EVOLUTION OF MINIBASIN STRATIGRAPHY ADJACENT TO THE PINE RIDGE SALT DIAPIR; PARADOX BASIN, SE UTAH

by

Cheryl Fountain
A thesis submitted to the Faculty and the Board of Trustees of the Colorado School of Mines in partial fulfillment of the requirements for the degree of Master of Science (Geology).

Golden, Colorado

Date: _________________

Signed: _________________________
Cheryl Fountain

Signed: _________________________
Dr. Bruce Trudgill
Thesis Advisor

Golden, Colorado

Date _________________

Signed: _________________________
Dr. M. Stephen Enders
Professor and Department Head
Department of Geology and Geological Engineering
ABSTRACT

The architecture of salt-controlled minibasins affects the distribution of hydrocarbon elements within a system, and may be different across associated minibasins adjacent to salt bodies. Therefore, it is vital to accurately correlate stratigraphic sequences in adjacent minibasins, as mistakes can be costly. This can be a difficult task in salt-controlled basins due to the transient nature of the evaporites, which in turn affects accommodation and depositional patterns. The task is complicated in areas with sparse data density. This can be seen in two previously published papers over the Pine Ridge salt diapir, SE Utah, which have conflicting interpretations due to lack of subsurface control. One interpretation is based on well data, while the other is based on 2D seismic data.

This study incorporates a 3D seismic dataset, regional 2D seismic lines, well data, and field data over the Pine Ridge Diapir, allowing a detailed analysis of minibasin stratigraphy both temporally and spatially. Our dense dataset in this study suggests that the evolution of the Pine Ridge diapir was significantly different to that of the Salt Valley salt wall to the north. Depositional geometries at Pine Ridge show that the salt rise rate was equal to the sediment accumulation rate. A lack of supra-salt faults indicates that the salt body was consistently below the surface, and did not experience salt dissolution or extension during its evolution. Salt geometries and isopach stacking patterns give insight into the evolution of the diapir. During interpretation it was determined that salt movement was not significant during the formation of the early to mid-Pennsylvanian evaporites, and continued to be quiescent into the late Pennsylvanian. However, rapid movement occurred during deposition of the lower Cutler beds, leading to drastic thickness changes across the minibasin. Salt movement generally stayed consistent for the rest of the diapir passive growth history with the exception of a few increases
of diapir rise rate versus aggradation during the deposition of the White Rim, Moenkopi, and Chinle formations. The findings at the Pine Ridge salt diapir can be applied to other salt diapirs in similar tectonic settings.
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ACKNOWLEDGEMENTS

This research would not have been possible without the discussion, insights and support of many colleagues, professors and professionals. Although they are too numerous to name, I would like to take a moment to highlight a few people that have had a great impact on my research.

First and foremost I would like to thank Bruce Trudgill for teaching me everything I know about salt. I would also like to thank Bruce for providing me with such an interesting topic for research, the many discussions, and providing so many opportunities. I would also like to thank my committee members Mary Carr for her helpful insights, willingness to discuss the research, and for a great literature list; and Donna Anderson who’s questions were always very perceptive.

Field work and much of the research would not have been possible without Jessica Jobe and Elizabeth Wilson. The many discussions in and out of the field have shaped my understanding of the basin and its development. A huge thank you to Tim McIntyre who was always willing to hear about my ideas and question their validity. Much of the progress made was thanks to you. Thank you to Oscar Valequez for your willingness, patience, and time devoted in helping build a velocity model. Your expertise and skill were invaluable. Thank you to Sarah King for sharing your knowledge of Petrel. Thank you to Wes Bucker for making InDesign approachable.

Thank you to my previous mentor, Dave List, who sparked my interest in the Paradox Basin and seismic interpretation, encouraged me to further my education, and provided me with an extensive literature list. I would also like to thank the individuals at Chevron who gave me a
lot of good feedback.

    I would also like to thank the organizations who helped fund the research through scholarships including the SEG foundation and Marathon Oil. A huge thank you to the Bill Barrett Corporation who was generous enough to donate their 3D data set to us. Thank you also to IHS for their donation of LAS files.

    Finally, and most importantly, I would like to thank my husband, Ryan Fountain, for his support and keeping me sane through the process.
CHAPTER 1
INTRODUCTION

1.1 Research Objectives

The architecture of salt controlled minibasins affects the distribution of hydrocarbon elements within a system including reservoir and seal position and trap geometries (Peel, 2014). Accurately modelling and understanding how the salt withdrawal minibasins relate both temporally and spatially is crucial in oil and gas exploration.

The Pine Ridge Salt Diapir and its associated minibasins exposed in southeast Paradox Basin affords the opportunity to correlate depositional packages across the diapir. Additional correlative minibasin comparisons drawn from the Salt Valley Salt Wall to the northwest gives a more complete picture of minibasin development in three dimensional space. This research has implications for modelling minibasin evolution as a whole, and can be used as an analog in understanding the history of minibasin development of other salt-controlled basins.

1.2 Significance of the Problem

Correlating formations in the subsurface can be a difficult task, especially when salt is continuously moving and evolving through time and space. Accommodation created by salt evacuation may not be equivalent across minibasins, therefore, sediment may be deposited at a deeper level in one minibasin while depositing at a much shallower depth in an adjacent minibasin.

The story is further complicated by the transitional nature of salt; emplacement causing deposition center migrations, and salt evacuation triggering subsidence of overlying units (Kluth, & DuChene, 2009). Hypothetically, one minibasin may experience collapse while a neighboring minibasin may experience emplacement of salt. This leads to discrepancies in depth correlation; the same depth across minibasins may contain sediments of widely different ages.

The incongruities across minibasins can generate costly problems during hydrocarbon exploration, either by over or undershooting a target, or drilling in a location where the target
formation was not deposited. Making it essential to understand how adjacent minibasins behave during the evolution of adjacent salt bodies in order to increase accuracy during hydrocarbon exploration.

1.3 Study Area:

The following research is located within the Paradox Basin, which is a northwest to southeast elongate (Chidsey et al, 2015) intraforeland flexural (Barbeau, 2003) basin defined by the extent of Pennsylvanian evaporite deposits of the Paradox Formation (Baars & Stevenson, 1981). The majority of the basin lies within southeast Utah and southwest Colorado, but a small portion reaches New Mexico (Figure 1.1). The basin is asymmetric, with the deepest point residing next to the Uncompahgre uplift to the northeast and becoming shallower to the southwest (Whidden et al., 2012). The basin is bounded to the northeast by the Uncompahgre Uplift, the San Juan Dome to the East, and the San Rafael Swell to the northwest (Heyman et al., 1986; Kelley, 1955) (Figure 1.1). Major salt tectonics are confined to what is referred to as the “Paradox Fold and Fault belt” (Kelley, 1955; 1958) which is located in the northern part of the basin (Figure 1.1).

The study specifically focuses on the Pine Ridge Diapir and associated minibasins located in San Juan County, Utah, southeast of the La Sal Mountains (Figure 1.2). The Pine Ridge diapir is oriented southeast to northwest and resides between the Moab Valley salt wall and the Gypsum Valley salt wall (Figure 1.2). Comparisons are drawn with the Salt Valley Salt Wall located in Grand County, east-central Utah, northwest of the Pine Ridge diapir. Figure 1.2 outlines the Pine Ridge diapir and Salt Valley salt wall study areas and the data used. Figure 1.3 is a detailed dataset for the Pine Ridge study area.

Access to the study area can be obtained via Highway 46 which runs east to west, and transects the Pine Ridge Diapir and the adjacent minibasins (Figure 1.4). However, much of the highway near the study area lies on private land, and therefore, outcrop data was only obtained immediately adjacent to the highway, and did not extend further. Additional site access was
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available along unpaved dirt roads from Buckeye Campground, UT along Forest Service roads 0371 Buckeye Rd. 4114, 4111, and 0072.

1.4 Data sets:

The research at Pine Ridge utilizes publicly available well logs and well reports, a proprietary three dimensional (3-D) seismic survey, and field data (Figure 1.3). Proprietary two dimensional (2-D) regional lines were viewed, but will not be shown in this study due to confidentiality restrictions. The comparative Salt Valley data set consists of publicly available well logs and well reports, a proprietary 3-D seismic cube and proprietary 2-D regional seismic lines. Each unique data type provides different information, and when combined provide a relatively complete data set around the areas being studied.

The well logs establish formation depth profiles and are subsequently used to calibrate the temporal seismic data set. The 3-D seismic cube exhibits reflectors that when tied to wells can be interpreted as formations. In this way, stacking geometries of minibasin deposits as well as the relationship of stratigraphic and diapiric contacts can be inferred. The seismic is used to derive evolution of events in the Pine Ridge northern minibasin, and understand relationships of adjacent minibasins of the Salt Valley salt wall. Regional 2-D seismic lines in Salt Valley give insight to the salt influence on depositional patterns. Finally, field data collected along the strike of the Pine Ridge Diapir constrains timing of events. Figure 1.2 shows the distribution of data types over the two study areas.

1.4.1 Well Logs

Available well logs and well files are used to evaluate lateral changes in formations, generate cross sections, and create models of the Pine Ridge Diapir. Well logs in the Salt Valley area are used to tie formation tops to seismic horizons.

Twenty-one wells are deemed important to the study of the Pine Ridge Diapir due to either close proximity to the 3-D survey, or position along the strike of the salt structure (Figure 1.3). Two wells were drilled within the seismic cube: the TXC-Huber Federal 1-15
Eighty three wells were used for the interpretation of the secondary study area at Salt Valley salt wall and associated minibasins. Forty two wells are located within the 3-D seismic, 41 regional wells are located in close proximity to the 3-D seismic and 2-D seismic lines. Well files are available for all 81 wells; raster well logs are available for 35 wells; 16 wells have associated LAS files. Table 7 in the appendix contains the wells used in the Salt Valley salt wall area.

Stratigraphic horizons were chosen on the well logs, this lithostratigraphy, inferred environmental setting and hydrocarbon potential is shown in F, (modified from Hintze & Kowallis, 2009; Trudgill, 2011; Whidden, 2014; and Rasmussen, 2014) however, these formations are discussed in more detail in section 2.3.

1.4.2 3-D Seismic

The Pine Ridge 3-D seismic cube is located on the northern flank of the Pine Ridge Diapir (Figure 1.3), and will be used to interpret depositional geometries of the northern minibasin. The seismic volume was donated to the Colorado School of Mines by the Bill Barrett Corporation. The seismic cube extends 39,050 ft by 33,550 ft or about 7.4 mi by 6.3 mi, and penetrates the subsurface to a two-way time interval of 3,000 ms. However, data are only present in dimensions of about 15,740 ft by 20,300 ft or 3.8 mi by 2.9 mi due to edge effects during acquisition.

The Salt Valley 3-D seismic cube is located over the Salt Valley Diapir and accompanying minibasins. The seismic cube extends 35,700 ft by 59,200 ft or about 6.8 mi by
**Figure 1.5:** Stratigraphy of the Pine Ridge and Salt valley Salt wall, depositional environment and hydrocarbon potential. (Modified from Hintze & Kowallis, 2009; Trudgill, 2011; Rasmussen, 2014; Whidden et al., 2014.)
11.2 mi, and penetrates the subsurface to a two-way time interval of 3,000 ms.

