highlands to the east (Stewart et al., 1972). The name Chinle Formation is generally used in Colorado and Utah, however, rocks of similar age and lithologic type are considered to be a part of the Popo Agie Formation to the north in Wyoming (Stewart et al., 1972). To the southeast, age equivalent rocks in southeastern Colorado, eastern New Mexico, Oklahoma, southwestern Kansas and Western Texas, are referred to as the Dockum Group (McKee et al., 1959). To the south, the Chinle Formation likely did not extend beyond the ancient Mogollan highland (Fig. 7) in modern day southern Arizona, New Mexico, and California (Stewart et al., 1972). The western terrane likely consisted of low-lying land, volcanic arcs and small marine basins, however, is less well understood (Blakey and Gubitosa, 1984).

Throughout most of the Colorado Plateau, the Chinle Formation unconformably overlies the early to mid-Triassic Moenkopi Formation and the contact represents a ~ 17 m.y. regional unconformity (Tr-3) where paleo-valleys of the Moenkopi are infilled by the Shinarump Member of the Chinle (Stewart et al., 1972; Doelling, 2000). The Chinle is unconformably (J-0) overlain by the Upper Triassic Wingate Formation (Pipiringos and O’Sullivan, 1978) of the Glen Canyon
Group (Stewart et al., 1972). Regionally, the Chinle is informally subdivided into two major parts based on lithologic characteristics and the source of material composing the rocks: (1) the lower bentonitic part consists of the Monitor Butte, Petrified Forest, and minor, ledge forming members of the Shinarump and Moss Back, and (2) the upper, red-bedded part that consists of the Owl Rock and Church Rock members (Fig. 8) (Stewart et al., 1972). The lower part is typically composed of abundant volcanic debris derived from the southern-sourced Mogollan highlands (Stewart et al., 1972). The upper part is commonly arkosic and largely derived from the igneous and metamorphic terrane of the Uncompahgre and Front Range highlands of Colorado and Northern New Mexico (Stewart et al., 1972). Chinle rocks are coarser grained and volcanic-rich to the south and southeast, whereas, they are finer grained and less volcanic to the north (Blakey and Gubitosa, 1984). Carbonate-rich sandstone and limestone are common in the central part of the Paradox Basin and on the western margin of the salt anticline region (Blakey and Gubitosa, 1984).

Chinle sandstone bodies and associated mudstones record the changing patterns in Late Triassic tectonism and sedimentation, where variations in facies architecture are mainly driven by fluctuations in regional accommodation space (Blakey and Gubitosa, 1984; Lucas, 1993; Lucas et al., 1997; Matthews, 2007). Changes in vertical and lateral sediment accumulation mainly reflect changes in sediment supply, climate, volcanic input into the basin, rate of subsidence, frequency of avulsion and other intra- and extra-basinal parameters (Blakey and Gubitosa, 1984). The extensive sheets representing braided fluvial sandstones and conglomerates of the Moss Back and Shinarump members correspond to a time of low regional accommodation, whereas, isolated sandstones and lacustrine mudstones of the Monitor Butte, Petrified Forest, Owl Rock, and Church Rock represent a period of high regional accommodation.
(Lucas, 1993; Lucas et al., 1997; Matthews, 2007). The differences in available accommodation space during Chinle deposition likely reflect a change in the rate of regional tectonic subsidence (Blakey and Gubitosa, 1984).

The Chinle was deposited approximately 5-15 degrees north of the paleoequator (Van der Voo et al., 1976; Habicht, 1979; Ziegler et al., 1983; Parrish and Peterson, 1988; Bazard and Butler, 1991) during a major global reorganization of plate regimes that included the initial breakup of Pangea and consequent westward motion of the North American plate (Coney, 1978; Hazel, 1994). Late Triassic climate was dominated by monsoonal circulation and alternating wet and dry seasons (Dubiel, 1991) with increasing aridity toward the end of the deposition of the Church Rock member (Stewart et al., 1972; Blakey and Gubitosa 1983, 1984; Dubiel, 1987; Blodgett, 1988; Blakey et al., 1988). Evidence documented by fluvial channels, crevasse splays, lakes, bogs, marshes and lacustrine deltas reflect abundant precipitation and shallow water tables (Dubiel, 1991), although paleosols and ichnofossils indicate water tables and lake levels fluctuated episodically (Dubiel, 1991; Prochnow et al., 2006a). Strata in the uppermost Chinle consists of lacustrine and marginal-lacustrine mudstones interbedded with minor eolian sand sheets and eolian dunes, indicating continued precipitation marked by pronounced and extended dry seasons (Dubiel, 1991). The dominant paleocurrent direction of streamflow in Western Colorado and southeastern Utah trended to the west and southwest (Figure 7) (Schultz, 1963; Stewart et al., 1972). The Chinle alluvial plain extended far (>100 km) from source areas and was dominated by seasonal precipitation and low-gradient deposition (Hazel, 1994).
Figure 7. Fluvial-lacustrine deposition of the Upper Triassic Chinle Formation (after Dubiel 1991). Location of Gypsum Valley is outlined red rectangle. Paleogeographic map of the southwestern United States during the Late Triassic. This interpretation shows the position of Gypsum Valley, the Ancestral Rocky Mountain highlands, and northwestward fluvial transport of the Chinle Formation and the regressing coastline moving west/northwest after Blakey (2008).
2.5 Chinle Formation Stratigraphy Within the Study Area

In southwest Colorado, Chinle-age strata are collectively termed the Dolores Formation (Lucas et al., 1997). It has long been recognized that Chinle and Dolores strata are equivalent and can be closely related (Stewart et al., 1972), however, the Dolores Formation is only applied locally in southwestern Colorado (Lucas et al., 1997). Stewart et al., 1972 summarized the Dolores Formation and correlated a lower, middle and upper member to the formal lithostratigraphic members within the Chinle Formation. The “lower member” was correlated by Shawe et al., 1968 to the Moss Back Member (Fig. 8), which contains as much as 27 m of greenish gray to tan, fine-grained quartzose sandstone and limestone-pebble clast conglomerate. The middle member was correlated by Shawe et al., 1968 to the Petrified Forest Member (Fig. 8), which contains as much as 83 m of grayish red mudstone, siltstone and minor beds of trough-crossbedded, fine-grained sandstone and limestone-pebble conglomerate. The “upper member” of the Dolores Formation is referred to as the Church Rock Member (Fig. 8) by Shawe et al., 1968, which contains as much as 220 meters of horizontally bedded, light brown and reddish brown fine-grained sandstone and non-bentonitic siltstone (Lucas et al., 1997).

The Chinle Formation in the area of Gypsum Valley outcrops as steep slopes of shale and siltstone interrupted by resistant ledges and cliffs of siltstone, sandstone and conglomerate (Shawe et al., 1968). The Chinle is exposed in it’s entirety in the Dolores River Canyon, where the stratigraphy was documented by Shawe et al., (1968), and reaches a maximum thickness of about 400 feet or approximately 121 meters (Fig. 8). Well logs adjacent to the study area in Andy’s Mesa field have logged up to 165.5 meters TVD in the Double Eagle Unit of Chinle Formation strata (Cole III et al., 2009). The Chinle locally thins abruptly or pinches out onto the
Figure 8. Lithostratigraphic units of the Chinle Formation in the Dolores River Canyon. Modified from Shawe et al., 1968. The formal units of the Shinarump and Monitor Butte are not correlatable in the basal sections of the Chinle Formation in the Gypsum Valley area. The formal naming of the Chinle Formation in the field area is adopted from Stokes and Phoenix (1948) mapping of Gypsum Valley.
Gypsum Valley anticline, where members become lithostratigraphically indistinguishable (Shinarump and Monitor Butte are not correlatable).

The siltstone is generally orangish-brown to reddish-brown, has a high clay content, and is indistinctly bedded, however, where clay content is low, the siltstone forms resistant ledges (Shawe et al., 1968). Mudstone is common as a greenish gray to reddish brown with a variable content of sand and silt, and is likely bentonitic if greenish gray in color (Shawe et al., 1968). Conglomerate is dark gray to greenish gray and is typically composed of limestone clasts and a small number of shale, quartzite, and chert clasts that range up to pebble in size in a calcareous-clay matrix (Shawe et al., 1968). The sandstone is generally light reddish-brown and light gray to greenish gray. Reddish-brown units are colored by abundant hematite films on detrital grains and light greenish gray sandstone is very fine to coarse grained with clasts composed of quartz, feldspar and minor amount of mica (Shawe et al., 1968). Bedding and sorting in the sandstone is variable, ranging from massive to thin bedded and crossbedded to horizontally bedded (Shawe et al., 1968). Carbonitized plant fragments and other carbonaceous material are common in light-gray and greenish-gray sandstones and conglomerates (Shawe et al., 1968).
Chapter 3: Methods

Stratigraphic analysis of the Chinle Formation Gypsum Valley salt shoulder involved a total of 20 measured sections extending from the base of the Chinle Formation to the base of the Wingate Formation on the northeast margin of Little Gypsum Valley (Figure 5). The sections range in thickness from 54.25 meters (178 feet) to 160 meters (525 feet). Sections were measured using an ASC precision Jacob’s staff with an Abney level for bed thickness and a Brunton Compass for bed orientation. Within the measured sections, units were described and photographed. Lithofacies, paleocurrent, strike/dip, sedimentary structures, grain size, sorting, angularity, color, and thickness were documented where units changed. The measured sections were georeferenced in cm-scale resolution using a Trimble GeoXH GPS device. Beds were walked laterally and documented to correlate each measured section. Faults were located and apparent offsets were measured and given a GPS location. This process was aided by aerial imagery in the software ArcGIS for areal continuity of faults along strike of the salt wall. Within each section, units were documented and marked by a GPS position. Paleocurrent measurements were collected from trough cross-stratified beds to determine current flow in 15 different locations with a Brunton Compass. Strike and Dip measurements were documented using a GPS-enabled digital clinometer (GeoClino-G) to generate a position with an x (longitude), y (latitude), and z (elevation) position in order to provide a 3-dimensional location.

Field data were combined and organized using ArcGIS. ESRI world imagery was downloaded for basemaps and GPS points from measured sections, lines from faults, and lines collected from laterally walked units were input onto the basemap. The points collected are associated with unit numbers and section references to ensure correct placement on the map.
ArcGIS was used as the primary source for maps, shapefile collection, and platform for organization from all GPS related field mapping.

In order to build a three dimensional model of the salt shoulder in Bridge Canyon, a total of 516 photographs were taken of the canyon walls, the salt-sediment contact, and the Dolores Canyon river valley in a different GPS location along strike using a Sony Cyber-shot DSC HX90V digital camera with a built in GPS sensor to tag photos with a three dimensional GPS location of where the photo was taken. These photographs were placed in a series of groups based on where the photograph was taken, the part of the field area being photographed, and connectivity of the photographs in the series. Once the series of photographs were arranged, utilization of the 3D scanning software AGISoft Photoscan processed and developed detailed point clouds of each photo grouping, known to the software as ‘data chunks’. The various ‘data chunks’ are processed by the software resulting in point clouds, or an array of georeferenced pixels in 3D space. Fifteen total point clouds generated in AGIsoft Photoscan were subsequently combined into a master, to-scale, 3D representation of Bridge Canyon with a total of 20 million georeferenced data pixels.