Salt is defined by poor seismic imaging due to steeply dipping salt wall flanks, as well as large velocity contrasts between the salt and the surrounding sediment, resulting in high impedance values between salt and other rock types. The sediments next to salt can also show lateral discontinuity and change in dip near the salt bodies (Jones and Davidson, 2014).

Interpretations of the seismic cube depositional geometries lead to an understanding of the timing of the minibasin development and salt evolution. The interpretations were then used in the reconstruction of the depositional history in the northern mini-basin off of the Pine Ridge Diapir, and in correlating depositional packages in the Salt Valley minibasins.

1.4.3 2-D Seismic

ConocoPhillips shot several proprietary regional 2-D lines in the area of interest. One of the regional lines crosses directly over the Pine Ridge minibasins and diapir. 2-D lines were observed in person, and notes were taken on the overall geometry of the minibasins. The Salt Valley area contains twenty regional 2-D seismic lines, which outline the regional extent of the salt bodies’ relationship to deposition.

1.4.4 Field Data

Field data were collected along the strike of the Pine Ridge Salt Diapir in order to help constrain the timing of salt movement within the minibasin. Measurements include sample descriptions to accurately identify formations and strike and dip measurements to understand lateral continuity of the formations. Outcrop exposure was poor near the Pine Ridge diapir, which limited the data density. Outcrops at the surface represented the Brushy Basin, Dakota, and Burro Canyon formations. The Burro Canyon formed cliffs which could be measured, while the Brushy Basin and Dakota had poor exposure due to erosion. The La Sal Quad 30’x 60’ Quadrangle by Doelling, (2004), was heavily referenced during field work.
Figure 1.6: Published interpretations of the Pine Ridge Diapir and associated minibasins A.) Rasmussen’s 2014 cross section over the diapir B.) Kluth & DuChene’s 2009 cross section across Lisbon Valley, Gypsum Valley and Paradox Valley. The two interpretations show a vastly different evolution of the Pine Ridge Diapir and minibasin development both in timing and minibasin geometries.
1.5 Previous Studies

Two studies and a United States Geological Survey (USGS) quadrangle map completed near the Pine Ridge Diapir were used extensively during the course of this research. The studies were Donald L. Rasmussen’s Utah Geological Association publication “Namakiers in Triassic and Permian Formations in the Paradox Basin (USA) with Comparisons to Modern Examples in the Zagros Fold Belt, Iran (2014),” and Kluth & DuChene’s “Late Pennsylvanian and Early Permian Structural Geology and Tectonic History of the Paradox Basin and Uncompahgre Uplift, Colorado and Utah (2009).”

Rasmussen (2014) used well log data to generate a cross section across the Pine Ridge Diapir and flanking minibasins (Figure 1.5). The interpretation suggests that the Honaker Trail and Moenkopi formations thicken in the SW minibasin, and shows drastic thickening of the White Rim formation in the SW minibasin. The paper also interprets a salt glacier or namakier as present in the SW minibasin. Basement faults were interpreted as reverse and post-salt counter-regional faults were also interpreted in this section. Geometries interpreted in the cross section suggest that the growth of the Paradox Valley salt wall and Pine Ridge salt wall coincide. It also suggests that minibasin fill occurred at relatively the same time.

Kluth & DuChene (2009) used 2-D line drawings to construct a history of the Pine Ridge Diapir and its minibasins (Figure 1.5). Basement faults are interpreted to be normal with no post-salt faults. Interpretations were based solely on seismic, and therefore lithostratigraphic changes between minibasins were not noted. Geometries interpreted suggest that minibasin deposits cannot be correlated across the diapir, and the minibasins have very different depositional histories. The Paradox Valley and associated minibasins had completely formed before the development of the Pine Ridge diapir.

The Utah Geological Survey Map 205, (Doelling, 2004), is a detailed map that focuses on the center of the Paradox basin, and includes the Pine Ridge northeastern minibasin. This survey was used extensively in field mapping in order to understand the relationship of outcrop
contacts. The data was later used to interpret formation geometries of units above the seismic datum, then subsequently used in reconstructions.

1.6 Software

The Fieldmove Clino App for the iphone 5C is a digital mapping tool that can record accurate satellite location, strike and dip measurements as well as Geographic Positioning System (GPS) measurements which positioned outcrop photos. This was used in conjunction with a Brunton compass for field measurements. Formation tops were picked from well logs using both Petra Version 3.8.3 and Petrel 2015 software packages. Interpretation of seismic volumes were completed in Petrel 2015. Structural restorations were completed using Midland Valley’s Move 2015.

1.7 Data Sources

Colorado and Utah Township and Sections were acquired from the Public Land Survey System’s (PLSS) Geocommunicator website, updated on December 2009. Colorado and Utah Counties, State outlines, hydrology, roads, railroad, and urban areas were acquired from Tiger Geodatabases (U.S. Census Bureau, 2015). Colorado well spots and raster logs were attained from the Colorado Oil and Gas Conservation Commission. Utah well spots, raster logs, Digital Elevation Map, and land ownership were acquired from the State of Utah Oil and Gas Program. Colorado land ownership maps were obtained from the Bureau of Land Management (B.L.M.).
2.1 Theory of Minibasin Deposits

Salt has several unique properties which allow it to behave differently than most sediments (Schoenherr et al., 2007), leading to many of the interesting structural styles that are well seen in the Paradox Basin.

Salt has a low permeability and porosity at very minimal burial depths, and porosity is almost non-existent by about 330 feet or about 100 m (Schoenherr et al., 2007). In addition salt is incompressible. This means that as burial continues, salt has a consistent density of 2.160 g/cm³, whereas most other sediments continue to compact with depth and increase in density. With deep enough burial, density inversion occurs and the salt becomes less dense than the overlying sediment. Typically this occurs between 5,450 ft and 9,850 ft or about 1,660 m to 3,000 m, depending on the overlying lithology (Schoenherr et al., 2007).

However, the density contrast between materials is not as influential as it was once believed, and has minimal effect on a rising diapir. One such example is the Paradox Basin, where the development salt walls began to form when overlying sediments were still thin, and less dense than salt (Hudec et al., 2009). Salt rise is instead primarily dependant on another unique property of the salt: it is mechanically weak and flows like a fluid under geologic time scales (Ratcliff, 1993; Jackson & Vendeville, 1994; Trudgill, 2011). Because of this crucial property of salt, as it is buried, if a differential load is created, the salt tends to flow. However, because salt is so weak, one of several factors has to be in play for movement to occur: available accommodation, an open conduit, or the differential pressure must exceed overburden strength (Vendeville, & Jackson, 1992).

Hudec et al. (2009) proposed several mechanisms that may drive salt movement including density driven subsidence, shortening, extension, erosion of surface salt, topographic loading on a slope, and sub-salt deformation. Topographic loading on a slope may be a driving
mechanism in the Paradox Basin. However, for migration of a minibasin down-dip, enough sediment must be deposited in order to keep the surface at a regional dip (Hudec et al., 2009). Peel (2014) completed forward modelling experiments in which he studied the initiation of salt wall and minibasin development with no heterogeneity in the system. The study found that the development of structures could be initiated by neighboring salt movement alone, and continued salt development progrades across the basin (Peel, 2014). However, the salt structures seen in the Paradox Basin more closely resemble models shown in Ge et al. (1997) using progradation in combination with basement structures to drive diapir growth. Ge et al. (1997) conducted research on scaled physical models to understand how the growth of a salt diapir may be triggered by progradation (Figure 2.1). In an experiment involving only progradation, the development of distorted salt, and remnant salt pillows formed, however prominent diapiric salt structures did not develop. When the experiment was repeated with the addition of basement steps along with progradation, then diapiric salt structures formed. The model also showed that as the experiment progressed the age and the complexity of the salt structures decreased basinward, which is consistent with what is seen in outcrop and well logs within the Paradox Basin (Ge et al., 1997).

Whatever the initial trigger, once the appropriate conditions are met, salt begins to flow by evacuation from the original depositional position, or the autochthonous layer, causing subsidence in the overlying sediments. Sagging of the sediment alters accommodation, which in turn affects deposition centers and thicknesses; generating key diagnostic features in the minibasin. These features can be used in order to interpret the evolution of salt diapirs and the associated sediment. Two main features were observed in this study: contact relationships between the salt diapir and associated minibasin deposits, and isopach-thick stacking patterns.

Over long time spans, relative diapir rise rates and minibasin aggradation rates generate characteristic salt morphologies and contact relationships (Giles & Lawton, 2002). Giles and Lawton (2002) defined three idealistic scenarios that generate diagnostic shapes; however, one salt system may evolve through multiple iterations of all or a few of the situations. Figure 2.2
Figure 2.1: Physical scaled model conducted by Ge et al. (1997) showing salt structures evolving over normal basement faults by progradation. Figure taken from (Ge et al., 1997).
shows the geometries of salt and minibasin deposits in relation to these rates.

In Giles and Lawton’s (2002) first scenario, the salt rise rate is faster than aggradation of surrounding sediment. The diapir is unconstrained, and bulges outward, and if the rates are fast enough, salt will flow at the surface. The evacuating salt exacerbates subsidence and the sediment aggradation that cannot keep up with the extra accommodation, causing the sediments to flare upward next to the adjacent emplaced salt. The minibasin deposits can feature overturned beds and even repeated sections if the diapir rise rate is much faster than sediment deposition (Giles & Lawton, 2002).

In Giles and Lawton’s second scenario, salt rise rate is equal to the aggradation rate and sediment is able to fill the generated accommodation. Sediments in this scenario form up-lap contacts at the salt body, high angle unconformities, and vertical salt wall contacts (Giles & Lawton, 2002).

In the last model, diapir rise rates are lower than the depositional rates. Because the salt cannot keep up with deposition, sediments bury the salt. This leads to narrowing salt nearer the surface and sediments on-lap and eventually overlap the salt diapir (Giles & Lawton, 2002).

Minibasin isopach-thick stacking patterns can also be insightful as to how the salt walls and structures move throughout time. Kluth & DuChene (2009) describe how salt walls and the surrounding sediment evolve along a NE/SW transect through Pine Ridge. A modified image from their paper illustrates the evolution of sequential minibasins by progradation, shown in Figure 2.3. As overburden pressure increases, the weak, pressurized salt begins to nucleate on pre-salt structure and will eventually flow upwards. As salt is evacuated from one area, accommodation is created, forming a new depositional site. As the salt migrates, so does the basin center. New deposits will be thicker where accommodation is greatest; these are called isopach thicks. The isopach thicks will move toward the salt body that is migrating through time, and in this manner a basin history can be reconstructed (Kluth, and DuChene, 2009).
Figure 2.2: Geometries of salt and minibasin in relation to diapir rise rates and aggradation. a.) Salt response b.) Contact Geometries c.) Unconformity relationships d.) Depositional stacking patterns e.) Salt flow direction (Modified from Giles & Lawton 2002)
Figure 2.3: Geometries of minibasin depositional packages in relation to salt movement (Modified from Kluth & DuChene, 2009). Salt wings, or christmas tree structures can be seen where salt rise rate was faster than sedimentation rate. Erosion occurred across the basin, removing units 4 and 5.
2.2 Tectonic Evolution

The Paradox Basin is defined as a thick-skinned intraforeland flexural basin (Barbeau, 2003) whose extent is defined by the presence of Pennsylvanian to Permian evaporites. Although the unique diapiric and minibasin features generated by salt movement are Pennsylvanian and younger, it is the older tectonic evolution that controls the original salt location and thickness. This makes it essential to understand the tectonic history of the basin to comprehend the present day basin morphology. Figure 2.4 is a time-line, not to scale, showing the various tectonic events throughout the basin’s history. The unique features seen in the Paradox Basin are controlled by two major tectonic events. The first influential event was the emplacement of the basement fault framework. These basement faults controlled where salt later accumulated, and also became nucleation sites for salt movement. The second defining tectonic event was the generation of the Uncompahgre uplift, which caused flexure of the basin, creating additional accommodation for evaporite deposition. The Uncompahgre also supplied sediments that prograded into the basin and led to differential loading of the salt.

The earliest and most influential fault framework seen in the basin was emplaced during the Precambrian (Figure 2.4). These faults are believed to have formed at a subduction zone between the Archean craton and offshore magmatic arcs during collision of plate boundaries (Condon, 1995). This resulted in two perpendicular, deep seated regional basement lineaments (Baars, & Stevenson, 1981) shown in Figure 2.5. The northwest trending, right lateral Olympic-Wichita Lineaments developed sometime in the Paleoproterozoic to Mesoproterozoic (1700 to 1400 Ma) (Baars, 1966); and the northeast trending, left lateral Colorado Lineaments developed sometime in the Paleoproterozoic (about 1700 Ma) (Figure 2.5) (Warner, 1978). The two lineament systems intersect near Moab, UT (Stevenson, 1981). These basement features control the orientation of salt wall growth by acting as a conduit for movement in the case of the Olympic-Withinita Linaments, and as a barrier to flow in the case of the Colorado Lineaments. The close relationship between the salt wall locations and the basement faulting can be seen in Figure 2.6.
Figure 2.4: Tectonic timeline showing important events in the Paradox Basin History. Timeline is not to scale. Data were compiled from (Wengerd, 1962; Baars, 1966; Warner, 1978; Baars & Stevenson, 1981; Kluth & Coney, 1981; Lemke, 1985; Heyman et al., 1986; Stevenson & Baars, 1986; Baars & Stevenson, 1981; Condon, 1995; Barbeau, 2003; Banbury, 2005; Kluth & DuChene, 2009; Trudgill, 2011).
Figure 2.5: Regional lineaments which intersect over the study area. Basin extent is shown in blue and study areas shown in red (north--Salt Valley Salt Wall, south--Pine Ride diapir) (Modified from Baars & Stevenson, 1981).
Figure 2.6: Salt wall extent interpreted from gravity gradient map (after Banbury, 2005). La Sal and basement faults interpreted from well data, gravity and seismic (referenced from Trudgill, 2011; Case and Joesting, 1972; Frahme and Vaughn, 1993; Friendman et al., 1994; Ross 1998; Banbury, 2005; Kluth & DuChene, 2009; Utah Division of Oil, Gas and Mining well-data files, 2009).
These basement features were later reactivated in response to the Gondwana and Laurentia plates colliding along the Appalachian-Ouachita-Marathon fold-thrust belt (Kluth & Coney, 1981; Barbeau, 2003). This generated the Uncompahgre uplift during the formation of the Ancestral Rocky Mountains (ARM) (Barbeau, 2003, Trudgill, 2011). The uplift is believed to occur from the mid-Pennsylvanian Desmonian (~310 Ma) through early Permian Wolfcampian (~260 Ma) (Figure 2.4) (Wengerd, 1962; Lemke, 1985; Stevenson & Baars, 1986; Barbeau, 2003).

Reactivation of the basement faults continued throughout the basin’s history. Locally, these faults were reactivated in the southwest Paradox during the Cambrian, Devonian, and Mississippian, and erosion associated with the uplifts locally removed the Ouray and Leadville formations (Figure 2.4) (Baars & Stevenson, 1981). However, the sediments in the Pine Ridge diapir area do not appear to be affected.

Later Laramide shortening events heightened the basement lineaments (Baars & Stevenson, 1981), but they are interpreted to be minor, and have little effect on the resulting structural features within the basin (Heyman et al., 1986; Kluth & DuChene, 2009; Trudgill, 2011).

2.2.1 The Uncompahgre Uplift

The second influential tectonic event in the basin’s history was the formation of the Uncompahgre uplift. The Uncompahgre uplift is confined by two major faults that trend northwest to southeast and are about 50 km (about 164,000 feet) apart (Barbeau, 2003). However, the faulting style changes to the southeast, and becomes several faults. The main southwest bounding thrust fault dips about 50 degrees to the northeast; the minor thrust faults to the south also dip to the northeast but range from 35-50 degrees (Trudgill, 2011). The northeast fault is sub-vertical dipping southwest, the faults are assumed to connect in the subsurface (Barbeau, 2003). The Uncompahgre uplift has about 5 km (about 16,500 feet) of vertical movement and about 10 km (about 32,800 feet) of horizontal shortening and extends 200 to 300
km (about 656,000-984,200 feet) (Barbeau, 2003).

Trudgill (2011) argued that the Uncompahgre uplift did not simultaneously rise, but vacillated through periods of uplift along strike at various times. Figure 2.7 and Figure 2.8 are schematics completed by this study of the interpreted history of the Uncompahgre uplift, with Figure 2.7 describing the early stages of the uplift’s development, and Figure 2.8 describing the late stages of the uplift’s development. During the first stages of the Paradox Basin development, the Uncompahgre was not a structural high, and the Paradox and Eagle Valley Basins were inter-connected (Rasmussen, 2014; Kluth & DuChene, 2009; as referenced by Trudgill, 2011). Between the Uncompahgre Uplift and the Paradox Basin, left lateral transpressional movement was occurring running east-west (Thomas, 2007; as referenced by Trudgill, 2011). As the Uncompahgre developed, still in the early stages of the Paradox salt deposition, there was a pulse on what is now the northwestern extent of the fault. The uplift became a structural high, although barely above sea-level. This formed as a series of disassociated thrust faults (Arbuckle, 2008; Trudgill & Paz, 2009; as referenced by Trudgill, 2011). During late Paradox deposition, the central portion of the fault became much more prominent, and eroded Uncompahgre arkosic material was shed onto the salt below (Elston & Shoemaker, 1960; as referenced by Trudgill, 2011). During the last stages of the uplift, from the end of the Pennsylvanian to the Permian, the entire uplift, including the southeast portion had become a prominent feature (Moore et al., 2008; as referenced by Trudgill, 2011)

The early flexure in the north led to accommodation and accumulation of early salt cycles that are not seen in the south (Hite & Buckner, 1981) as shown in Figure 2.9.

2.2.2 Differential Loading

The additional load caused by the crustal thickening during the Uncompahgre uplift is the mechanism of flexural loading that leads to accommodation generation immediately adjacent to the uplift. Because of the additional accommodation, salt deposits are thickest near Uncompahgre uplift in the foredeep, and become thinner to the south away from the uplift in
Figure 2.7: Schematic of the evolution of the early Uncompahgre uplift. Data compiled from (Elston & Shoemaker, 1960; Elston et al., 1962; Barbeau, 2003; Paz, 2006; Rasmussen, 2006; Thomas, 2007; Arbuckle, 2008; Moore et al., 2008; Kluth & DuChene, 2009; Trudgill & Paz, 2009; as referenced by Trudgill, 2011).
Late Pennsylvanian (End of Paradox deposition)

End Pennsylvanian to Permian

LEGEND
Location of future salt walls in the Paradox Basin

S.V.: Salt Valley
F.V.: Fisher Valley
C.V.: Castle Valley
M.S.V.: Moab Spanish Valley
Sin.: Sinbad Valley
P.x.: Paradox Valley
P.R.: Pine Ridge
G.V.: Gypsum Valley
L.V.: Lisbon Valley

Eroded Uncompahgre material
Approximate Sea Level
Evaporation
Principal Stress Direction

Figure 2.8: Schematic created by this study, of the evolution of the Uncompahgre uplift from the Late Pennsylvanian to the Permian. Data compiled from (Elston & Shoemaker, 1960; Elston et al., 1962; Barbeau, 2003; Paz, 2006; Rasmussen, 2006; Thomas, 2007; Arbuckle, 2008; Moore et al., 2008; Kluth & DuChene, 2009; Trudgill & Paz, 2009; as referenced by Trudgill, 2011).
Older evaporite cycles exist in the north that do not exist in the South. This may be due to the Uncompaghre’s uplift history, with uplift beginning in the north, generating flexure and additional accommodation. (Modified from Hite & Buckner, 1981)
the forebulge and the backbulge (Herman & Barkell, 1957; Goldhammer et al., 1991; Barbeau, 2003; Whidden et al., 2014). Figure 2.10 is a schematic of depositional extent of evaporites and arkosic material shed from the Uncompahgre across the basin. Sediments eroding from the uplift prograded into the basin, and infilled the accommodation generated by flexure. As deposits thickened, the weak salt became differentially loaded and began to form salt walls and diapirs (Rasmussen, 2014). The initiation of salt movement occurred over basement faults which acted as nucleation sites for the subsequent salt wall structures (Kluth & DuChene, 2009).

2.3 Stratigraphic Background

The Paradox salt movement has varying degrees of influence on the depositional patterns of younger stratigraphy. These salt deposits range in thickness across the basin from about 10,000 to 15,000 feet, closer to the former at the Pine Ridge Diapir. Deposits of the Paradox salt and younger sediments lie on top of older stratigraphy that is not defined by the same depositional extent or tectonic forces (Condon, 1995). However, the older basin history and associated deposits influence the younger depositional packages, as well as the formation of salt wall location and are therefore described. The stratigraphic units, their inferred environmental setting and hydrocarbon potential can be seen in the Figure 2.11 (modified from Hintze & Kowallis, 2009; Trudgill, 2011; Whidden, 2014; and Rasmussen, 2014).

2.3.1 Precambrian

Precambrian basement rocks of the Paradox Basin are comprised of gneiss, schist, and intrusives, but mainly granite near the Pine Ridge area, and are about 1,800 to 1,740 Ma (Condon, 1995). This was followed by extensive erosion and potentially non-deposition (Condon, 1995).

2.3.2 Cambrian and Devonian

Deposits from the Cambrian to the Devonian suggest that the basin was near the Equator, and sediments were deposited in a warm-shallow marine to shelf environment (Condon, 1995). Formations deposited during this time include the Cambrian Ophir, Lynch, and the
Figure 2.10: Cross section of the asymmetric Paradox Basin from the shallow southwest to the deep northeast where the thick skinned Uncompahgre uplift resides (modified from Goldhammer et al., 1991; Barbeau, 2003; Whidden et al., 2012). Depositional extent of deposits modified from Herman & Barkell (1957).
Figure 2.11: Stratigraphy of the Pine Ridge and Salt valley Salt wall, depositional environment and hydrocarbon potential. (Modified from Hintze & Kowallis, 2009; Trudgill, 2011; Rassmussen, 2014; Whidden et al., 2014).
Devonian McCracken, Elbert and Ouray, which include sedimentary clastics and conglomerates, limestones, dolomites and small amounts of shale (Condon, 1995). A major unconformity between the late Cambrian and the late Devonian sediments represents a time of erosion and/or non-deposition, but it is believed that the craton was stable due to the current horizontal position of successive units. The end of the Devonian is punctuated by a mass extinction, and an associated sea level fall (Condon, 1995).