The full reconstruction of Bridge Canyon was viewed and geology was added using the software Vulcan Maptek. Within this software program, GPS data from measured sections (points and lines) representing bedding contacts, lateral changes of units, and faults were imported and placed onto the master point cloud. Strata were correlated across the canyon using three dimensional polygons, faults were correlated and offset within the polygons was inferred. The resulting polygons honor structural and stratigraphic relationships. Eight total polygons were generated in Maptek and imported into Midland Valley Move. These polygons were then combined with the georeferenced strike and dip orientation points that had 3D data associated
with them along with their strike and dip values and orientations collected with the GeoClino-G. These are referred to as ‘structure disks’ in the software Move and were combined with the associated unit polygon that the strike and dip data were collected on. Combining polygons with orientation data allowed projection of surfaces with meter scale field-accuracy. Eight total surfaces were created, seven within the Chinle Formation and one of the top of the caprock surface representing the salt-sediment interface of the shoulder.

Photographs were taken with a Sony Rebel T3i DSLR camera and merged in Adobe Photoshop in order be finalized as high resolution photo-panoramas. These panoramas aided in the documentation of lateral changes in facies, sand-body geometries, and channel dimensions. It also aided in arranging measured sections in order to overlay facies distribution, halokinetic sequences, and halokinetic sequence boundaries. Photo-panels were interpreted and figures were completed using Adobe Illustrator.
Chapter 4: Lithofacies and Facies Associations of the Chinle Formation at the Gypsum Valley Salt Wall Shoulder

Seven facies associations were determined within the Chinle Formation designated from groups of lithofacies classified using Miall (1978) facies codes. The facies associations are based on lithology, sedimentary structures, geometry, nature of bedding contacts, and presence of fossils from measured sections starting at the top of the caprock (salt-sediment interface) to the base of the Wingate Formation. The following facies associations were identified: 1) non-caprock bearing channel-fill sandstone and stratified conglomerate (FA1), 2) caprock-bearing channel-fill sandstone and stratified conglomerate (FA2), 3) tabular shale, siltstone (FA3), 4) unsorted conglomerate lense (FA4), 5) caprock-bearing heterolithic channel sandstone and conglomerate (FA5), 6) fossiliferous mudstones & sandstones (FA6), 7) paleosols (FA7).

In addition, laminated & brecciated caprock is mapped where it is in contact with the Chinle Formation in the study area. The Gypsum Valley salt wall unconformably underlies the Chinle Formation on the northeast margin of GV salt wall in Little Gypsum Valley. It is represented by caprock, a mixed lithology, ledge-forming facies that is massive to crudely bedded. It most often represents an erosional contact with the overlying Chinle Formation. This facies commonly has a fetted odor when broken apart. Units range in thickness from .5 to 3.5 meters folded (Fig. 9a), laterally continuous sequences of dolomite (Fig. 9b), gypsum (Fig. 9c), anhydrite, and shale. Carbonate sequences typically become increasingly brecciated toward the top of a sequence (Fig. 9d). This facies contains the following colors: tan, yellow, gray, and white. The occurrence of a scoured contact between the FA2 of the Chinle Formation and suggests that the caprock (Fig. 9d) was present prior to deposition of the Chinle (Fig. 9e). Within caprock bearing facies, evidence is drawn from Lawton and Buck (2006) at Castle
Valley, a direct comparison to lithologies seen at Gypsum Valley where there is no regional source for this type of lithology.
Table 1. Lithofacies codes, descriptions, sedimentary structures, and interpretation. Modified from Miall, 1978, 1996.

<table>
<thead>
<tr>
<th>Facies code</th>
<th>Facies</th>
<th>Sedimentary structures</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gmm</td>
<td>Matrix-supported, massive gravel</td>
<td>Weak grading</td>
<td>Plastic debris flow (high strength, viscous)</td>
</tr>
<tr>
<td>Gcm</td>
<td>Clast-supported, massive gravel</td>
<td>Weak grading</td>
<td>Low-strength, psuedo-plastic debris flow</td>
</tr>
<tr>
<td>Gh</td>
<td>Clast-supported, crudely bedded gravel</td>
<td>Horizontal bedding, imbrication</td>
<td>Longitudinal bedforms, lag deposits, sieve deposits</td>
</tr>
<tr>
<td>Gt</td>
<td>Gravel, stratified</td>
<td>Trough cross-beds</td>
<td>Minor channel fills</td>
</tr>
<tr>
<td>Gp</td>
<td>Gravel, stratified</td>
<td>Planar cross-beds</td>
<td>Transverse bedforms, deltaic growths from older bar remnants</td>
</tr>
<tr>
<td>St</td>
<td>Sand, fine to very coarse, may be pebbly</td>
<td>Solitary or grouped planar cross-beds</td>
<td>Transverse and linguoid (3-D) dunes</td>
</tr>
<tr>
<td>Sp</td>
<td>Sand, fine to very coarse, may be pebbly</td>
<td>Solitary or grouped planar cross-beds</td>
<td>Transverse and linguoid bedforms (2D dunes)</td>
</tr>
<tr>
<td>Sr</td>
<td>Sand, very fine to coarse</td>
<td>Ripple cross-lamination</td>
<td>Ripples (lower flow regime)</td>
</tr>
<tr>
<td>Sh</td>
<td>Sand, very fine to coarse, may be pebbly</td>
<td>Horizontal lamination parting or streaming lineation</td>
<td>Plane-bed flow (critical flow)</td>
</tr>
<tr>
<td>Sl</td>
<td>Sand, very fine to coarse, may be pebbly</td>
<td>Low-angle (&lt;15°) cross-beds</td>
<td>Scour fills, hummock or washed-out dunes, antidunes</td>
</tr>
<tr>
<td>Fl</td>
<td>Sand, silt, mud</td>
<td>Fine lamination, very small ripples</td>
<td>Overbank, abandoned channel, or waning flood deposits</td>
</tr>
<tr>
<td>Fh</td>
<td>Sand, silt, mud</td>
<td>Horizontal lamination to massive</td>
<td>Deposition from suspension, lower flow regime</td>
</tr>
<tr>
<td>Fm</td>
<td>Mud, silt</td>
<td>Massive, dessication cracks</td>
<td>Overbank, abandoned channel, or drape deposits</td>
</tr>
<tr>
<td>P</td>
<td>Paleosol</td>
<td>Pedogenic features: nodules, filaments</td>
<td>Soil with chemical precipitation</td>
</tr>
</tbody>
</table>
Table 2. Facies association chart displaying photo description, symbol, (Miall, 1996) lithofacies present, odor (‘scratch and sniff’), grain size, sedimentary features, and geometry of strata for the Chinle Formation and the adjacent salt wall gypsum & dolomite caprock.

<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Photo</th>
<th>Description</th>
<th>Symbols((\text{Miall}, 1978, 1996)) Lithofacies Present</th>
<th>Color</th>
<th>Grain/clast size</th>
<th>Sedimentary Features</th>
<th>Geometry</th>
</tr>
</thead>
<tbody>
<tr>
<td>FA1</td>
<td></td>
<td>Non-caprock bearing channel fill sandstone and stratified conglomerate</td>
<td>St, Sp, Sh, Sr, Gr, Gp, Gh, Fh, Fl</td>
<td>Tan, red, light red, brick red, dark gray</td>
<td>Well sorted, upper very fine to upper medium s, Clasts: granule-pebble</td>
<td>Flame structures, silicified wood</td>
<td>(1) Amalgamated, multi-story sheets (25-50 m wide) and (7-20 m thick) (2) Isolated, multi-story ribbons (75-150 m wide) and (3-5 m thick)</td>
</tr>
<tr>
<td>FA2</td>
<td></td>
<td>Caprock-bearing channel fill sandstone and stratified conglomerate</td>
<td>St, Sp, Sh, Gr, Gp, Gh, Fh, Fl</td>
<td>Tan, tan-yellow, red, light red, gray, light gray</td>
<td>Well sorted, upper very fine to lower medium s matrix, Clasts: granule-cobble</td>
<td>Flame structures, silicified wood</td>
<td>(1) Narrow, single-story sheets (150-250 m wide) and (2.5 m thick) (2) Narrow, multi-story sheets (250-400 m wide) and (10-25 m thick) (3) Simple, single-story isolated ribbons (60-75 m wide) and (3-4 m thick)</td>
</tr>
<tr>
<td>FA3</td>
<td></td>
<td>Tabular shale, siltstone</td>
<td>Sr, Fh, Fl, Fm</td>
<td>Dark red, brown, light pink, gray, green</td>
<td>Siltstone, claystone, sandy siltstone, very fine grained s</td>
<td>Flame structures, siltstone nodules, desiccation cracks</td>
<td>Tabular and commonly laterally extensive; up to 60 meters thick and 10-100's of meters wide</td>
</tr>
<tr>
<td>FA4</td>
<td></td>
<td>Unsorted conglomerate</td>
<td>Gmm, Gcm</td>
<td>Tan, gray, purple, green, dark red</td>
<td>Very poorly to poorly sorted angular to sub-rounded granule-to-boulder clasts; upper medium to coarse grained matrix</td>
<td>Calcite, siltstone nodules, silicified wood</td>
<td>Lenticular and predominantly isolated: (3-15 meters thick and up to 125 meters wide)</td>
</tr>
<tr>
<td>FA5</td>
<td></td>
<td>Caprock-bearing heterolithic channel fill</td>
<td>St, Sp, Sh, Gr, Gp, Fl, Fm</td>
<td>Brown, light red</td>
<td>Well sorted, upper very fine to lower medium s matrix, Clasts: granule-pebble, siltstone, claystone</td>
<td>Mud drapes</td>
<td>Lenticular and isolated: (75-100 meters wide and 3.5 to 7.5 m thick)</td>
</tr>
<tr>
<td>FA6</td>
<td></td>
<td>Fossiliferous mudstones and sandstones</td>
<td></td>
<td>Light red, brown, gray</td>
<td>Modernly sorted fine-grained sandstone and siltstone matrix</td>
<td><em>Triassic</em>uroidea (unianulated bivalves), concretions, carbon films, silicified wood, <em>Cambrygyna</em> burrows</td>
<td>Tubular and isolated: (25-75 meters wide and 3.2-3.5 meters thick)</td>
</tr>
<tr>
<td>FA7</td>
<td></td>
<td>Paleosols</td>
<td>P</td>
<td>Red, dark red, gray, white</td>
<td>Siltstone, Charyses and very well sorted very fine to fine grained s</td>
<td><em>Seponia</em> burrows, rhizocretions, blocky peds, reduction zones, concretions</td>
<td>Tubular and laterally extensive; (75-125 meters wide and 3.5 to 5 meters thick)</td>
</tr>
<tr>
<td>Gypsum Valley Salt Wall</td>
<td></td>
<td>Coherent and brecciated caprock</td>
<td></td>
<td>Tan, yellow, gray, white</td>
<td></td>
<td>Folded beds, breccia</td>
<td>Lenticularly extensive</td>
</tr>
</tbody>
</table>
Figure 9. Select outcrop photos of GV salt wall caprock from various measured sections (Fig. 5). (A) Folds in carbonate-caprock along the salt-sediment interface at the base of section 7, rock hammer for scale in center of photograph is 12.5” in length. (B) Coherent carbonate caprock at the base of section 5, scale ruler is 15 cm wide. (C) Isoclinal fold of gypsum along the margin of the Gypsum Valley Salt Wall with overlying Chinle Formation at the base of section 14. (D) FA2 incising into caprock, the dark-red sandstone is the Chinle Formation lithified in a carbonate caprock breccia located at the base of section 8, lense cap for scale is 3” in diameter. (E) Elongate bed of caprock along the salt-sediment interface with overlying Chinle beds located at the base of section 15, human for scale is approximately 1.8 meters tall.
4.1 Facies Association 1. Non-Caprock Bearing Channel-Fill Sandstone and Stratified Conglomerate (FA1)

**Description:**

FA1 is less common in positions near the Gypsum Valley salt wall (< .5 km, measured perpendicular to strike to the salt-sediment interface). However, it can be found in isolated positions in normally graded channel successions from conglomerate to well-sorted, subarkose to lithic arkose sandstones. FA1 contains trough (St) (Fig. 10a), planar/tabular (Sp) (Fig. 10b), horizontal (Sh) (Fig.10c), and ripple/climbing ripple (Sr) stratified sandstone (Fig. 10d) lithofacies. Trough cross stratified conglomerate (Gt) (Fig. 10e) is most common with rare planar/tabular (Gp) (Fig.10f), and horizontal (Gh) stratified conglomerate lithofacies, and horizontally laminated (Fh) (Fig.10g) and finely laminated (Fl) (Fig. 10g) fine-grained sandstone and mudstone are common. Matrix grain size variation does not change considerably except for transitions from conglomerate lithofacies (Gt,Gp, Gh) into a well-sorted sandstone (St,Sp,Sh,Sr) within a channel fill sequence (Fig.10h). Flame structures are most commonly present in the Sh and Sp (Fig. 10i) sandstone lithofacies (Fig. 10j).

FA1 is characterized by 0.1 to 3.5 meter thick, and 7-10 meter wide, channel-form beds of well sorted, upper very fine- to upper medium-grained sandstones that most commonly overly granule- to cobble-sized clast conglomerates that grade upward from an erosional, scoured basal surface. The most common stacking pattern from bottom to top for this facies is Gt, Sr, Fl (with nodule-development and color mottling). Trough crossbed sets are commonly 1-3 meters wide, and .4 to .8 meters thick. Conglomerates at the bases of channel sequences contain sub-rounded to rounded mud rip-up clasts, silicified wood, and chert clasts that commonly overly an erosional surface. These beds are contained within individual channels with dimensions ranging from 75-
250 meters across and 2-5 meters thick. This facies has the following colors: tan, red, light red, and brick red and dark gray.

Fining upward channel successions are often stacked vertically and laterally into amalgamated, multi-story sheet sand bodies. Amalgamated, multi-story channels are grouped into sand bodies ranging in thickness from 7 to 20 meters and 250 to (>500) meters in width. Single-story, isolated channels, though less common, also exist, and are more typical in positions near the diapir. Channels range in thickness from 3-5 meters thick and 75-150 meters wide. Paleocurrent measurements from beds in this facies dominantly trend south-southeast.

**Interpretation:**

Lithofacies Gt, Gp, and Gh represent gravel bars and bedforms deposited by traction currents (Miall, 1985). Channel floor lags are interpreted where pebble conglomerates occur at the base of a channel sequence (overlying a scoured surface) and bar platforms are interpreted where conglomerates are documented at the middle or toward the top of the channel sequence (Miall, 1985). Lithofacies St, Sp, Sh, and Sr represent minor bar forms and channel fills deposited by traction currents as bed load or by intermittent suspension (saltation) (Miall, 1996). The presence of flame structures is interpreted to represent water-escape due to rapid aggradation (Boggs, 1987).

FA1 represents various fluvial channel depositional systems based on the diversity of sand body geometries, sedimentary structures, and composition of clasts. The most common succession of Gt, Sr, to Fl with nodule development and mottling is indicative of deposition in rivers with traction flow scouring and incorporating mud chips from the floodplain, at low flow speeds (<1 m/s) for suspended load sands due to the presence of ripples (Miall, 1996), and on a
floodplain environment where isolated, periodic wetting and drying exists (Brown and Kraus, 1981) due to nodule development in fine-grained sandstones and mudstones.

Isolated channels form narrow ribbons (Fig. 11a) that are defined by low width/depth ratios (5:1-20:1; Hazel, 1994) and can be both single or multi-story (Blakey & Gubitosa, 1984). Isolated ribbons are interpreted to represent meandering river deposits based on their association with (1) lateral accretion deposits showing point bar migration, (2) their isolated character within overbank mudstones, and (3) low width/depth ratios (5:1-20:1; Dubiel, 1987; Hazel, 1991, 1994; Matthews, 2007). The presence of amalgamated, multi-story channels that form both narrow and broad sheets (Fig. 11b) and have high width/depth ratios (>100:1; Hazel, 1994) are interpreted to represent a relatively unrestricted braided fluvial system. Amalgamated, multi-story sheets of FA1 represent braided river deposits based on the following components: (1) multiple, stacked channels, (2) association with downstream accretion deposits representing mid-channel bar migration, and (3) high width/depth ratios (>100:1) (Hazel 1991, 1994; Matthews 2007).
Figure 10. Outcrop photographs of FA1 in various measured sections (A) St, trough cross-stratified sandstone located within section 5, scale ruler is 15 cm wide. (B) Sp, planar cross-stratified sandstone located within section 6. (C) Sh, horizontally laminated sandstone located within section 6. (D) Sr, ripple cross-laminated sandstone located within section 5, scale ruler is 14 cm wide. (E) St overlying Gt located within section 15. (F) Gh, horizontally stratified conglomerate overlying Gp, planar/tabular stratified conglomerate located within section 15. (G) Fh, horizontally laminated fine grained sandstone and mudstone overlying Fl, finely laminated sandstone and mudstone located within section 13. (H) Channel-fill sequence located within section 8. (I) Sh, horizontally laminated sandstone overlying Sp, planar tabular laminated sandstone located within section 12. (J) Soft-sediment deformation compaction flame structures located within section 12.
**Figure 11.** Classification of fluvial sand body geometries modified from Blakey and Gubitosa (1984). (A) Ribbon geometry, (B) Sheet geometry.
4.2 Facies Association 2. Caprock-Bearing Channel-Fill Sandstone and Stratified Conglomerate (FA2)

Description:

FA2 is confined to positions near the GV salt wall (< 1 km perpendicular to strike of the margin of the salt-sediment interface). Caprock-bearing units have been uniquely identified in this study in order to document episodes of fluvial interaction with the GV salt wall. FA2 is characterized by gypsum, limestone, or dolostone clasts (Heness, 2016) in gravel conglomerates commonly found at the bases of channel successions and subordinately in stratified conglomerates in the middle of the channel sequence. The conglomerate facies most commonly fines upward into a well-sorted, subarkose to lithic arkose sandstone. This facies contains trough (Gt) (Fig. 12a), planar/tabular (Gp) (Fig. 12b), and horizontal (Gh) (Fig. 12c) stratification, planar/tabular (Sp), horizontal (Sh) stratified (Fig. 12d), trough (St) (Fig. 12e), sandstones lithofacies, and (Fl) and (Fh) (Fig. 12f). Caprock-bearing conglomerate facies commonly overly scoured contacts at the bases of channel sequences.

The FA2 contains 0.1 to 2.5 meter thick (Fig. 12g), and 5 to 7 meter wide beds of well sorted, upper very fine- to lower medium-grained sandstone and granule- to cobble clast conglomerate (Fig. 13a). The most common matrix grain size is a lower medium sand (Fig. 13b). The predominant colors for this facies are tan, tan-yellow (Fig. 13c), red, light red, gray and light gray. Common facies stacking patterns are Gt, St, Sr, and Fl or Fh (Fig. 13d). A less frequent but still common facies stacking pattern is Gp, Sr or Sp, and Fl or Fh.

Caprock-bearing channel-fill conglomerates and sandstones occur in multi- and single story sheets in positions further from the diapir and single- and multi-story ribbons in positions near the diapir. Single story sheets occur as broad units with vertical thicknesses ranging from 2
to 5 meters thick and widths ranging from 150-200 meters overlying the salt shoulder. Multi-story units form narrow sheets with thicknesses ranging from 10-25 meters and widths ranging from 250-400 meters. Single story, multilateral ribbons are confined to positions that directly overly the shoulder with dimensions of 50-75 meters in width and 2-5 meters in thickness. Trough cross-bed sets in single story ribbons range in thickness from .1 to .5 meters and .5-1 meter in width (Fig. 13e). These beds are cyclical (Fig. 13f) and are often overlain by a thin (.1-.3 meter) laminated mudstone beds (Fig. 13g). Broad sheets also exist but are rare in diapir proximal positions and have thicknesses ranging from 10-25 meters and have widths greater than 500 meters.

**Interpretation:**

Conglomerates located at the bases of channel successions are interpreted to be deposited by traction flow as ‘channel lags’. The incorporation of carbonate caprock clasts indicates a local, proximal source of the Gypsum Valley salt wall and fine-grained subarkose to lithic arkose (Henness, 2015) sandstone indicates a distal source derived from the Uncompahgre Plateau (Stewart et al., 1972). The caprock-bearing lithofacies Gt, Gp, and Gh represent gravel bars and bedforms deposited by traction currents (Miall, 1985). The incorporation of caprock clasts likely occurred either by incision of the GV salt wall caprock or debris-flow facies (FA4) by fluvial erosive processes. Channel floor lags are interpreted where pebble conglomerates occur at the base of a channel sequence (overlying a scoured surface) and bar platforms are interpreted where conglomerates are documented at the middle or toward the top of the channel sequence (Miall, 1985). Lithofacies St, Sp, Sh, and Sr represent minor bar forms and channel fills deposited by traction currents as bed load or by intermittent suspension (saltation) (Miall, 1996).
Multi- and single story sheets have abundant planar and trough cross stratified (Gp, Gt, Sp, St) and lenticular scours filled by sandstone overlying gravel lags are consistent with transverse bar deposits in braided rivers (Miall, 1985; Hazel, 1994). Zones with abundant rounded to sub-rounded caprock clasts likely represent active channel erosion of GV salt wall or FA4 due to bank-collapse or channel incisement.