2.3.3 Mississippian

The overlying Mississippian age deposits are the fossil bearing Leadville limestone. The unit ranges in thickness from about 200 feet in the southeast to about 700 feet in the northwest (Morgan, 1993; Hintze & Kowallis, 2009; Chidsey, 2016). The Leadville displays very similar depositional environments to the underlying Devonian units: warm-shallow marine and near the equator. Transgressive and regressive cycles can be seen within the unit, caused by tectonic forces during the Antler orogeny as well as sea level fluctuations. By the end of the Mississippian, sea level dropped, and the basin became exposed to subaerial weathering and erosion for millions of years, forming the Molas Formation (Condon, 1995).

2.3.4 Pennsylvanian

During the early Pennsylvanian, subaerial conditions prevailed. The Paradox Basin was still near the equator, and intense weathering generated the red, clay rich regolith of the Molas formation as well as dissolution of the underlying Leadville which created karstic features (Condon, 1995). A transgression then changed the depositional style, and shallow marine conditions created the cyclic Pinkerton Trail limestones, and dark shales (Nuccio and Condon, 1996).

The middle Pennsylvanian to the late Pennsylvanian consists of cyclic marine and evaporitic deposits of the Paradox Formation, which define the basin extent. During this period, the basin was in a restricted marine environment because the Uncompahgre had become enough of a structural feature to isolate the basin from open marine conditions. The Pine Ridge salt body
deposits were formed during this semi-arid to extremely arid climate condition, resulting in thick accumulations of evaporites. Original salt deposition is believed to be up to between 7,000 and 8,000 feet thick (Hite et al., 1984).

The Paradox formation includes the Alkalai Gulch/Cane Creek, Barker Creek, Akah, and Desert Creek members (Figure 2.11). The Ismay evaporites also formed during the Pennsylvanian. These are cyclic deposits of dolostone, black shale, anhydrite, and halite deposits (Nuccio & Condon, 1996), which were deposited in a series of cycles. The cycles are mainly controlled by glacioeustatic sea level fluctuations and can be correlated to Milankovitch cycle timing (Hite & Buckner, 1981; Trudgill, 2011; Rasmussen, 2014). Sea level fluctuations closely mirror the evaporite deposits because of solubility of minerals and the salinity of the water. When sea level was at its highest, black shale was deposited. As increasing evaporation occurred and salinity increased, dolomite, anhydrite, and then halite were deposited. Likewise, when an influx of less saline water enters the system, the minerals deposit in reverse order (Hite, 1961). Many of the halite zones show disconformities, which is indicative of sea level rise and subsequent dissolution of halite. The Paradox formation has 29 defined cycles down to the decimeter scale (Rasmussen defined 83 cycles down to the millimeter scale), each varying in completeness (Hite, 1961).

Many of the shales and carbonate members within the formation are hydrocarbon producers, sources and/or seals. The Desert Creek carbonates and anhydrites act as a seal. The Cane Creek shale, and organic-rich shales in the Akah salts are sources. The Chimney Rock shale, Gothic Shale, and Hovenweep shale are both seals and sources. The Cane Creek shale, Barker Creek Member, Chimney Rock Shale, organic-rich shales in the Akah salts, the Desert Creek carbonates and anhydrites, Gothic Shale, Lower Ismay carbonates, Hovenweep shale and Upper Ismay carbonates are all producers (Whidden et al., 2014).

The Honaker Trail was deposited in the Upper Pennsylvanian during open marine conditions, and continued to experience glacioeustatic forcing cycles similar to the Paradox
Formation. The Honaker varies laterally because it is interpreted to form on the broad offshore shelf, and deposits are strongly controlled by sediment supply from proximity to terrestrial fans as well as water depth. The formation is also believed to represent paleotopography (Barbeau, 2003), and is composed of carbonate shoals, channels and fans (Trudgill, 2011; Rasmussen, 2014). The formation can vary from a dark gray to brown to red limestone or dolostone, and is locally fossiliferous (Trudgill, 2011; Rasmussen, 2014).

2.3.5 Permian

The time-transgressive Cutler Formation displays profound thickness changes across the basin in response to the rapid erosion and subsequent differential loading of the Uncompahgre Uplift (Figure 2.12). The lower Cutler group may have been deposited in the lower Pennsylvanian during the Desmoinesian (~315 Ma) (Barbeau, 2003), depending on the proximity to the Uncompahgre uplift, (Condon, 1997), with sediments closer to the uplift being coeval with salt deposition and as young as the Permain Wolfcampian (~262Ma) (Barbeau, 2003) toward the basin. The lower Cutler were deposited as alluvial fans and debris flows which generated from the Uncompahgre uplift and prograded into the basin (Figure 2.12) (Condon, 1997; Mack & Rasmussen, 1984; Barbeau, 2003).

During the Permian, the Elephant Canyon Formation was deposited locally in southeast Utah between Moab and Lisbon Valley and the confluence of the Green and Colorado rivers (Campbell, 1987). The Elephant Canyon is interpreted to have influences from eolian, fluvial and marine sources. The eolian deposits form to the west and the northwest, marine influences from the west and southwest, whereas the fluvial deposits come from the east. The environments are interpreted to coastal setting, shallow marine, and coastal plain respectively (Campbell, 1987).

After the lower Cutler and Elephant Canyon formations were deposited, during the early Permian Leonardian (Kungurian) age (~272.95+/−0.11Ma) (Cohen et al., 2013), there was a “regional” erosional event, the cause of which is still unknown (Rasmussen, 2014). This can
Figure 2.12: Images from Banham & Mountney, (2013) and Colorado Plateau Geosystems, (2013) shows the paleogeography and depositional environments at the time of deposition.
be seen as a prominent angular unconformity in the Pine Ridge northern minibasin, and across much of the basin.

The Organ Rock Formation was deposited in the Permian during subaerial and increasingly arid conditions. The environment is interpreted as being fluvial to eolian and is controlled through spatial distribution. Many plant and fossils are found within the fluvial section of the Organ Rock. Several unconformities are also found in this formation (Condon, 1997).

The White Rim Sandstone is an extensive eolian deposit with wind directions from the northwest to southeast that formed in the late Permian (Lawton et al., 2015).

### 2.3.6 Triassic

The Moenkopi Formation unconformable overlies the White Rim Sandstone, and was deposited after the regional Triassic unconformity (Rasmussen, 2014). The Moenkopi is interpreted to be deposited in a marginal marine environment, tidal flats and floodplains during transgression (Figure 2.12). The Moenkopi also appears to have sediments which are deposited in a more continental setting towards the Uncompahgre (such as fluvial deposits) and more marine environments to the southwest (Trudgill et al., 2004; Doelling, 1988).

The Chinle Formation was deposited in the Upper Triassic and unconformably overlies the Moenkopi formation. Interpreted to be deposited in a continental setting in a fluvial channel environment (Molenaar, 1981), both perennial and semi-arid (Doelling, 1988; Hazel, 1994), as well as a lacustrine environment (Figure 2.12) (Molenaar, 1981).

### 2.3.7 Jurassic

The Wingate sandstone formed in the Lower Jurassic, and is of eolian origin. It has similar wind directions as the White Rim sandstone, with dominant directions from the northwest to the southeast (Molenaar, 1981). The Wingate also contains deposits that are indicative of sand sheets, sabkhas and lacustrine deposits (Nation, 1998)
The Kayenta formed in the Lower Jurassic; it grades to a more fluvial dominated environment but still contains eolian deposits and even some conglomerates (Molenaar, 1981).

The Navajo is conformable with the Kayenta, and was deposited during arid conditions in an eolian environment (Molenaar, 1981). Again, dominant wind directions are from the northwest to the southeast (Molenaar, 1981).

The middle Jurassic Carmel is not present near the Pine Ridge salt diapir area, but can be seen in units to the north at the Salt Valley Salt wall; interpreted as depositing in an intertidal environment (Trudgill, 2011).

Entrada Sandstone formed during the middle Jurassic and contains deposits that are both aeolian dune, interdune (Doelling, 2000; Trudgill, 2011) and tidal flats in near-shore environments (Doelling, 2000).

The middle Jurassic Summerville Formation shows deposits that are consistent with tidal flat deposits on a marginal marine environment next to an interior seaway (Molenaar, 1981).

The Tidwell member formed during the middle Jurassic and is part of the Morrison formation.

The late Jurassic Salt Wash, is also part of the Morrison formation, and formed as a braidplain deposit as a distributive fluvial system (Parrish et al., 2004). This unit contains many terrestrial organics such as petrified wood. Provenance deduced from crossbedding dips suggests it is from the southwest (Moelenaar, 1981). The Salt Wash crops out in the Pine Ridge study area.

Brushy Basin Member formed during the upper Jurassic mostly in a broad floodplain environment and to a lesser extent, a lacustrine environment (Moelenaar, 1981; Doelling, 2000). The Brushy Basin member outcrops in the Pine Ridge Study area.
2.3.8 Cretaceous

The lower Cretaceous Burro Canyon has an environment similar to the Brushy Basin member in that there were fluvial channels and lacustrine deposits in a floodplain environment (Moelnaar, 1981). The Burro Canyon contains extensive fossils within the formation, and many terrestrial organics can be seen in outcrop near the northern Pine Ridge minibasin.

The Dakota Sandstone formed during the lower Cretaceous on a coastal-plain environment. Near the end of the Dakota deposition, the Mancos interior sea way began to form by regional sea-level rise (Molenaar, 1981).

The lower Cretaceous Tununk Shale Member is a unit within the basal Mancos shale. The Salt Valley salt wall project has state picked tops available for the Tununk Shale member. The Mancos was deposited in the foredeep in a shallow marine environment. Clastics from the Uncompahgre uplift can found within the unit (Chidsey et al., 2015).

Ferron Sandstone Member formed during the lower Cretaceous also as a lower Mancos shale unit, and is located in the northern part of the Paradox Basin. It is absent from the depositional record in the Pine Ridge area but is present in the Salt Valley salt wall with available state picked tops. This member is interpreted to be a shallow sea with very little sediment accumulation (Molenaar, 1975).

2.3.9 Tertiary

The Tertiary La Sal mountains are composed of three intrusive centers, with the southern most center emplacing into and to the north of the Pine Ridge Salt Diapir. Data suggest that the intrusives followed pre-existing weaknesses along faults and along salt cored anticlines. The intrusives were emplaced between 1.9 and 6.0 km deep (Ross, 1998).

Aeromagnetic data observes magnetic anomalies in the subsurface; rocks that contain strong enough magnetism to generate an anomaly include PreCambrian basement rocks and the La Sal intrusives (Steenland & Nelson, 1962). Likewise, gravity data observes gravity anomalies in the subsurface; salt can be seen due to low gravity anomalies in comparison to
surrounding sediments, whereas more dense intrusives and PreCambrian basement can also be seen (Steenland & Nelson, 1962).

Because the Pine Ridge Diapir is directly adjacent to the La Sal mountains, the gravity gradient map and the aeromagnetic map are affected by density and magnetic contrasts. The Pine Ridge diapir appears to curve, and trend east to west, however, the seismic data, and other salt wall trends suggest that this is an artifact of the intrusives.

2.4 Petroleum System

Oil and gas exploration in the Paradox Basin began in the 1900’s, however, there was very little success, and discoveries were inconsequential. That is until 1956 when Texaco had successes at the Aneth Field drilling carbonate buildups (Peterson, 1989). This southern Paradox field was developed by Texaco, Superior, Phillips, Shell, and Carter (Peterson, 1989). Around the same time the Big flat field, west of Moab, UT had moderate success in the Pennsylvanian and Mississippian reservoirs. Following these discoveries was the 1959 detection of Lisbon Valley field which had success in the Mississippian reservoirs (Peterson, 1989). Historically, most of the production of the basin has been on the conventional side (Intelligence Press, 2016) from carbonate buildups (Peterson, 1989). Continued exploration is still sporadic, and is strongly influenced by market pricing (Peterson, 1989).

In 2011, the USGS published an assessment of the undiscovered hydrocarbon in the Paradox Basin. The publication looked at elements needed for a hydrocarbon play including source generation, reservoir and seal distribution, and formation of traps. Calculations were made when all play elements were present, defined as an assessment unit (AU). Each AU was categorized based on geologic, source and hydrocarbon similarities. Using these criteria, seven AUs were defined and subsequently tallied. According to their calculations, the Paradox Basin has conventional resources between 25 to 176 million barrels of oil (MMBO), 234 and 1,591 billion cubic feet of gas (BCFG), and 4 to 37 million barrels of natural gas liquid (MMBNGL). The basin also has unconventional resources between 229 and 831 MMBO, 6,195 to 20,032
BCFG, and 221 to 854 MMBNGL (USGS, 2012).

However, all play elements do not exist over the entire basin. The spatial distribution of the conventional AU’s can be seen in Figure 2.13, and the unconventional AU’s can be seen in Figure 2.14. The Pine Ridge Diapir northern and southern minibasins are completely within three of the conventional AU’s including the (1.) Leadville McCracken AU (2.) Pennsylvanian Carbonate Buildups and Fractured Limestone AU and the (3.) Upper Paleozoic Mesozoic Reservoirs. The Pine Ridge northern minibasin is within two of the unconventional AU’s including (1.) Cane Creek Gas AU, and (2.) Gothic, Chimney Rock, Hovenweep Shale Gas. While the southern minibasin is within three unconventional AU’s (1.) Cane Creek Gas AU, and (2.) Cane Creek Shale Oil, although only a portion to the south, and (3.) Gothic, Chimney Rock, Hovenweep Shale Gas (USGS, 2012).

The Salt Valley diapir northern and southern minibasins are mostly within the conventional assessment units (1.) Leadville McCraken AU (2.) as well as the Pennsylvanian carbonate buildups and fractured limestone AU (1.) Cane Creek Shale gas AU, (2) the Gothic, Chimney Rock, Hovenweep Shale Gas AU, (3) and the Gothic, Chimney Rock, Hovenweep shale oil AU (USGS, 2012).

2.4.1 Thermal Maturity

As with the depositional distribution, the thermal maturities are not uniform across the basin. Instead, the maturation trends mimic burial history, and are modified by intrusives. Hence, the greatest maturities exist next to the Uncompahgre Uplift to the northeast, and decrease to the southwest and where sediment has not been buried as deeply. Likewise, formations within a single well have different burial histories dependant on where they reside in the section (Nuccio & Condon, 1996). Nuccio & Condon (1996) determined the thermal maturities for various points across the Paradox Basin for two intervals, the Cane Creek and the Desert Creek. Data used for their analysis included the Production Index (PI), Maximum Hydrocarbon Yield ($T_{\text{max}}$), and Vitrinite reflectance ($R_o$).
Figure 2.13: Conventional total petroleum system assessment units near the study area. (Modified from USGS, 2011)
Figure 2.14: Unconventional total petroleum system assessment units near the study area. (Modified from USGS, 2011)
The Production Index is a measure of thermal maturity, and the petroleum generation window (Nuccio & Condon, 1996) and is calculated using the following equation:

\[
\text{Production Index} = \frac{S1}{S1 + S2}.
\]

S1 and S2 are peaks calculated during rock evaluation. The S1 peak is a measure of free hydrocarbons while S2 is the quantity of cracked hydrocarbons, or the breaking of larger hydrocarbons into smaller hydrocarbons (Peters, 1986). Using this equation overall maturities can be determined because the PI goes up with increasing maturity. However, the equation assumes that there has been no hydrocarbon migration, and values were calculated with this assumption (Nuccio & Condon, 1996). The Desert Creek is most mature next to the Uncompahgre with maturities in the hydrocarbon window at 0.5, and becomes only marginally mature to the southwest. The Cane Creek values follow the same trend, but have higher PI values; they are overmature next to the Uncompahgre with PI values greater than 0.5 and in the hydrocarbon generation window with values greater than 0.1 to the southwest.

\[T_{\text{max}}\] values were also calculated, which also comes from Rock-Eval. The temperature value is recorded when the S2 peak is seen. Again, the S2 peak records when larger hydrocarbons start breaking down. As maturation increases, the temperature at which the S2 peak occurs also increases (Nuccio & Condon, 1996).

Finally vitrinite reflectance (\(R_0\)) values were calculated across the basin. This looks at the amount of light reflected off of a grain of woody organic matter. The amount of light reflected back is indicative of the thermal maturities of the basin, and change depending on the type of hydrocarbon being observed (Nuccio & Condon, 1996).

Nuccio & Condon (1996) used the three of these methods together in order to generate thermal maturity maps. This was done because samples collected did not display matching maturities for the various methods used (Nuccio & Condon, 1996). These results may not be consistent due to the migration of hydrocarbons during the basin history, giving one value for PI and another for \(T_{\text{max}}\) and \(R_0\) values. Finally, they generated burial history curves based on
stratigraphy and thermal maturity values. An assumption was made that the unconformities were non-depositional events, and did not result in any additional burial, and that the heat flow was constant through time (Nuccio & Condon, 1996). This will mean that most likely, hydrocarbon generation occurred sooner than these burial histories indicate (Figure 2.15, 2.16).

Burial history curves were generated for an area in Moab as well as an area in Lisbon Valley. The burial history curves can be seen in Figure 2.15 for Moab, UT and Figure 2.16 for Lisbon Valley. The Pine Ridge diapir resides between these two locations, however, it is closer to Lisbon Valley.

Although Lisbon Valley is only about ten miles away, the burial history curve does not show any additional heating due to the La Sals (Nuccio & Condon, 1996). This shows just how localized the intrusives influence on thermal maturity is. However, the southwestern La Sal Intrusive was emplaced directly into and adjacent to the Pine Ridge diapir (Ross, 1998). This will undoubtedly have increased thermal maturity during the Tertiary, yet, quantifying the additional maturation is difficult to discern. Studies have been completed to model the influence of intrusives on maturation values such as Peace et al. (2017), Wang et al. (1989), Alalade et al. (2013) as well as many others. But to accurately understand intrusive’s influence both on temperature and distance, these models need specific inputs such as diffusion coefficients, specific heat capacity of the intrusion as well as the host and conductivity (Peace et. al, 2017). These values are not publicly available for the Paradox Basin, and therefore, it can only be said qualitatively that the thermal maturation will be greater in the Pine Ridge due to the La Sal intrusives.
Early generation of hydrocarbons 10%-25%

Main phase generation of hydrocarbons 25%-90%

Late generation of hydrocarbons 65%-90%

Figure 2.15: Burial Thermal history map near Moab from Nuccio and Condon, (1996).
Figure 2.16: Burial Thermal history map near Lisbon Valley from Nuccio and Condon, (1996).
CHAPTER 3
METHODOLOGY

Several workflows were performed on the available data sets with the aim to accurately model and understand the evolution of the Pine Ridge salt withdrawal minibasins temporally and spatially. Workflows include field data collection, well log interpretation, seismic interpretation including horizon and fault picks, cross section generation, and structural restorations. However, the validity of interpretations must be scrutinized by examining the assumptions and the uncertainties associated with interpretation. This chapter discusses the workflows used and the associated assumptions made in the research.

3.1 Field Data

Field data collected along strike from the Pine Ridge Diapir constrains the timing of salt movement in the area. Salt withdrawal will generate accommodation, and minibasin deposits will respond. Thickness changes within the field would suggest that salt withdrawal was still occurring during the time of deposition, whereas uniform thicknesses would suggest a lack of salt withdrawal. Additionally, an active diapir could lead to increasing minibasin deposit dips closer to the diapir. These data were collected using the FieldMove Clino app as well as a Brunton compass. Unfortunately, the iphone 5C gyroscope was not gathering accurate compass direction readings, but was accurate with the GPS measurements. Therefore strike and dip were obtained using a Brunton compass while the FieldMove Clino app was used for location information. Extensive weathering made it difficult to distinguish bedding surfaces and get an accurate reading on strike and dip measurements, especially in the Brushy Basin and Dakota formations. However, the Burro Canyon formation was more resistant than surrounding formations and bedding planes could readily be measured.

3.2 Well Log Interpretation

Well logs were used in order to establish formation depth profiles as well as to calibrate the seismic data set. Well logs were downloaded from both Colorado and Utah state websites.
Colorado wells are in the coordinate reference system NAD83 UTM zone 13 meters. The Utah wells are in the coordinate reference system NAD83 UTM zone 12 meters. Wells were loaded in Global Mapper and converted to NAD27 Utah State Plane Southern Zone (4303), US foot and imported into the Petrel project. Discrepancies are present between the aerial photography and wellhead locations provided by the state which is likely due to the coordinate reference system differences. To account for this, wellhead latitude and longitudes were moved up to about 200 feet to match the photographed borehole where it was visible on the air photography. However, not all boreholes can be seen on the aerial images. Here an assumption was made that shifts to some of the boreholes and not others is not significant enough to cause concern, and still depicts a truthful image of the subsurface. Coordinate reference system conversions in Global Mapper were also completed for all cultural data from the state websites. Differences in datums generate inherent inaccuracies within the data set.

Well tops were gathered from the state websites, and imported into the project. Formation tops did not always match from well to well, and therefore formation tops were chosen on wells where the log response matched the published literature lithology description of the formation. These log responses were then used as a type log for other well horizon picks to re-correlate tops.

All wells were assumed to be vertical except the Pine Ridge Location #1 (API 4303731890). The Pine Ridge Location #1 was used to tie the seismic time volume to the well depth profile, and therefore the most accurate location was desired, and the directional survey was entered. The other vertical wells were assumed to give an accurate depiction of the subsurface without the directional surveys. This will generate slight changes to the depth profiles of the other wells, however, these changes are believed to be minor, as the wells in this area are vertical. The well’s kelly bushing, ground level, derrick floor and total depths were checked with the state websites. However, inaccuracies may arise when original measurements were mis-recorded.
3.3 Depth Conversions

It is important to understand how features exist in depth, as time volumes may generate misleading information about thickness values. Since the seismic volume was acquired in time, a depth conversion must be performed on the data. Several steps were undertaken in order to generate a velocity model which most accurately represents the data. Depth conversions were applied solely for restoration purposes, while seismic package and fault interpretations were completed in time.