Single and multi-lateral ribbons that exhibit relatively uni-directional paleocurrent azimuths (SE), a high proportion of associated fine-grained facies (FA3) suggest off-axis deposition on the shoulder by low sinuosity, meandering streams prone to seasonal fluctuation in runoff (Hazel, 1991). Deposition on the shoulder was likely prone to frequent avulsion events from axial channels flowing adjacent to the salt wall in the Dry Creek minibasin. The relatively high percentage of Sr, climbing ripple sandstone, and interbeds of trough cross stratified sandstone (St) thin to medium thick bedforms (10-50 cm) individually capped by mudstones indicate ‘flashy deposition’ or rapid aggradation, followed by periods of channel abandonment in relatively narrow 75-150 meter wide and deep 2-5 meter thick channels. This style of deposition is interpreted to represent the monsoonal climate (Dubiel, 1991) during Chinle deposition where ‘off-axis’ streams were not active until periods of flooding. The inclusion of caprock clasts likely coincides with fluvial reworking of debris flow facies (FA4) and incorporation into channel basal lags and bar forms.
Figure 12. Outcrop photographs of FA2 within various measured sections (Figure 5). (A) Gt, trough cross-stratified conglomerate containing clasts of cap rock derived from the Gypsum Valley salt wall located within section 15, scale ruler is 15 cm. (B) Pebble conglomerate in sand matrix containing granule to pebble sized dolostone, gypsum, calcite, and mudstone clasts located within section 8. (C) Interbedded sandstone and pebble conglomerate containing granule to pebble sized dolostone, chert, and mudstone clasts located within section 8. (D) Gh, Gp horizontally laminated conglomerate overlying planar cross-stratified conglomerate located within section 14. (E) Interbedded Sp and Gt, planar laminated sandstone and trough cross-stratified conglomerate located within section 15, scale on exposed jacob staff in photograph is 1.25 meters in height. (F) Fh, horizontally laminated fine-grained sandstone and mudstone underlying Fl, rinely ripple laminated fine-grained sandstone and mudstone located within section 12. (G) Channel sequence of Gt to St located at the top of section 14.
Figure 13. Outcrop photographs of FA2 within various measured sections (Figure 5). (A) Granule to cobble-sized caprock clasts in a medium grained sandstone matrix located within section 9, scale ruler is 15 cm. (B) Tan caprock conglomerate overlying red sandstone beds located within section 7. (C) Tan-yellow color caprock conglomerate located within section 8. (D) Common stacking pattern for FA2 erosionally overlying FA4 & GV salt wall caprock, rock hammer for scale is 12.5” in length. (E) Tan carbonate caprock bed located with a single-story ribbon within section 2. (F) Cyclical caprock conglomerate beds located within section 8. (G) Fl, finely laminated mudstone bed within FA2 located in section 10.
### 4.3 Facies Association 3. Tabular shale, siltstone (FA3)

**Description:**

FA3 occurs throughout the entire study area, and is predominantly found overlying the tops of channel sequences, underlying the bases of channel sequences, and along margins of FA1 and FA2. This facies commonly represents a scoured contact with overlying FA1 and FA2 and is often interbedded with these two facies. This facies association often forms laterally extensive, tabular successions (10-100’s of meters wide) that commonly form units 15-35 meters thick. FA3 is characterized predominantly by 0.1 to 2 meter thick beds of parallel laminated claystones, siltstones, sandy siltstones, and very fine-grained sandstones. Lithofacies that comprise this facies include: climbing ripple and inclined climbing ripple fine-grained sandstone (Sr), (Fig. 14a,b), interlaminated mudstone (Fl) (Fig. 14c), horizontally laminated to massive siltstone (Fh) (Fig. 14d), dessicated to massive mudstone (Fm) (Fig. 14e). Other sedimentary features include flame structures, siltstone nodules, and dessication cracks. Common colors for this facies are dark red, brown, light pink, gray and green.

**Interpretation:**

The thin-bedded, fine-grain size, and consistent fine-scale laminations of this facies likely indicate deposition in overbank areas (Miall, 1996). The high concentration of Fl, Fh and Sr lithofacies represent deposition from suspension and from weak traction currents (Miall, 1996). Where this facies overlies channels indicates deposition from suspension in overbank areas during floods (Miall, 1996). Fining-upward successions and the presence of repeated cycles of mud cracks represents multiple episodes of flooding events and associated drying out during sub-aerial exposure on the floodplain between each flooding event (Matthews, 2007). Rippled and climbing rippled sandstones represent rapid episodic sand deposition from high-
velocity, unidirectional flows, which may represent more energetic floodplain traction currents, such as crevasse splays (Miall, 1996). In laterally extensive units, these types of deposits could represent splays or sheets associated with overbank flooding events from neighboring channels (Dubiel, 1987; Matthews, 2007). Climbing ripples in sandstone beds record episodic flooding that deposited large amounts of sediment in a short period of time (Ashley, 1970; Ashley et al., 1982), which was likely influenced by the monsoonal climate during deposition. Abundant ripple cross-laminae (Fig. 13f) have been documented in overbank deposits of modern ephemeral streams, sometimes extending un-interrupted for kilometers (McKee et al., 1967, Karcz, 1970; Stear, 1985).
Figure 14. Outcrop photographs from various measured sections of FA3. (A) Sr, climbing ripple cross-laminated sandstone located within section 5, fingernail for scale is approximately 1.5 cm wide. (B) Sr, climbing ripple cross-laminated sandstone located within section 6, scale ruler is 15 cm. (C) Fl, interlaminated siltstone and mudstone exhibiting small-scale ripples located within section 6. (D) Fh, horizontally laminated to massive siltstone located within section 6, jacob staff for scale is .5 meters. (E) Outcrop photograph showing relationship and position of photograph A located within section 5, jacob staff for scale is 1.5 meters. (F) Outcrop photograph showing relationship between photographs B,C, and D located at the top of section 6, human for scale is approximately 1.5 meters tall.
4.4 Facies Association 4. Unsorted Conglomerate Lense (FA4)

**Description:**

FA4 is confined within close proximity (< 150 m) to the margin of the GV salt wall. It crops out in isolated areas along strike of the salt wall and rarely shows bedding (Fig. 15a). This facies is characterized by lenticular, 2-15 meter thick (Fig. 15a) units of very poorly- to poorly sorted, angular- to sub-rounded, granule to boulder-sized clasts (Fig. 15b) in an upper medium- to very coarse-grained sandstone matrix. Lithofacies included in FA4 include Gmm and Gcm. Clast types include angular to sub-rounded gypsum, anhydrite, dolostone, sandstone, siltstone nodules, chocolate brown mudstones, purple siltstones, and silicified wood (Fig. 15c). Measured sections containing unsorted conglomerate are generally thicker than sections without it in similar positions to the GV salt wall. This facies is most commonly clast supported, massive, exhibits no preferred orientation of clasts, and contains no sedimentary structures (Fig. 15d). The overall color for FA4 is tan, gray, purple, green and dark red (Fig. 15e). Individual clasts are gray, tan, yellow, chocolate-brown, purple and dark red (Fig. 15f).

**Interpretation:**

The proximity of this facies relative to the GV salt wall, diversity of carbonate caprock clasts, and lack of grading, rounding, or sorting, suggest that this facis was deposited in localized, small-scale debris cones sourced from local topographic highs of the GV salt wall. The increased thicknesses of measured sections containing debris flow deposits are interpreted as paleo-valleys occurring along strike of the GV salt wall. The relatively high concentration of sediment dominated by diapir-derived detritus (Lawton and Buck, 2006) represents the erosion of soluble evaporites (gypsum, anhydrite, and halite) from the top of the diapir (topographic high) into a slurry-like, mudflow conglomerate that was subsequently drained and deposited into valleys associated with the salt wall.
The lithofacies Gmm is indicative of the process of high-strength debris flows (Miall, 1996). Flows passively occupy preexisting alluvial topography such that they assume a channelized or lenticular form (Miall, 1996). Flows are commonly lobate in plan-view form and develop lobate convex up geometries, which is preserved when the flow stops forward movement as a result of internal friction due to water loss (Miall, 1996). The lithofacies Gcm is a result of low-strength, pseudo-plastic debris flows deposited from viscous, laminar or turbulent flows (Miall, 1996).

Remaining clasts of the Moenkopi Formation (chocolate brown mudstones & carbonate caprock) on the top of the diapir and Chinle-aged sandstones and mudstones were combined into the mudflow deposits as remnant clasts. The monsonal climate during the Late Triassic created episodes of aridity and associated weathering followed by episodes of heavy rainfall that may have initiated gravity flows off the weathered diapir top and transported material down slope off toward the Dry Creek minibasin. Few debris flow conglomerates exist within the field area greater than 150 meters from the salt wall, which is likely due to re-working of this facies by rivers and streams that were deposited as FA2.
Figure 15. Outcrop photographs of FA4 from various measured sections (Figure 5). (A) Outcrop showing 6.5 meter thick deposit of FA4 located at the top of section 300 meters west of section 13. (B) Unsorted conglomerate facies containing sub-angular granule to cobble sized clasts of diapir-derived dolostone, gypsum, calcite, and chocolate brown clasts of the Moenkopi Formation at the base of section 1. (C) Boulder-sized gypsum clast in unsorted conglomerate located due west of section 13. (D) Sub-rounded pebble conglomerate containing mudstone clasts, gypsum clasts, and dolostone clasts located at the base of section 15. (E) Red color in unsorted conglomerate lense within section 14. (F) Gypsum clasts and grey color of FA4 located west of section 13.
4.5 Facies Association 5. Caprock-Bearing Heterolithic Channel Fill (FA5)

Description:

This facies most commonly contains low angle (Sl) and climbing ripple (Sr) stratified sandstone (Fig. 16a), trough (St) (Fig. 16b), planar (Sp) (Fig. 16c), horizontally (Sh) (Fig. 16d). Trough cross-stratified (Gt), planar stratified (Gp) conglomerate (Fig. 16d), interlaminated mudstone and siltstone (Fl), and dessication cracked mudstone (Fm) (Fig. 16d). Conglomerates are typically composed of sub-angular to sub-rounded pebble-sized carbonate caprock clasts derived from the GV salt wall that average 2-5 cm in diameter, with the largest clast around 10 cm in diameter in an upper fine- to lower medium-grained subarkose to lithic arkose sandstone matrix (Henness, 2015). Mud rip up are commonly combined with carbonate caprock clasts in conglomeratic lithofacies. ‘Mud chips’ are angular to sub-angular and range in size from 2-10 cm in diameter with the largest clast reaching 25 cm in diameter. The most common matrix grain size is a lower medium sand and the most common lithofacies stacking succession is Gt, St, Sr, climbing ripple sandstone and Fm. Paleocurrent azimuths range from NE, E, and SE, though are predominantly directed toward the SE (measured off of the trends of crests of ripples).

Conglomerate beds range in thickness from 25-60 cm. Upper very fine-grained to lower medium-grained sandstone beds range in thickness from 5-60 cm and are often well sorted. The sandstone beds are commonly draped by a thin layer of clay (1-10 cm) with gradational bottoms and erosional tops. Each scoured surface fines upward from interbedded pebble conglomerate into sandstone to draped mud. The mudstone (Fig. 16e) at the tops of sandstone beds commonly have dessication cracks. The entire complex contains sigmoidal beds that exhibit offlaped upper terminations and downlapped lower terminations, and a preserved cutbank. Sand-body geometry (Fig. 11) is 75 to 100 meters in width and 3.5 to 7.5 meters thick. Common colors for this facies are brown and light red.
**Interpretation:**

The cyclical, interbedded nature of conglomerates, sandstones, and dessicated mud drapes represents periodic aridity followed by ‘flashy’ discharge (Miall, 1996). Erosional scouring surfaces at the tops of mud-draped sand indicates episodic ‘slack-water’ deposition of mud followed by higher energy deposition of a conglomerate or sandstone lithofacies. The abundance of Sh and Sl lithofacies indicate the common occurrence of high energy, shallow flow near the transition from subcritical to supercritical (Miall, 1996). Preserved dessication cracks on mud drapes occur on second- and third order surfaces (bedforms) of lateral accretion sand and gravel deposits and are typical of ephemeral, meandering rivers.