Because the well logs are in depth and the seismic is in time, it is important to tie the two together so that they can be viewed simultaneously. The first step in tying the data sets involved is to generate a synthetic seismogram. A synthetic seismogram is the idealistic signal that would be seen from a seismic response at a reflection coefficient boundary, and is the sum of cosine waves with amplitudes and phases (Ewing, 2001). This idealistic generated wave pattern is then cross-correlated to the actual seismic data in order to place the well in time. Synthetics generated for the Pine Ridge Location #1 used the edited velocity curves as well as the seismic with a cross line window of 10. Reflection Coefficients used the despiked sonic values from the Pine Ridge Location #1, and edited density values from the TXC-Huber Federal 1-15. Because there are only two wells within the 3D seismic volume, and both the sonic and density logs were not available for either well, assumptions were made in order to depth convert. The first assumption being that the Pine Ridge Location #1 is an accurate representation of the subsurface across the minibasin. The second assumption is that the density log suite acquired in the TXC-Huber Federal 1-15 is representative of the density changes seen in the Pine Ridge Location #1 well. However, because formations occurred at different depth profiles within the two wells, the TXC-Huber Federal 1-15 density values were modified in depth in order to match the sonic curve well response in the Pine Ridge Location #1, generating a synthetic density log. Once this was completed, the well could be correlated to the seismic. Assumptions were made in order to tie seismic reflections to the formations seen in well logs. This was based on a well log depth profile in addition to approximate and idealized velocity values of the well log formations. By
generating the synthetic, the well and the seismic are tied by effectively converting the well to
time (See Figure 3.1).

Once the well and the seismic were in the same units, a velocity curve was generated.
The Pine Ridge Location #1 was used to generate the curve due to the available sonic curve,
delta t (DT) in addition to the well’s location within the seismic cube. The DT curve measures
slowness, and is the reciprocal of velocity (Asquith et al., 2004):

$$DT = \frac{1}{V} = \frac{\text{time (µsec)}}{\text{distance (ft)}}$$

This tool measures the sum of several waves within the borehole including the acoustic,
shear and tube wave. Because the tool measures wave energy, if there are poor hole conditions,
or there is gas present, then the energy of the wave will decrease and the tool will measure
additional time. This will result in an inaccurate slowness measurement and is called “cycle
skipping (Asquith et al., 2004).” The Pine Ridge Location #1 well has a P-wave measurement
which describes the resistance to shear and the incompressibility of rock when subject to
seismic waves. The P-wave measurement is also known as the compressional wave or push
and pull wave, and is the first to arrive (Asquith et al., 2004). This curve was used to calculate
the velocities of the subsurface (Figure 3.1). First, the curve was de-spiked to eliminate any
un-geologic jumps that are most likely due to poor borehole conditions, and would generate
erroneous velocity calculations. The borehole conditions (Figure A.2) are shown in the appendix
as a black curve, and was used in determining which spikes were removed. Three standard
deviations within a spike window were used in order to clean up the curve (Figure 3.1). Then the
transverse time was calculated for the well.

$$V_{dt} = \frac{(1,000,000 \text{ms})}{(\text{DT}_{\text{despiked}})}.$$ 

The velocity curve was then compared to horizon picks (Figure 3.1). Significant velocity
changes occur when a new formation is encountered. Normally this sonic velocity curve would
be calibrated against a checkshot or other type of velocity survey. Comparing the sonic log to
the checkshot can help detect errors in the log and or the checkshot, or other velocity surveys.
Figure 3.1: Synthetic generation used in calculating velocity data later used in Move. A.) Synthetic Generation created in Petrel B.) P-Wave velocity vs MD, velocities calculated by formation is m/s and ft/s. C.) Despiked DT curve D.) Formation boundaries show a drastic change velocity.
However, a check shot was not available for the well, and therefore was not completed. Because of this, potential errors in the data are not caught, and this must be noted during interpretation, and may cause either a shallower or deeper depth than is actually present. The velocity curve was then applied to the rest of the cube. Once again, an assumption was made that the velocity curve generated would match the rest of the seismic cube where data were not present.

A second method was also used in order to depth convert for comparison purposes, and was ultimately adopted because it honored the data more accurately than the velocity curves. This was completed in Move 2016.2. Time horizons, interpreted in Petrel, were brought into Move, in addition to the depth profile wells. A depth conversion was completed within the Move 2016.2 software package using a fixed model, which states that the velocity increases with depth (Move Tutorial, 2016), defined by the function:

\[ Z = V_0 \left( e^{kt} - 1 \right) / k \]

Where \( Z \) is the depth of the formation, \( V_0 \) is the initial velocity, \( k \) is the rate of velocity increase with depth, \( t \) is the one way travel time, and \( e \) is the natural logarithm base (Move Tutorial, 2016). These parameters were input from the values obtained from the velocity curve generated in Petrel (Figure 3.1). Once this was completed, minor adjustments were made to the horizons by manually bulk shifting the horizons to the depth profile seen from the well logs.

### 3.4 Seismic Interpretation

Seismic interpretations were completed in order to understand minibasin stacking geometries, and the relationship of contacts, in order to derive the evolution of salt movement and minibasin development. The seismic data record indirect geophysical measurements which are used to interpret geological information (Herron, 2011). Assumptions must be made about the physical properties of the rock and what geophysical signature they will exhibit. The 3D seismic cube located over the Pine Ridge minibasin was donated, and did not contain acquisition or processing information. Therefore, it is unknown how much signal to noise was filtered out, what the resolution of the cube is, and what focusing techniques were used. Processing seismic
involves assumptions including subsurface velocities. If inaccurate assumptions are made, velocity pull ups and/or velocity pull downs can be generated. This can lead to the creation of structures which do not really exist in the subsurface, such as an anticlines or synclines (Herron, 2011). Accurately imaging formations under salt bodies is difficult because the thickness distribution of salt, as well as velocities must be known. Due to the complex nature of salt this can be difficult to fully understand. For example, salt can have a wide variety of lithologies and velocities within the salt body which can be highly distorted as salt flows. In addition, because salt flows, there can be an irregular distribution of salt (Herron, 2011). Although it can be difficult to image subsurface salt, by using a combination of a seismic volume, Bouger Gravity maps, and well data, an insight into large scale features within the sub-salt section can be seen.

Other data are also unknown, including the polarity of the data. The polarity of the data can sometimes be determined by looking at the sea-floor to understand the seismic response. However, this is not possible in the land-locked data set. Instead the top and base of the salt were observed and the polarity of the data determined this way. The top of the salt shows a positive reflection as red, and the base of the salt as a negative reflection as blue, and therefore is in American polarity.

The donated 3-D cube has both a PreStack 3-D volume as well as a PostStack 3-D volume. The PostStack volume was clipped in order to visualize smaller changes in amplitude. 0.07252% of the outlier data were eliminated, and these were most likely false data. This was done because there are only so many color range values, and they must represent the range of values seen. Therefore, the data are binned, and color bars are split between the amplitude ranges. The original, unclipped data had a range of minimum amplitude -63,224.78 to a maximum amplitude of 64,769.42. This means that there are 127,994 unique data values that were then grouped into 4,096 bins, which is the pre-set value in Petrel. This means that amplitude ranges were binned every 31 values. However, once the outliers were eliminated, the minimum amplitude was -20,000 and the maximum amplitude was 20,000.00. The data values clipped are 0.03643% for the minimum amplitude and 0.03609% of the maximum amplitude.
were clipped. Taking this value range, and having a total of 4,096 bins, this means that binning was completed on a value range closer to 9.8 for every color. This allows for greater detail, and more subtle changes to be seen within the data. The data was clipped using a filter, zero centric, floating point 32 bit (so as not to lose information), and 4,096 bins. The seismic data were then colored for optimal viewing including the seismic default color bar, limits defined in the color table, RGB -40,000 to 40,000, and a non-linear gradient.

A series of standard seismic volume attributes were run in order to better visualize the data including structural smoothing, relative acoustic impedance, instantaneous frequency, envelope, simple sweetness, and cosine of phase. Each of these volume attributes helped to better understand the subsurface.

*Structural smoothing* was completed on the clipped volume Tricon Post Stack Migration with Petrel’s Dip-Guided edge enhancement. This increases the continuity of the seismic reflections (Pepper & Bejarano, 2005), and eliminates noise. The smoothing is completed by using dip and azimuth to determine the structure than a Gaussian smoothing is run on the structure, while the edge enhancement detects the edges.

The *Relative Acoustic Impedance* is completed on the clipped and smoothed 2D volume. This is a sum of amplitude values which can be used to show apparent acoustic contrast-accentuating boundaries, and discontinuities.

*Instantaneous Frequency* was run on the clipped, smoothed and relative acoustic impedance volume. This was done on a window size of 33, which is the automatic setting and worked well for the dataset. The data were then clipped from -68.8340 to 113.7199 to -20 to 80. This removes any outliers which skew the data. The attribute is used in correlation of horizons and can especially be useful across faults, as well as determining the location of salt bodies.

The *Envelope* was run on the *Instantaneous Amplitude* volume and measures the magnitude of reflection. Since this attribute is only looking at the magnitude, it is independent of phase. This can be used to detect any changes in lithology and can also be used to detect gas.
The automatic settings were used to run this attribute, and the window was set to 33, and the volume was realized to a 32 bit volume in order to preserve data.

The *Simple Sweetness* can detect energy signature energy changes and can be especially useful in detecting channel bodies (Hart, 2008). The data were clipped from 0-10,000, which clips the upper 0.203% of the data. The simple sweetness attribute is generated using the equation:

$$\text{Simple Sweetness} = \frac{\text{Instantaneous Amplitude}}{\text{SQRT (Instantaneous Frequency)}}.$$  

The *Cosine of Phase* is also known as the cosine of the instantaneous phase or the normalized amplitude. This can be used to enhance features, and is especially useful in detecting salt bodies. With the cosine of phase, salt bodies are delineated by the discontinuous nature of the signal.

In using different seismic attributes, various features became highlighted, including the location of the salt body, faults and horizon continuity.

### 3.5 Cross Section Construction

Cross section construction was completed using the 3D seismic cube and well log data in the northern Pine Ridge minibasin. Cross section construction for the southern minibasin utilized well log data, field data, and projected interpretations from Kluth & DuChene (2009) 2-D seismic line drawings. 3D seismic and well data were interpreted in Petrel and transferred to Midland Valley’s Move 2016.2. 2D interpretations were georeferenced into Midland Valley’s Move 2016.2, and subsequently projected onto the cross section line. Field data were collected and transferred into the Move project via Midland Valley’s field move clino app for the iphone supplemented by Brunton measurements.

### 3.6 Restorations

Restorations were completed in order to validate interpretations, as well as understand the evolution of the Pine Ridge minibasins through time. Midland Valley’s Move tutorial in addition
to Rowan’s (1993) paper on restoration were used to determine the best workflow for restorations and backstripping. Backstripping is the process of removing any influence which has shaped the present day subsurface configuration starting with the most recent process and working methodically backward through time (Rowan, 1993). This includes the processes of mechanical compaction, isostatic adjustment, and diapirism (Rowan, 1993).

Many assumptions are made in order to generate the restoration models. Restorations were completed on 2D cross sections which run perpendicular to the Pine Ridge diapir axis. An assumption was made that the salt stays within the 2D cross-section, window, and does not move in and out of the plane (Rowan, 1993). This is most likely an incorrect assumption, but one that must be made to simplify modelling. A second assumption is that all of the diapirism present can be seen on the cross section. Again, this is an invalid interpretation, as the actual amount of salt is unknown, and can only be constrained through seismic and well log data, and is sparse in many areas. Since the seismic dataset does not cross into the southern Pine Ridge minibasin, volumes and placement of salt is estimated using 2D published line drawings from Kluth & DuChene (2009). Another important assumption to note is that unconformities are assumed to remove little to no material. Although the contact relationships at the unconformity contacts suggest that material was removed, it is difficult to determine just how much material was removed. Another factor not accounted for in the reconstruction model includes any chemical compaction caused by smectite and kaolinite changing to illite (Move, 2016). Greater error in decompaction values are generated in formations with more smectite and illite. Units with the greatest chemical decompaction error are the Morrison and Mancos formations, and have the least error in the White Rim sandstone.

Units are decompacted in order to understand original thicknesses before compaction occurred and see how rock volumes and porosity loss have changed in response to burial (Move Tutorial, 2016). Rock properties were determined for each horizon based on Nuccio and Condon’s (1996) burial and thermal history research near Lisbon Valley, UT and Moab, UT. An average between the two locations were used in order to determine an estimated sandstone, shale
and limestone percentage for the Pine Ridge area, which lies in between these two locations. The differences in lithology between formation changes the amount of mechanical compaction that occurred.

Move has three built in equations for porosity changes due to compaction. The first is the Sclater & Christie (1980) equation which can be described by the equation: $f = f_0(e^{-cy})$, where $f$ is the current porosity at depth, $f_0$ is the porosity at the surface, $c$ is a porosity-depth coefficient ($\text{km}^{-1}$), and $y$ is depth. The surface porosity is determined using Sclater & Christie’s (1980) calculations from North Sea lithologies which were under normal pressure conditions (Move Tutorial, 2016). Since the lithologies in Pine Ridge have mixed lithologies, the surface porosity of lithology components are averaged. The porosity-depth coefficient is also calculated using the North Sea values determined by Sclater and Christie (1980), and where mixed lithologies are involved, this depth coefficient is also averaged. Since salt is incompressible, the decompaction is set to zero. Another equation used by Move 2016 in determining the effect of compaction is the Baldwin & Butler (1985) which uses the equation: Burial depth (km) = $6.02xS^{6.35}$ Where $S$ is solidity which is similar to porosity. However, this equation is used in shales and limestones, and does not apply to sandstones (Move Tutorial, 2016). Since much of the Pine Ridge area is composed of sandstone, this equation was not used. Finally, the last equation used for decompaction comes from Dickinson (1953) which was calculated using shales in the Gulf of Mexico, and is defined by the equation: Burial depth (km) = $15xS^8$. Again, where $S$ is solidity. This equation best describes overpressured areas. The Sclater & Christie method was ultimately used because of its ability to represent mixed and sandstone lithologies more accurately. Decompaction values ranged from 7.8% in more shallow formations to 52.5% in the deepest Honaker Trail.

Once units are removed, and subsurface units are decompacted, the effects of isostatic adjustment also need to be removed (Rowan, 1993). Here the Airy isostasy equation was used because it is one dimensional (Rowan, 1993), meaning that the isostasy is only compensating above and below the additional load, and not laterally (Move Tutorial, 2016). This one
dimensional model is sufficient to explain the isostasy occurring in the Pine Ridge area where much of the load is compensated for by the underlying Paradox Salt due to its ductile nature (Rowan, 1993). The Airy isostasy calculation can be described by the equation:

\[ Z = S - (H_1 - H_2) \frac{P_c - P_w}{P_m - P_w} \]

“Where \( Z \) is the amount of subsidence, \( S \) is the loaded or unloaded sediment thickness, \( H_1 \) is the crustal thickness before the load, \( H_2 \) is the crustal thickness after the sediment load, \( P_c \) is the crust density, \( P_w \) is the water density, \( P_m \) is the mantle density (Move Tutorial, 2016, airy isostatic compensation).”

Once decompaction and isostatic adjustment had been accounted for, the underlying formation was then restored to the regional datum. This was done using the 2D unfolding with the Flexural Slip algorithm. The flexural slip algorithm allows for the formation to be restored to a target (Move Tutorial, 2016). If the paleobathymetry is known, then the formation will be restored to that horizon. However, for this dataset, the paloetopography was unknown, therefore, a regional datum was used. The regional datum was determined for each individual formation by selecting areas which did not show depositional growth, suggesting that the units were deposited on a horizontal plane. The underlying units were also passively unfolded. This was systematically completed for every formation.
4.1 Stratigraphic Architecture from Seismic Interpretation

There is a elevational tilt that can be seen in the 3D cube, with the north consistently deeper than the south. This change in elevation is more prominent in the deeper horizons, and especially in the sub-salt horizons.

Fifteen packages were interpreted across the 3D seismic volume based on the unique seismic signature of the reflection patterns, contacts, reflection continuities, and reflection amplitudes (Table 1, Table 2, Table 3). These packages were then assigned standard Paradox Basin formation names based on rock property response predictions and depth callibrations from well data. The predicted rock responses are based on expected changes in lithology between formations, since the seismic response is based on the sum of changes in acoustic impedance and reflection coefficients. When acoustic impedance increases and reflection coefficients are positive, than a seismic peak is generated. In this same way, if the acoustic impedance decreases and the reflection coefficient is negative, than a trough is generated (Herron, 2011). In general, shales (1,790-5,805m/s) have a slower velocity value than sandstones (5,490-5,950m/s) (Glover, 2016). Although a few reflections can be seen below the salt, only one seismic package was interpreted in this area due to the lack of well control and the discontinuous nature of the reflections across most of the cube. These fifteen seismic packages are assigned in Table 4.1, Table 4.2 and Table 4.3. The seismic packages can be seen on the composite line in Figure 4.1.

4.1.1 Salt Geometries

The named horizons were picked across the 3D seismic cube using a combination of manual picking, 2-D auto-tracking and 3D auto tracking tools. Several attribute volumes were also used to determine the most accurate contact between the Paradox salt and the minibasin sequences. Attribute volumes used were relative acoustic impedance, instantaneous frequency, envelope and cosine of phase. These attributes were viewed both individually in addition to
Table 4.1: Table of seismic packages picked and assigned formations. Packages are determined using reflection patterns, continuity, and relationship of contacts.

<table>
<thead>
<tr>
<th>Seismic Package</th>
<th>Reflection Patterns and Continuity</th>
<th>Reflection Amplitudes</th>
<th>Contacts</th>
<th>Peak or Trough</th>
<th>Comments</th>
<th>Assigned Formation</th>
<th>Age Range (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Parallel Continuous</td>
<td>High</td>
<td>Continuous with younger unit, except where faulting has juxtaposed the unit 1 with unit 2.</td>
<td>Peak</td>
<td></td>
<td>Leadville Limestone</td>
<td>~323.2+/-.0.4 to ~358.9+/-.0.4 (Cohen, 2013)</td>
</tr>
<tr>
<td>2</td>
<td>Parallel Continuous</td>
<td>High</td>
<td>Continuous with older unit and unknown relationship with overlying unit, due to salt movement. Overlies seismic package 1 across most of the cube, except where faulting has juxtaposed seismic package 1 and 2/3.</td>
<td>Trough</td>
<td></td>
<td>Base Paradox Salt</td>
<td><del>298.9+/-.0.15 to</del>323.2+/-.0.4 (Cohen, 2013)</td>
</tr>
<tr>
<td>3</td>
<td>Random Discontinuous</td>
<td>Low</td>
<td>Juxtaposes seismic packages 4 through 15</td>
<td>Peak</td>
<td>Interpreted using the multi-z function within Petrel 2016.2. This unit had extremely variable thickness. Further refinement was interpreted using the cosine of phase seismic attribute.</td>
<td>Paradox Salt</td>
<td><del>298.9+/-.0.15 to</del>323.2+/-.0.4 (Cohen, 2013)</td>
</tr>
<tr>
<td>4</td>
<td>Parallel Continuous</td>
<td>Lower than seismic package 11</td>
<td>Truncates against seismic package 6 (erosion surface) and is juxtaposed next to seismic package 3</td>
<td>Peak</td>
<td></td>
<td>Honaker Trail</td>
<td><del>298.9+/-.0.15 to</del>307.0+/-.0.1 (Cohen, 2013)</td>
</tr>
<tr>
<td>5</td>
<td>Continuous</td>
<td>High amplitude to the north lower amplitude to the south near the truncation.</td>
<td>Truncates against seismic package 6 (erosion surface)</td>
<td>Trough</td>
<td></td>
<td>Elephant Canyon</td>
<td>~272.95+/-.0.11 (Cohen, 2013)</td>
</tr>
</tbody>
</table>
Table 4.2: Continued table of seismic packages picked and assigned formations. Packages are determined using reflection patterns, continuity, and relationship of contacts.

<table>
<thead>
<tr>
<th>Package</th>
<th>Formation</th>
<th>Description</th>
<th>Contact</th>
<th>Age (Cohen, 2013)</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>Continuous</td>
<td>Higher amplitude toward the south, lower amplitude toward the north</td>
<td>Defined by a high angle unconformity, truncating the reflections below.</td>
<td>Peak</td>
</tr>
<tr>
<td>7</td>
<td>Parallel Continuous</td>
<td>High</td>
<td>Reflections from seismic package 8 onlap onto seismic package 7 reflections</td>
<td>Peak</td>
</tr>
<tr>
<td>8</td>
<td>Parallel Continuous</td>
<td>High</td>
<td>Reflections onlap onto seismic package 7</td>
<td>Trough</td>
</tr>
<tr>
<td>9</td>
<td>Parallel Continuous to the south, more discontinuous to the north</td>
<td>High</td>
<td>Truncate against seismic package 10 (erosion surface)</td>
<td>Peak</td>
</tr>
<tr>
<td>10</td>
<td>Parallel continuous</td>
<td>High</td>
<td>Low angle truncation of seismic package 9</td>
<td>Follows the same dip trend as seismic packages 11 and 12</td>
</tr>
<tr>
<td>11</td>
<td>Lower than seismic package 12</td>
<td></td>
<td>Follows the same dip trend as seismic package 12.</td>
<td>Peak</td>
</tr>
<tr>
<td>12</td>
<td>Continuous</td>
<td>Higher amplitude than seismic packages 13, 14, and 15.</td>
<td>The dip trend is steeper than seismic packages 13, 14 and 15.</td>
<td>Peak</td>
</tr>
</tbody>
</table>
Table 4.3: Continued table of seismic packages picked and assigned formations. Packages are determined using reflection patterns, continuity, and relationship of contacts.