This style of deposition represents sediment removed from the cutbank that becomes incorporated into the overall sediment load of the river (Miall, 1996). Material is broken down into individual grains and swept downstream as part of the bed and suspended load, and eventually deposited as bedforms and bars (Miall, 1996). The relatively high percentage of caprock clasts indicates that this process was either occurring in direct association with coherent and brecciated caprock or erosional detritus derived from the GV salt wall. Heterolithic channel complexes are interpreted to represent floodplain, levee and point bar deposits of meandering streams. This interpretation is reinforced by the presence of lateral accretion sets, fining-upward accretionary bundles, and channel plugs within a preserved cutbank (Hazel, 1991). These complexes were also are interpreted by (Dubiel, 1987; Hazel, 1991, 1994; Matthews, 2007) as partly confined suspended and mixed-load meandering-stream deposits.
Figure 16. Outcrop photographs of a well-exposed heterolithic channel fill sequence (FA5) in section 13 (Figure 5). The center photograph shows relationships and positions of the various photographs around it and human for scale is approximately 1.75 meters tall. (A) Mud-draped sandstone, thin beds (cm-scale) of mud-cracked clay gradationally overlies fine grained sandstone indicating episodic flooding events and back-filling during deposition of a channel, scale ruler is 15 cm. (B) light pink, upper fine grained sand, lenticular in nature and approximately 60 cm thick. (C) Upper fine-grained, moderately sorted sandstone. (D) trough cross stratified heterolithic channel-fill sequence representing a laterally accreting point bar with interbedded lithofacies: Gt, St, and mud drapes at the tops of each sand succession. (E) Interbedded mudstone and sandstone beds, Jacob staff for scale is 1.5 meters tall.
4.6 Facies Association 6. Fossiliferous Mudstones & Sandstones (FA6)

Description:

This facies occurs in isolated areas on the shoulder, but has a greater regional extent within the Chinle as shown by Heness (2016). FA6 is characterized by isolated, 25-75 meter wide and 0.5 to 2.5 meter thick, crudely bedded, poorly sorted fine-grained sandstones and siltstones with high concentrations of Unionid bivalves (Fig. 17a,b) (*Triasmanicola*, Parish and Good, 1987), predominantly silicified wood (Fig. 17c), concretions, preserved plant material, carbon film casts, and subordinate vertical burrows ranging from 6 to 12 cm in width and no longer than one meter in length (*Camborygma*, Hasiotis & Mitchell, 1993). It is most commonly light red, brown, and gray in color. This facies is dominantly clast supported with well preserved, intact fossils. However, fossiliferous clasts occur with sub-rounded sandstone and siltstone rip-up clasts up to 25 cm in diameter.

Interpretation:

The abundance and high concentration of fossil material including silicified wood, preserved carbon films, and thin shelled Unionid bivalves (Fig. 17a) (Dubiel, 1987a; Parish and Good, 1987) indicate a very hospitable environment to support life. The presence of concretions, articulated freshwater bivalves (Fig. 17b), a high concentration of silicified wood (i.e. stumps, limb-casts, and petrified logs) (Fig. 17c), represent seasonally re-supplied pond deposits from surrounding channels via crevasse splays (Good, 1998). The presence of elongate, vertical burrows represent crayfish that lived away from open bodies of water in areas such as proximal and distal floodplains and burrowed to the water table to depths of up to 4 meters (Hobbs, 1981; Hasiotis and Mitchell, 1993). Crayfish burrows that exhibit long length and simple architectures reflect a deep and fluctuating water table, likely related to seasonal and annual precipitation during the Triassic monsoonal climate (Hasiotis and Mitchell, 1993; Hasiotis, 2002).
Figure 17. Outcrop photographs of FA6 from measured section 15. (A) Triasmanicola, a type of unionoid bivalve that lived in a freshwater environment, scale ruler is 15 cm. (B) Articulated, Triasmanicola bivalve fossils in a mudstone matrix, GPS device for scale is 8 cm wide. (C) Meter wide stump of silicified wood.
4.7 Facies Association 7. Paleosols (FA7)

Description:

This facies is composed of claystones, siltstones, and very well sorted, very fine- to fine-grained sandstones. It is most often horizontally bedded (Fig. 18a) (20-80 cm) with ellipsoidal and tabular mottling (Fig. 18b). These zones commonly overly an unaltered, tabular horizon (Fig. 18c). An abundance of root traces, blocky peds, and rhizocretions are commonly associated with reduction haloes. Zones of highly concentrated 4-8 cm in diameter, cylindrical, sinuous, sub-horizontal burrows (Fig. 18d) with internal preserved ‘scratch marks’ represent soil-burrowing arthropods (*Scoyenia*, Hasiotis, 2002). These burrows typically exist in zones of ellipsoidal (Fig. 18e) and tabular (Fig. 18f) mottling and overly an unaltered zone (Fig. 18g). Common colors for this facies are red, dark red, gray, and white.

Interpretation:

This facies association is interpreted to represent moderately developed vertisol paleosols by the abundance of pedogenic features such as root traces, mottling, concretions, and burrow morphology types (Prochnow et al., 2006). The presence of iron-depletion around root traces and fine, blocky peds and biological traces that overly relatively unaltered parents material in an A horizon are consistent with Prochnow et al., 2006 interpretation of vertisols. *Scoyenia* trace fossils (Fig. 18h) indicate burrowing in moist, compact substrates and are common in floodplain mudstones and paleosols (Hasiotis, 2002). High concentrations of these burrows are indicative of high soil and sediment moistures approaching 100 % freshwater (Hasiotis, 2002).
Figure 18. Select outcrop photographs of FA7 in various measured sections (Figure 5). (A) Horizontally bedded mudstone with tabular discoloration located within section 11, rock hammer is 12.5’ in length. (B) Variegated vertical zones of discoloration located within section 11, human for scale is approximately 1.5 meters tall. (C) Laterally extensive horizontal tabular zones of discoloration located within section 14, human for scale is approximately 1.5 meters tall. (D) Preserved trace fossils of Scoyenia burrows in siltstone located at the top of section 10, scale ruler is 15 cm. (E) Ellipsoidal discoloration in mudstone located within section 11. (F) Vertical and horizontal discoloration interpreted to be exposure surfaces in a paleosol located within section 11, lense cap is 3” in diameter. (G) Ellipsoidal and tabular discoloration surfaces located within section 10, field notebook is 20 cm in length. (H) Amalgamated Scoyenia burrows in siltstone located at the top of section 10.
Chapter 5: Halokinetic Sequences and Stratigraphic Distribution of Facies Associations

The Bridge Canyon study area exposes a 500 meter (measured perpendicular to the trend of the salt wall) by 850 meter area of the salt shoulder (parallel to the trend of the salt wall) (Fig. 5). Neither the outboard margin nor the inboard margin of the diapir are exposed. The outboard margin lies in the subsurface and modern erosion has removed the Chinle Formation before it reaches the inboard margin of the salt shoulder (Fig. 19). However, the debris flows of FA4 and the linear edge of cliffs that mark the inboard limit of exposure suggest that the inboard wall is very close to the remaining outcrop (Fig. 5). The exposed shoulder can be divided into two domains (Figs. 19, 23, 24, and 25). The outer domain comprises the “lateral” and is influenced by minibasin subsidence so that units thicken into the minibasin. The inner domain lies on the inboard, “collar” of the shoulder. An anticlinal fold present between measured sections 2 and 6 forms the boundary between the two domains (Figs. 19, 23, 24, and 25). Steeply dipping normal faults between measured sections 2 and 6 are present inboard of the steepest curvature of the fold (Figs. 19, 23, 24, and 25). The faults extend through the Navajo Formation at the top of the outcrop. Stratigraphic thickness and facies within the Chinle Formation do not change across the faults indicating that the faults were not active during Chinle time and post-date deposition.

Halokinetic sequences are defined by Giles and Lawton, 2002 as “relatively conformable successions of growth strata genetically influenced by near-surface or extrusive salt movement and are locally bounded at the top and base by angular unconformities that become disconformable to conformable with increasing distance from the diapir”.

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Figure 19. Oblique cross section of the salt shoulder showing the distribution of facies associations, halokinetic sequences, and measured sections. Solid lines are unit contacts based on channel sands. Facies associations match colors in Table 2.
Figure 20. Diagram of halokinetic sequences displaying facies and stratal geometries within the Chinle Formation. The view is looking west to east. Flattened lines at the top of each sequences are the associated halokinetic sequence boundaries for each sequence.
Halokinetic sequences form two geometrically different end member types (Giles and Rowan, 2012): (1) hook, containing a narrow zone of deformation (50-200 meters), > 70 degrees of angular discordance, common mass wasting deposits, and abrupt facies changes, and (2) wedge, containing a broad zone of folding (300-1000 meters), low-angle truncation, and gradual facies changes. Halokinetic sequences have thicknesses and timescales equivalent to parasequence sets and stack into composite halokinetic sequences (CHS) representing time scales equivalent to third order depositional cycles. Stacked hook halokinetic sequences form tabular-CHS and wedge halokinetic sequences form tapered-CHS (Giles and Rowan, 2012).

The Chinle on the shoulder contains three wedge halokinetic sequences (WHS-1, WHS-2, WHS-3; Fig. 19) locally bounded by angular unconformities less than 5°. Each sequence: (1) contains various stratigraphically unconformable relationships with the GV salt wall (e.g. onlap and overlap), (2) thins towards the GV salt wall, and (3) contains fluvial facies changes towards the GV salt wall. The Chinle on the salt shoulder thins from 160 meters (distal to the diapir) to 44 meters (proximal to the diapir) over a distance of approximately 500 meters (Fig. 20). WHS-3 varies in thickness along strike (Fig. 21) in an undulatory fashion from 44 m to 88 meters (Fig. 21).

WHS-1 and WHS-2 each directly onlap the salt shoulder erosionally truncated updip by the overlying sequence and dip into the subsurface on the lateral of the shoulder (Fig. 19). Each displays a broad zone of upturn (> 200 meters). The broad zone of thinning and upturn with low angle truncation define these sequences as wedge halokinetic sequences (Giles and Rowan, 2012). WHS-3 overlaps the GV lateral and extends onto the collar of the shoulder (Fig. 19). WHS-3 forms a low-angle unconformity with the underlying wedge halokinetic sequences and a five degree angular unconformity with the overlying Wingate Formation. A broad zone of
folding (c. 500 meters) expresses the transition from lateral to collar (Fig. 19, 25). All combine
to suggest it also classifies as a wedge halokinetic sequence (Giles and Rowan, 2012). The three
wedges of the Chinle Formation stack into a single, tapered-CHS, which thins over a distance of
approximately 500 meters and exhibits angular unconformities of 5° or less and thus matches the
criteria of a tapered-CHS by Giles and Rowan (2012). A series of high-angle normal faults
paralleling the GV salt wall offset the Chinle, Wingate, Kayenta, and Navajo formations at the
salt shoulder.

The distribution of facies associations within the Chinle Formation adjacent to the GV
salt wall are described within a halokinetic sequence stratigraphic framework (Fig. 19). The
three wedge halokinetic sequences (WHS-1, WHS-2, WHS-3) within the Chinle Formation were
documented overlying the GV shoulder and the distribution of facies associations within each
sequence are described in detail in this chapter. Distribution of caprock-bearing channel fill
sandstones and conglomerates (size, relative percentage, and angularity of caprock clasts), fine-
grained fluvial overbank deposits, and debris-flow conglomerates vary between the halokinetic
sequences as well as with increasing distance from the GV salt wall (Fig. 19). Trending towards
the inboard edge of the shoulder, all three wedge halokinetic sequences: (Fig. 19) (1) transition
into relatively finer grained fluvial overbank dominated facies, (2) show individual channel
geometries become increasingly narrow, and (3) exhibit sand bodies with more common ribbon
geometries than sheet geometries (Fig. 19). Conversely, coarser-grained facies and thick,
amalgamated sheets are more common on the lateral of the salt shoulder (Fig. 19).
**Figure 21.** Isopach map of the Chinle Formation on the northeast margin of the GV salt wall in Little Gypsum Valley. Blue numbers and points represent measured sections, fine scale stippled line infers 20 meter topo intervals which lower confidence, coarse-scale solid black lines infer 20 meter topo intervals with high confidence, coarse scale stipple lines infer 10 meter topo intervals with lower confidence and coarse-scale solid black lines represent 10 meter topo intervals with higher confidence. The range in thickness is from 160 meters to 44 meters.
Figure 22. (A) Photo panorama looking to the northwest in Bridge Canyon of the Chinle Formation onlapping the salt shoulder, (B) 3D-modeled surface of the top of the caprock unit, still image generated in 3D Move, (C) Fence diagram corresponding to panorama in frame A. The light pink color corresponds on the photo-panorama to the top of caprock surface. The furthest north edge of the reconstructed three dimensional surface begins in the plane where the pink line is illustrated on the photo-panorama and extends to the east.
5.1 Wedge Halokinetic Sequence 1 (WHS-1)

The lowest halokinetic sequence (WHS-1) (Fig. 23) is a relatively thin unit (12.5 meters at the maximum recorded thickness at section 6 that thins towards the GV salt wall until it onlaps onto the lateral of the shoulder approximately 325 meters from the inboard margin of the diapir (Fig. 19). Onlap near the outboard edge of the shoulder where the contact with the carbonate caprock surface flattens from 25° to lower dips of 6-8° (Figs. 19 & 20). WHS-1 is gently folded and thins over the shoulder for a distance of approximately 200 meters. An angular unconformity of 2° separates WHS-1 from the overlying, thicker WHS-2 (Figure 19) and represents a halokinetic sequence boundary (HSB-2).

Outcrop of WHS-1 is dominated by FA2 the caprock-bearing channel-fill sandstone and stratified conglomerate facies (Fig. 20) contained in one single story sheet (3-5 meters thick and 150 meters wide) (Fig. 23). Pebble to cobble caprock clasts are sub-rounded to rounded and reach a maximum diameter of 15 cm. The percentage of caprock clasts in FA2 within WHS-1 is > 50 %, and is restricted to the bases of channel fill sequences. To the margins of the sheet are fluvial overbank deposits (FA3). Fluvial overbank deposits are also present locally underlying FA2, with an erosional contact. Paleocurrent measurements from the base of section 6 in WHS-1 (Fig. 19) are unidirectional, and trend to the south east. Unsorted debris flow conglomerates are not present in WHS-1.
Figure 23. (A) Photo panorama looking to the northwest in Bridge Canyon of the Chinle Formation onlapping the salt shoulder, (B) 3D-modeled surface of the top of WHS-1, still image generated in 3D Move, (C) Fence diagram corresponding to panorama in frame A. (D) Flattened surface along HSB-2. Blue line and arrow are highlighted on the photo-panorama and corresponds to the top of the 3D WHS-1 surface.
5.2 Wedge Halokinetic Sequence 2 (WHS-2)

WHS-2 reaches a maximum thickness of 27.5 meters before plunging into sub-crop. It onlaps onto the flat, inboard domain of the shoulder, termed the “collar” (Figure 19) within approximately 175 meters of the inboard diapir margin. It is gently folded and extends over the shoulder for approximately 325 meters (Fig. 24). WHS-2 overlies an angular unconformity of 2º with WHS-1 (HSB-2) and underlies an angular unconformity of 3º with WHS-3 (HSB-3). The angular unconformity overlying WHS-2 represents a second halokinetic sequence boundary beneath the overlying sequence of WHS-3 (HSB-3).

WHS-2 is composed of two relatively thin, (12-15 meter) fluvial units (defined by channel body geometry and conglomerate clast percentages) that onlap onto the shoulder further inboard than WHS-1 (Fig. 20). It contains a greater diversity of sand body geometries and a higher percentage of fine-grained fluvial overbank deposits relative to WHS-1. Near it’s inboard pinchout, FA2 channels incise directly into GV diapir carbonate caprock where the base of WHS-2 has onlapped 150 m further onto the shoulder than WHS-1 (Fig. 24). These channels have single story ribbon geometries (50-75 meters wide and 3-5 meters thick) that pinch out near the outboard edge of the shoulder where they transition into fine-grained fluvial overbank deposits (FA3) toward the inboard margin of the shoulder. The conglomerate facies found at the bases of these channels contain a relatively high percentage (> 50 %) of rounded to well rounded, pebble-size carbonate caprock clasts. The maximum recorded diameter of a carbonate caprock clast in the lower unit of WHS-2 is 14 cm. Paleocurrent measurements within measure section 6, 5, and 3 from WHS-2 are relatively uni-directional and trend south-southeast.

The upper of the two units contains channels that stack into multi-story (10-15 meters thick) and multi-lateral sheets (>200 meters wide). These channels are primarily FA1 in all but
three of the channels documented. The caprock conglomerate facies (FA2) contain < 50% carbonate caprock clasts in unit 2 of WHS-2. The channel sequences transition laterally toward the GV salt wall into fine-grained fluvial overbank deposits from the edge of the shoulder. Subordinate fluvial overbank deposits are found near channel margins and at the tops of fining-upward channel successions. Overbank outcrops are also exposed below channel fill conglomerates, commonly with a scoured erosional contact. Unsorted debris flow conglomerates are absent in WHS-2.
Figure 24. (A) Photo panorama looking to the northwest in Bridge Canyon of the Chinle Formation onlapping the salt shoulder, (B) 3D-modeled surface of the top of the WHS-2, still image generated in 3D Move, (C) Fence diagram corresponding to panorama in frame A. (D) Flattened surface along HSB-3. The purple line correlating the unit across the photo-panorama corresponds to the top surface generated in the 3D model.
5.3 Wedge Halokinetic Sequence 3 (WHS-3)

WHS-3 reaches a maximum thickness of 120 meters before diving into the subsurface and thins to approximately 53 meters on the inboard margin over a distance of approximately 500 meters (Fig. 25). The full extent of WHS-3 over the diapir is unknown due to modern erosional beveling of the salt-cored anticline, which has removed the inboard salt-Chinle diapir margin. WHS-3 overlaps the lateral and continues onto the collar/inboard part of the GV salt shoulder. WHS-3 underlies a regional unconformity with the Late Triassic Wingate Formation.

WHS-3 is composed of one thin (10-15 meter) lower unit (3A) and four relatively thick (20-30 meter) upper units (3 B, C, D, & E) (Fig. 20). The lower unit is dominantly composed of one narrow, single-story sheet of caprock-bearing sandstone and conglomerate. The sandstone and conglomerate channel facies pinch out and transition into fine-grained fluvial overbank facies at the transition point from the lateral into the collar of the salt shoulder (Fig. 19).

Westward, further onto the collar of the shoulder, the unit transitions from fine-grained fluvial overbank facies into unsorted debris flow conglomerate (FA4) that ranges from 7.5 to 12 meters thick. The unsorted cobble to boulder clast conglomerate contains sub-angular to angular caprock clasts with a maximum clast size of 30 cm in diameter. A lens of FA4 has a width of 150 meters on the shoulder where it is truncated to the east by the first unit of WHS-2. The relative percentage of caprock clasts in the caprock-bearing facies is >50 % in unit 1 of WHS-3. The second unit in WHS-3 is dominated by fluvial overbank deposits and contains one multistory channel ribbon that is laterally discontinuous (100 meters wide) and pinches out on both margins into FA3. This isolated multistoried ribbon has a scoured erosional basal contact with the underlying fine-grained fluvial overbank deposits, as well as fining-upward into fine-grained fluvial overbank deposits.
The third and fourth units of WHS-3 are dominated by thick, multistory sheets with both FA1 and FA2 channel sandstones and conglomerates on the lateral part of the shoulder. Multistory, isolated channel bodies occur more commonly on the shoulder collar. On the shoulder lateral, individual channels stack laterally and vertically into broad, multi-story sheets (widths > 250 meters) and reach total thicknesses of 15-30 meters. These sheet sandstones contain basal lags of caprock-bearing conglomerate facies. Caprock clast percentages do not exceed 50% in these channels. The channels that lack caprock bearing gravel lags, instead exhibit mud rip up clasts incorporated into the bases of the channel sequences. On the inboard shoulder collar, the dominant facies are fine-grained fluvial overbank deposits. Channels form isolated, multistory bodies and contain caprock bearing facies at the base of the channel sequence. The uppermost unit of WHS-3 is dominantly fine-grained fluvial overbank deposits that transitions to isolated, multistory bodies of FA1 on the shoulder collar at the inboard margin (Fig. 19). Paleo-current measurements (measured sections 2, 5, and 6) are the most varied among the three wedge halokinetic sequences: ranging from south to southeast.
Figure 25. (A) Photo panorama looking to the northwest in Bridge Canyon of the Chinle Formation onlapping the salt shoulder, (B) 3D-modeled surface of the top of the WHS-3, still image generated in 3D Move, (C) Fence diagram corresponding to panorama in frame A. (D) Flattened surface along the Chinle-Wingate contact. The bright pink line correlating the unit across the photo-panorama corresponds to the 3D surface generated.
5.4 Interpretation of Halokinetic Sequences

Deformation and associated impact on deposition is entirely related to the height and extent of local topographic relief over passive diapirs, which are in turn controlled by the interplay between sediment-accumulation rate and diapir-rise rate (Giles and Rowan, 2012). Tapered-CHS form when overall sediment-accumulation rate adjacent to the diapir exceeds diapir-rise rate over the timescale of a third-order (1-10 Ma) depositional sequence, where parasequence-set-scale fluctuations in sediment-accumulation rate form individual wedge halokinetic sequences (Giles and Rowan, 2012). The overall more rapid sedimentation results in depositional overlap of the diapir, forming a relatively thick roof and a correspondingly wide zone of drape folding (Giles and Rowan, 2012). Matthews (2007) and Andrie (2012) cite localized sediment-accumulation rates versus salt-rise rates to control of geometry and facies distribution in fluvial strata deposited adjacent to passive diapirs. Overall, the three wedge halokinetic sequences within the Chinle Formation represent a period where sediment-accumulation rate is greater relative to salt-rise rate. The three wedge halokinetic sequences within the Chinle Formation stack into a tapered-CHS and thus formed when overall average third-order sediment-accumulation rates were high relative to diapiric salt-rise rates during Chinle deposition. Additionally, an overall upsection decrease in salt rise relative to sediment accumulation rate occurred due to the change in halokinetic sequence geometry resulting in WHS-3 overlapping the lateral of the shoulder and burying the shoulder collar, an effect of discontinued rise of this part of the diapir initiation of subsidence of the shoulder along with the minibasin.

WHS-1 is interpreted to have developed during as an erosional period of fluvial interaction with the margin of the salt wall, relatively high topography of the GV salt wall and/or
subsidence of the Dry Creek minibasin. This interpretation is based on the presence of: (1) a low
diversity of sand body geometries; (2) a low angle onlapping relationship onto the salt shoulder,
(3) a relatively high percentage of rounded carbonate caprock clasts incorporated at the bases of
channel sequences as gravel lags, (4) a broad zone of thinning (325 meters), and (5) a low angle
halokinetic sequence boundary separating it from the overlying halokinetic sequence. Single
story, two to five meter deep and up to 75 meter wide multilateral channels with little occurrence
of vertical stacking and relatively consistent paleocurrent azimuths (E, SE) represents a braided
fluvial system that dominantly flowed toward and sub-parallel to the GV salt wall. At this stage
of deposition the fluvial system was likely influenced by topographic relief of the GV salt wall,
evident from the onlapping relationship of this sequence.

WHS-2 likely represents relatively high topography of the GV salt wall, and relatively
high levels of subsidence in the Dry Creek minibasin. This sequence contains unidirectional
paleocurrent azimuths, caprock-bearing heterolithic channel facies (FA5), a broad zone of
thinning (325 meters), and low angle unconformities bounding it. Relative percentages of
caprock clast conglomerate are lower in this sequence relative to WHS-1. Sand-body
architecture is relatively thicker and narrower than in WHS-1. A change in sand-body geometry
is observed within this sequence as isolated ribbons occur toward the inboard margin of the salt
shoulder (Fig. 19) and narrow sheets occur on the lateral of the salt shoulder. The onlapping
relationship of the sequence and compartmentalization of depositional environments within this
sequence is due to differential subsidence and topography of the salt wall. Individual channel
dimensions are narrower and thinner than channel dimensions observed in WHS-1. Lateral
accretion sets in mixed-load heterolithic channel fills represent ephemeral, gravel bed
meandering streams, and likely confined to partially confined flow also reflects higher
restriction. This indicates that during WHS-2, topography was relatively lower than during deposition of WHS-1, where the sequence onlaps further onto the shoulder.

The rate of subsidence relative to deposition in the minibasin likely decreases throughout deposition of WHS-3 based on amalgamated channels stacking into broad sheets, a higher concentration of thick, multi-lateral ribbons increasing upward on the GV shoulder, a broad zone of thinning (500 meters), and the overlapping relationship of Chinle strata with the salt wall. The presence of debrite facies in the base of the section 1 (Fig. 19) in the lowest unit of this sequence indicates exposure of the diapir and subsequent debris flows where the lowest unit fluvial facies onlaps the debris flow conglomerate at the inboard margin of the salt wall. The relatively lower concentration of caprock clast conglomerate in WHS-3 reflects fewer caprock clasts being re-worked into stratigraphically higher channel successions or intra-formational erosion due to more extensive covering of the shoulder of the diapir. The highest diversity of paleocurrents occurs in this sequence, which likely represents multiple azimuths in an unconfined braided fluvial system that flowed over the shoulder and eventually over the GV salt wall. This indicates deposition of fluvial systems that are the least confined relative to WHS-1 and WHS-2.
Chapter 6: Discussion

6.1 Controls on Halokinetic Sequences

The angular unconformities of all three halokinetic sequences reflect drape fold rotation associated with subsidence of the Dry Creek minibasin. The low angle unconformities reflect slow subsidence relative to the rate of deposition. The preservation of these sequences, along with the overlying geometries of the Wingate, Kayenta and Navajo strata reflects the abandonment of diapir rise on the shoulder and evidence that the shoulder area started to subside with the rest of Dry Creek minibasin. Continued diapiric rise, must have been accommodated along the inboard diapir margin. The onlapping relationships of WHS-1 and WHS-2 with the salt wall are indicative of the diapir forming a paleotopographic high during deposition of these units. The presence of debris flows composed of caprock clasts (FA4) and caprock clast conglomerates (FA2) in close association with the inboard salt wall indicates that the diapir was at least locally exposed, and older strata had been removed by erosion.

Controls on the overall decrease in salt rise rate relative to sediment accumulation rate through deposition of the Chinle Formation can be attributed to reduced salt flow from the Dry Creek and Disappointment minibasins. It may also be partially attributed to changes in Chinle climate. However, the effect of the differential subsidence of the Dry Creek minibasin has a more pronounced effect than climate. Channel sandstones pinch out into finer-grained overbank deposits, and thinner and narrower sand-bodies as the strata thin toward the inboard margin of the salt wall. The change from channel to overbank probably is the result of localization of channels in the subsiding minibasin, due to the subtle topographic effects of differential subsidence on the shoulder. This seems to have had profound influence on Chinle deposition as paleocurrents in the lower two WHS are directed to the southeast, nearly 180º to the regional
transport directions (Blakey and Gubitosa, 1984; Heness, 2016). This suggests a local rapidly subsiding depocenter near the southern end of the diapir, where the Dry Creek Minibasin exhibits the greatest subsidence (see below). Strata farther from the diapir exhibit northwest directed transport (Heness, 2016). Shawe et al., (1968) described a similar concentration of channel sands in the Latest Jurassic Morrison Formation in the center of the Disappointment Valley minibasin.

An overall regional increase in red color at the top of the Chinle Formation, the overlying Wingate Formation, and presence of eolian dune deposits in Church Rock equivalent members indicate an overall increase in aridity through deposition of the Chinle Formation documented by Stewart et al., (1972), Blodgett, (1982), and Blakey and Gubitosa (1984). Therefore, the upward increase in aridity through the Chinle to transition into the Wingate would indicate an overall decrease in sediment-accumulation rate, indicating climate as a subordinate control to shoulder development relative to changes in salt rise.

6.2 Moenkopi Age of Caprock

Regionally, the Chinle Formation unconformably overlies the mixed fluvial and shallow marine deposits of the Middle Triassic Moenkopi Formation, which does not crop out at Gypsum Valley, but is identified in the subsurface in Dry Creek minibasin where it is up to 256.95 meters TVD in the Double Eagle Unit well in Andy’s Mesa Field (Cole III et al., 2009). The exposed base of the Chinle Formation on the GV shoulder unconformably overlies carbonate caprock and Permian Cutler Fm. strata along the NE margin of GV salt wall, commonly represented by an erosional contact. The absence of Moenkopi outcrop at Gypsum Valley is likely controlled by the regional erosional event referred to as the Tr-3 unconformity (e.g. erosional un-roofing of the diapir) coupled with subsidence of the minibasins flanking the GV salt wall. Chocolate-brown
mudstone clasts, commonly a distinguishing feature of the Moenkopi Formation (Shoemaker and Newman, 1959) are found as sub-angular to sub-rounded clasts in the Chinle Formation lithofacies FA4 at Bridge Canyon (Fig. 19). The caprock clasts and debris flows are evidence that caprock had to have developed prior to Chinle deposition. Regional documentation of caprock development during deposition of the Moenkopi Formation is supported at Castle Valley salt wall by Shock (2012) and Foster (2015) who also describe diapir-derived caprock clasts within the Moenkopi fluvial channels. Using the concept of caprock events in the Paradox Basin outlined by Giles et al., (2013), it is suggested here that the GV shoulder caprock formed by groundwater processes associated with deposition of the Moenkopi Formation over the GV salt wall. The Moenkopi strata over the diapir were only partially stripped off the top of the GV salt wall during the Tr-3 regional unconformity. Continued erosion of the diapir incorporated these Moenkopi clasts into Chinle debris flows along with the carbonate caprock.

6.3 Flow Direction of Chinle Rivers

Regional paleocurrents were documented by Shawe et al., (1968) and Stewart et al., (1972) to flow north, west and northwest in southwest Colorado. However, along the northeast margin of the Gypsum Valley salt wall, dominant paleocurrent directions are to the south and southeast. The majority of the paleocurrent measurements are directed south-southeast (e.g. 165°-180°), which would indicate that the dominant direction of flow was into and sub-parallel to the salt wall. These data, along with the facies changes described in section 6.1 above support an interpretation that the Dry Creek minibasin was a topographic low throughout deposition of WHS 1 and WHS 2, and possibly through the entirety of Chinle deposition. Channels were diverted to flow along the diapir to the southeast, possibly in a drainage along the Dry Creek minibasin between the Gypsum and Paradox Valley anticlines. This observation is supported by
the presence of caprock and gypsum clasts incorporated into the bases of channels, which were likely eroded clasts derived from the topographic high of the salt wall. This is consistent with findings in the Moab Valley salt wall by Matthews et al. (2004). When the deposition rate exceeded subsidence in the study area, channels flowed along the margin of the diapir, eroding it as is seen in the channels onlapping the diapir in WHS 1 and WHS2. During deposition of WHS-3, the Chinle strata covered the now inactive shoulder, thus allowing the relatively less restricted braided fluvial system with the most varied paleocurrent directions to flow across the shoulder as salt rise had ceased on the shoulder.

6.4 Possible Mechanisms to Produce a Shoulder Geometry

This study evaluated four possible mechanisms of GV shoulder formation (Figure 26). The mechanisms are: (1) erosional thinning over the diapir triggering piston-like uplift inboard of previous diapir edge, (2) local dissolution of salt and subsequent collapse of overlying strata along the diapir margin, (3) local minibasin subsidence and the formation of a rim syncline on the margin of the diapir, (4) local erosion and incisement of the diapir top by onlap of successive units, and local subsidence of the flanking minibasin. This process indicates a process of growth of the minibasin toward the diapir and abandonment of the previous diapir margin. However, the development of a salt-shoulder structure is not limited to one of these mechanisms and may involve any number and combination of the four listed.
Figure 26. Conceptual models illustrating proposed mechanisms for the development of salt shoulders. Gray colors represent the diapir and tan colors represent flanking strata. The various mechanisms are: (A) salt rise at an inboard position by piston-like uplift after Hearon (2014b); (B) local dissolution of salt and subsequent collapse of overlying strata along the diapir margin; (C) local loading, salt withdrawal, subsidence, and the formation of a rim syncline on the margin of the diapir; and (D) local erosion and incision of the diapir top, caused by a regional unconformity involving initial beveling of the top of the diapir followed by onlap of successive units and abandonment of the diapir margin.
The oldest Chinle strata exposed on the shoulder (WHS-1 and WHS-2) exhibit erosional onlap of the salt. The burial of the margin of the salt was probably a recurring event during diapir growth, and normally, a reduction in the rate of deposition or increased rate of diapir rise would have folded the Chinle strata up against the diapir flank, as is true in southeastern Gypsum Valley, they would have been onlapped by younger strata. However, diapir rise along the outboard diapir margin did not continue at this location after deposition of WHS-2. Instead, WHS-1 and WHS-2 were rotated to 25 degree dips by continued minibasin subsidence while diapiric rise continued inboard of the salt shoulder (Figs. 19 & 29). Episodic erosion of caprock indicates that at least episodically, the caprock of the diapir formed a topographic high at the inboard margin of the diapir adjacent to the shoulder. WHS-3 covered a larger area of the shoulder and developed the youngest sequence in a halokinetic growth monocline overlying the shoulder. The rotation of wedge halokinetic sequences 1 and 2 indicate that the diapirism was active on the outboard margin of the GV salt wall during and after their deposition. However, diapirism discontinued before or during deposition of WHS-3, evidence for the formation of the shoulder, the shoulder began to subside as part of the minibasin, creating a halokinetic growth monocline overlying the shoulder (Fig. 29). Alternatively, the sequences began to subside along with the Dry Creek Minibasin as salt rise continued at an inboard position. Continued diapirism and exposure of the caprock inboard of the shoulder is evident in the caprock clast conglomerates and debris flows within WHS-3 and the overlying Wingate and Kayenta formations.

The development of the Gypsum Valley salt shoulder is documented within this study to have resulted from a combination of the mechanisms illustrated in Figure 26. The width of the shoulder does not vary considerably along strike of the GV salt wall, maintaining a consistent
width of .5 km for approximately 19 km along strike. This geometry and the facies present on the shoulder (i.e. dominated by overbank deposits, lakes/ponds) makes erosion by fluvial incision an unlikely driving mechanism to have developed the salt shoulder. However, dissolution causing localized collapse of salt is a likely driver to create topographic lows in isolated positions along the salt shoulder (i.e. lakes/ponds; Heness, 2016). The shoulder is thought to have developed during the beginning of deposition of WHS-3 due to the presence of FA4 debris flow at the base of this sequence, indicating there was topography on the inboard margin of the salt wall. However, the development of the shoulder cannot be limited to timing exclusively at the beginning of deposition at WHS-3 because of the presence of the onlapping relationship of fluvial overbank deposits (FA3). The possibility of shoulder formation prior to WHS-3 could be explained by the re-working of FA4 (debris flow conglomerates) by erosion and inclusion into channel facies FA2, observed in both WHS-2 and WHS-1.

6.6 Formation of the Shoulder Anticline and Normal Faults

The change from 25 degree dips into the Dry Creek Minibasin to gentle southerly dips (4° to 7°), creates an anticline that formed because the outboard half of the shoulder continued to rotate with minibasin subsidence, whereas the inboard domain did not. The differential rotation between sequences would have put tensional stresses on the strata deformed on the lateral domain of the shoulder. High-angle normal faults mark the change from the lateral to the collar part of the salt shoulder and offset strata from the Chinle Formation through the overlying Entrada Formation (Fig. 5). These have been interpreted to be caused by local dissolution during breaching of the salt wall (Trudgill, 2011). The faults mark the change from the outboard shoulder domain, where Chinle strata are rotated, to the inboard domain, on which strata exhibit local unconformites and basins, but do not exhibit growth into or away from the diapir. High-
angle faults overlying the shoulder are restricted to the lateral, or outboard margin of the shoulder. The faults are preserved in strata overlying the shoulder along the northeastern flank of the GV salt wall along strike for 12 miles (19 km) and are constrained to the 500 meter extent of the shoulder. The collapse blocks of the overlying Wingate, Kayenta, Navajo are indistinguishable in certain positions along collapsed fault boundaries. The normal faults exhibit steep dips (85-88º) toward the GV salt wall. Fault planes are shared in Chinle through Navajo formations indicating timing of the faults were post-depositional, and thus likely related to dissolution and subsequent gravitational collapse into the topographic low of the GV salt core. The combination of the geometry of the drape fold monocline on the shoulder and subsequent post-depositional gravitational collapse faulting led to the overall antiformal geometry observed in the field. This combination has resulted in the “shoulder-roll” trap geometry discussed in the introduction.
Figure 27. Conceptual illustrations of time steps and for the development of the Gypsum Valley salt shoulder. Time sequence as follows: (A) Development of Moenkopi-aged caprock, (B) Beveling at the Tr-3 regional unconformity, (C) Onlap of WHS-1 of the Chinle Formation, (D) Continued salt rise and subsidence of WHS-1 and progressive onlap of WHS-2, (E) Continued salt rise and subsidence of WHS-2 and overlap and covering of GV salt wall at deposition of WHS-3, (F) Continued salt rise at an inboard position and development of piston-like uplift and associated fault, and abandonment of the diapir margin halokinetic-drape folding on the outboard margin of the shoulder,
Figure 28. Conceptual illustration of Chinle fluvial paleoflow on the northeast margin of Gypsum Valley salt wall, where sub-regional streams and rivers flowed dominantly west and south west, and measured paleoflow azimuths in the field area were dominantly south-southeast.
Chapter 7: Conclusion

7.1 Chinle Depositional Environments

Deposition of the Late Triassic Chinle Formation adjacent to the GV salt wall influenced the distribution of facies and the geometry of onlapping and overlapping strata. Chinle facies depositional assemblages documented on the shoulder include: (1) non-caprock bearing channel-fill sandstone and stratified conglomerate, representing both meandering rivers and streams as well as relatively unrestricted braided fluvial systems, (2) caprock-bearing channel-fill sandstone and stratified conglomerate, representing meandering streams and relatively unrestricted braided fluvial systems that incorporated eroded diapir-derived detritus (3) tabular shale, siltstone, representing overbank deposits (4) unsorted conglomerate lenses, representing debris flows (5) caprock-bearing heterolithic channel fill, representing meandering rivers with a preserved cutbank (6) fossiliferous mudstones and sandstones, representing ponds and or small lakes and (7) paleosols, representing preserved soil development in overbank areas.

7.2 Halokinetic Sequences

The Chinle Formation on the shoulder forms three wedge halokinetic sequences that progressively onlap and overlap the GV salt shoulder. The Chinle Formation thins from approximately 160 meters at the outboard margin of the shoulder to 53 meters at the inboard margin. The lowest of the three halokinetic sequences (WHS-1) thins from 12.5 meters to an erosional pinchout over a distance of 200 meters, and is distinguished from the overlying halokinetic sequence (WHS-2) by an angular unconformity of 2°. The WHS-2 reaches a maximum thickness of 27.5 meters and onlaps onto the shoulder over a distance of 325 meters, and is distinguished from the overlying halokinetic sequence by an angular unconformity of 3°. The third wedge halokinetic sequence reaches a maximum thickness of 120 meters and thins to
53 meters where it onlaps and overlaps onto the shoulder. The three wedge halokinetic sequences of the Chinle Formation stack into a tapered-CHS. The tapered-CHS displays a broad zone of thinning (> 500 meters) and upturn, and low-angle unconformities at halokinetic sequence boundaries (< 5°).

An overall decrease in grain size, sand-body width/height, and channel width/height was documented from diapir distal to diapir proximal positions. Proximal facies are finer grained and are packaged into single and multi-lateral, single story ribbons. Facies assemblages positioned near the outboard margin of the shoulder are packaged in single and multi-story sand sheets. The relative percentage of caprock clasts decreases upward through the Chinle, where channel bases average caprock-bearing percentages greater than 50 % in the first two wedge halokinetic sequences and < 50 % in WHS-3. Above HSB-3 (Figure 19), the relative percentage of caprock clasts decreases below 50 %. Paleocurrent measurements are recorded to be generally counter-regional (Fig. 19) to regional paleo-flow directions (NW) and at least 90° from sub-regional stream and river flows (SW-W), likely influenced by subsidence of the Dry Creek minibasin (flanking the GV salt wall to the NE). Paleocurrent measurements are relatively consistent in WHS-1 and WHS-2 (south-southeast), however, within WHS-3 paleocurrent measurements are relatively more variable (northwest, south-southeast, east).

7.3 Fluvial Architecture

Fluvial architecture of the Chinle Formation on the salt shoulder is controlled by the interplay between salt rise and sediment accumulation. The rate of salt rise relative to sediment accumulation rate decreased through deposition of the Chinle, which is evident by the onlapping relationships of WHS-1 and WHS-2 with the GV salt wall and overlapping of WHS-3 and overall increasing aridity in the Late Triassic climate documented regionally. Sheet sandstone
facies transition laterally toward the inboard margin to overbank fine-grained facies and smaller, narrower isolated ribbon sand bodies. Single story sand bodies with relatively unidirectional paleocurrents were documented in the first and second halokinetic sequences, whereas multi-story, amalgamated channels and a higher diversity of sand bodies as well as paleo-currents indicates a relatively less restricted fluvial system flowing across the GV salt wall. Coarser grained sand facies and sheets are found in the lateral domain of the salt shoulder, whereas fine-grained facies and ribbons are found in the collar domain of the shoulder.

7.4 Development of GV Shoulder

Gypsum Valley salt shoulder (Fig. 27) is interpreted to have developed during or before deposition of WHS-3 in the Chinle Formation. The sequence of events that most likely explains the formation of the GV shoulder-roll begins with the development of a Moenkopi-aged caprock on the GV salt wall through meteoric groundwater influx followed by regional beveling at the end of Moenkopi/beginning of Chinle deposition, documented as the Tr-3 unconformity. An overall decrease in salt rise rate during Chinle deposition led to progressive onlap and eventual overlap over the shoulder. Rotation of WHS-1 and WHS-2 into the Dry Creek minibasin flanking the still active northeast margin of the GV salt wall occurred, evident by halokinetic sequence formation. Subsequent overlap by WHS-3 documents the abandonment of the northeast margin of the GV salt wall and initiation of subsidence along with the Dry Creek minibasin. Salt rise continued at an inboard position creating topography explaining the presence of the FA4 debris flows observed at the inboard margin of the salt wall documented in the basal unit of WHS-3. This relationship records the timing of shoulder formation as rotation of the three sequences discontinued, thus creating a drape fold monocline overlying the shoulder. This timing has an effect that is two-fold: the outboard margin of the shoulder discontinues drape
folding strata, as the drape folding zone transitions to the inboard margin of the salt wall, where salt continues to rise creating a new zone of drape folding (Fig. 29). The older zone of drape folding (drape monocline zone 1) subsides with the minibasin, and rotation of strata continues inboard as salt rise and subsidence on the shoulder couple to create the younger zone of halokinetic folds in younger strata (i.e. drape monocline zone 2) (Fig. 29). The antiformal geometry formed after deposition of the Chinle Formation. Post-depositional faulting due to dissolution collapse of strata overlying the shoulder occurred after deposition of the Navajo Formation. Chinle through Navajo strata share common high angle normal fault planes that collapsed strata overlying the shoulder that are dipping into the salt wall. The mechanism of shoulder geometry formation is controlled by a fault at the inboard margin, allowing the abandonment of the shoulder and explains the relationship between inboard salt rise and coupled subsidence of the shoulder with the Dry Creek minibasin.
Figure 29. Conceptual illustration showing the effect of shoulder development and abandonment of the diapir margin. This process creates an inboard step of drape folding and abandonment of drape folding on the outboard diapir margin. Drape fold monocline zone 1 is shown overlying the collar and lateral domains of the shoulder. This zone constitutes the area of the shoulder as it subsides with the minibasin after shoulder formation. Continued salt rise at an inboard position on the diapir becomes the new zone for drape folding, thus continued rotation of outboard sequences overlying the shoulder become abandoned. This leads to the creation of drape monocline zone 2. Debris flow conglomerate (FA4) represented directly above the collar is evidence for topography during deposition of WHS-3 and likely beginning of shoulder development.
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Curriculum Vita

Joshua Coleman McFarland was born in Fort Hood, Texas on April 30th, 1990 and raised in Alexandria, Virginia. The second of three children of Richard and Peggy McFarland, the McFarland’s relocated three times before ending up in Northern Virginia, the result of being a military family. Mr. McFarland graduated from West Potomac High School in 2008 where he competed as a three sport athlete. In the Fall of 2008, he enrolled at Montana State University in Bozeman, Montana. He worked for the Forest Service on a wilderness trail crew in Wisdom, Montana, a production geologist for Stillwater Mining Company, and as a raft guide on the Yellowstone River. He qualified for the dean’s list in the Spring of 2010 and Fall of 2012. In the Fall of 2013, he enrolled at the University of Texas at El Paso (UTEP) to pursue a Master of Science degree in geological sciences. While a graduate student, he received research grants from the West Texas Geological Society, Roswell Geological Society, Geological Society of America, American Association of Petroleum Geologists, and Rocky Mountain Association of Geologist’s Stone-Hollberg Award for excellence in structural geology. He served as the vice president for UTEP’s AAPG chapter, worked as a teaching and research assistant at the university and helped achieve a third place finish for the UTEP Imperial Barrel Association team in Spring 2014. Josh completed internships with Marathon Oil Company in Oklahoma City, Oklahoma and BP in Anchorage, Alaska and currently works full time for BP Exploration Alaska.

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