<table>
<thead>
<tr>
<th>Package</th>
<th>Description</th>
<th>Characteristics</th>
<th>Horizon</th>
<th>Formation</th>
<th>Age Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>13</td>
<td>Parallel semi-continuous</td>
<td>Low</td>
<td>This seismic package shows significant thickening toward the diapir and thinning to the north.</td>
<td>Peak-dominant reflection</td>
<td>Following the horizon across the survey is a difficult task due to the data quality near the surface. Although the seismic packages of 1, 2 and 3 can be described in the same manner, individual packages were distinguished using the more prominent reflection at the top of the seismic package, where the body of the seismic package became more discontinuous and difficult to distinguish reflections.</td>
</tr>
<tr>
<td>14</td>
<td>Parallel semi-continuous</td>
<td>Low</td>
<td>This seismic package follows the same dip trend as seismic package 13 and 15.</td>
<td>Trough</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Parallel semi-continuous</td>
<td>Low</td>
<td>Due to the poor data quality near seismic package 3 (salt) it is unclear whether this package was deposited on top of seismic package 3 (salt) or if it is truncated. However, based on outcrop at the surface, restorations, and the cosine of phase attribute, it is interpreted that the reflections truncate at the seismic package 3 (salt).</td>
<td>Peak</td>
<td>Because the seismic package assignment is based on seismic reflection and a basic depth profile, the actual rock member is unknown, and therefore cannot be given a more precise age range.</td>
</tr>
</tbody>
</table>
Figure 4.1: Composite section of Inline 1214 and Crossline 1229 smoothed volume in order to show the fifteen interpreted horizons. The display is shown with the seismic colorbar with red displaying peaks and blue displaying troughs.
being viewed as an overlay on the smoothed volume, or as an overlay on top of each other. The repetitive nature of observing a time slice with many different volume attributes allowed for the most consistent and accurate boundary of the salt body. Figure 4.1 is a composite section that shows a representative sample of what the salt diapir and minibasin geometries look like. This composite section is located in the northern minibasin, the seismic volume does not cover the southern minibasin. It shows that the autochthonous salt welds out to the north, and has a bulge located after the erosion surface of horizon 6.

A 3D surface model of the Paradox top salt is shown in Figure 4.2, generated using the 3-D seismic cube. The figure displays the salt geometry using a tessellated surface converted to point data and colored by elevation (in ms). The highest elevation is colored in warm colors and the lower elevation in cool colors. This helps draw the eye to specific elevations, and enhances features that would otherwise be missed. Small pulses within the evolving salt can be seen, which are indicative of minor shifts in salt rise rate both temporally and spatially as well as deposition center changes. A map view of the diapir is also displayed with points data and colored on elevation. The red lines represent the composite section used in Figure 4.1. The map view shows how the present day salt changes in elevation across the northern minibasin (Figure 4.2).

The southwestern edge of the 3-D seismic cube shows the highest diapir elevation. The outcrop data shows that salt is not present at the surface, so it is assumed that this is the peak elevation of the Pine Ridge diapir. From this peak, there is a rapid decrease in both elevation and salt isopach thickness into the minibasin. The salt quickly tapers off to the north, and although elevation still decreases quickly to the east, it is more gradual than the north. The Pine Ridge diapir is an elongate body that has a northwest to southeast trend. This is consistent with other diapir trends throughout the basin. Data documenting salt geometries in the Pine Ridge southern minibasin are restricted to a few wells in addition to Kluth & DuChene’s 2009 interpreted cross section based on 2-D seismic lines.
Figure 4.2: Figures are colored on elevation (ms) with shallower elevations in warm colors. A.) Pine Ridge diapir tessellated surface converted to point data. B.) map view of the northern minibasin diapir C.) Intersection of the tessellated diapir converted to point data with the composite section from Figure 4.1.
4.1.2  Fault Framework

Basement faults in the Pine Ridge minibasin study area have been interpreted as both normal faults (Kluth & DuChene, 2009), and reverse faults (Rasmussen, 2014). Figure 4.3 compares these basement fault interpretations as well as the fault interpretation completed for this study. Kluth & DuChene (2009) interpreted a normal fault on the southwestern side of the Pine Ridge diapir, whereas the northern end of the Pine Ridge diapir is unaffected by faulting. An additional basement fault is interpreted beneath the Paradox Valley salt wall in the same relative position as the Pine Ridge diapir. Rasmussen (2014) interpreted no offset directly under the Pine Ridge diapir, however, he interprets reverse faulting both in the northern and southern Pine Ridge minibasins. Rasmussen also interpreted both reverse and normal faulting beneath the Paradox Valley.

Interpretations for this study were completed on a coherence attribute volume over the 3-D cube for the northeast portion of the diapir. Faults on the southwestern end of the diapir were schematically drawn from observing 2-D seismic lines. The basement fault interpretations to the southwest are consistent with Kluth & DuChene’s (2009) interpretations (Figure 4.3). Faults were interpreted to be normal, striking northwest to southeast below the northeast portion of the diapir. Kluth & DuChene’s interpreted basement fault was adopted for the southwest portion of the diapir. Figure 4.4 shows the interpreted basement normal faults across the 3-D seismic cube. Figure 4.4 A is a structurally smoothed composite section of Inline 1227 and crossline 1064 with a variance overlay in order to enhance changes in coherency and hence faulting. Figure 4.4 B shows a map view of the faulting across the seismic cube. The faults have a northwest to southeast trend which the salt diapir outline mimics. However, the greatest thickness variation within the Paradox occurs to the southwest of these faults. Figure 4.4 C shows the base of the salt section in order to show the amount of offset on these faults.

4.1.3  Deposition center changes in the northeastern Pine Ridge minibasin

Isochron minibasin stacking patterns were generated in the northern Pine Ridge minibasin to understand how depocenters evolved through time. Isochrons were used instead of
Figure 4.3: Published interpretations over the Pine Ridge Diapir in addition to this thesis interpretation. A.) Kluth & DuChene (2009) interpreting basement faults as normal. B.) Rasmussen, 2014 showing interpreting basement faults as predominantly reverse. C.) This thesis interpreting basement faults as normal.
Figure 4.4: Interpreted basement normal faults across the 3-D seismic cube. A.) Coherency volume with normal faults picked B.) Fault framework across the 3-D seismic C.) Base of salt surface showing offset and influence of the basement faults
isopachs in order to eliminate errors associated with the depth conversion. This was done both in cross section view (Figure 4.5) as well as a map view (Figure 4.6, Figure 4.7, and Figure 4.8) to better visualize changes within the minibasin.

Figure 4.5 shows a cross section through the north, center and south sections of the northern Pine Ridge minibasin. The location of greatest deposition is indicated by isochron thicknesses, shown with a red arrow, and changes within these vertical isopach stacking pattern are indicative of salt migration through time. Additional thickness in deposition packages are interpreted to indicate salt withdrawal and subsequent accommodation. As salt withdrawal continues to move into a diapir, the accommodation also shifts. Autochthonous salt withdrawl will move in the direction of salt wall growth. In this way the isochron thicknesses are a proxy for understanding salt withdrawl histories.

Depocenter migrations were also observed in map view. Although the overall changes remain constant from the cross section view, more subtle details can be observed within the isochron thicks and can be seen in Figure 4.6, Figure 4.7, and Figure 4.8. Areas of greatest depositional thicknesses are shown in cool colors, with the depocenter identified. Shifts in the depocenter migration are shown with a red arrow. The migration within the depocenter is indicative of salt evacuation under the overburden, most likely from the forcing factor of changes in the sediment supply pattern. The approximate location of the salt wall is shown in dashed blue, unconformities are shown in undulating red.

_Time 1: Base of Honaker Trail to the top of Honaker Trail isochron, see Figure 4.6._ The Honaker trail shows a change in isochron thickness within the northern Pine Ridge minibasin, with the greatest thickness at 325 ms and the thinnest at 175 ms (or a change of about 150 ms). Thinner deposition occurred to the southeast, while thicker deposition occurred directly adjacent to the present day diapir to the northwest. Although not prominent, this would suggest salt withdrawal had begun by Honaker Trail deposition and was evacuating more quickly out of the northern portion of the minibasin. A rapid change in thickness can also be seen adjacent to the
• 1-2 Honaker Trail to lower Cutler: shift to northeast
• 3 Elephant Canyon: southeast portion of minibasin had depocenter shift to southwest
• 3-4 Intra-Cutler unconformity: depocenter change south and slightly west
• 4 Organ Rock: prominent shift of depocenter to the southeast
• 6 Moenkopi: shift towards Uncompahgre and Paradox Valley
• 7-8 Chinle: migration to a position adjacent to present day Pine Ridge Diapir
• 9-10 Wingate to Enrada: shift towards northeast

**Figure 4.5:** Seismic Interpretation across the northern Pine Ridge minibasin
Figure 4.6: Isochron of the northern Pine Ridge minibasin Honaker through Organ Rock time.
salt wall. However, this is an artifact created by the “regional” Cutler unconformity, (see Figure 4.6).

**Time 2: Top of Honaker Trail to lower Cutler beds isochron, see Figure 4.6.** The lower Cutler beds experienced significant salt movement. As salt evacuated, additional accommodation was generated and thicker deposits were able to form. Here the change in isochron from the thickest to the thinnest deposits is 225 ms; thickest in the northeast, suggesting that withdrawal may be moving into the Paradox Valley salt wall at this time. There is no direct contact relationships between the Lower Cutler Beds and the salt due to the unconformity.

**Time 3: Lower Cutler beds to Elephant Canyon isochron, see Figure 4.6.** Isochron thicks continued depositing in the northeast part of the minibasin, however there was a slight shift to the north. Here the isochron range in time is about 200ms. Although there is a shift in the depocenter, the depocenter axis still runs northwest to southeast like the previous two isochrons. Again, there is no direct contact with the Elephant Canyon and the Paradox salt due to the “regional” unconformity.

**Time 4: Elephant Canyon to Organ Rock isochron, see Figure 4.6.** The location of the thickest depocenter continues to be to the northeast, with changes in the isochron of 140 ms. However, the depocenter has shifted to the southeast from the previous isochron. This could be an artifact within the data, rather than an actual depocenter shift. This may arise because the dataset does not extend far enough to see the actual location of the isopach thick. Salt movement is interpreted to continue shifting towards the Paradox Valley salt wall during this time, generating accommodation in the minibasin.

**Time 5: Organ Rock to White Rim isochron, see Figure 4.7.** During the Organ Rock deposition, the depocenter shifted back to the northwest. The Organ Rock is fairly consistent throughout the minibasin and only changes about 50 ms. The axis of deposition continues to run northwest to southeast.
Figure 4.7: Isochron of the northern Pine Ridge minibasin. White Rim to Chinle time.
**Figure 4.8:** Isochron of the northern Pine Ridge minibasin Wingate to Navajo time.
Time 6: White Rim to Moenkopi isochron, see Figure 4.7. Here the isochron varies about 180 ms from the thickest point to the thinnest point across the northern minibasin. The thickest section still resides to the northwest, however it has shifted slightly further north since Time 5. The driving mechanism is interpreted to be salt evacuation into the Pine Ridge Diapir. The southwest section had little accommodation and/or was sediment starved.

Time 7: Moenkopi to Lower Chinle isochron, see Figure 4.7. A significant change in the depositional center occurred during the lower Chinle beds, and minibasin thick patterns shifted to the southeast when salt was removed from the center of the basin along the axis. Salt flow was both into the Paradox Valley salt wall as well as the Pine Ridge diapir. The isochron has about 110 ms difference between the thinnest and thickest sections.

Time 8: Lower Chinle to Upper Chinle isochron, see Figure 4.7: The depocenter again shifted during the Upper Chinle and once again was thickest to the northwest directly adjacent to the present day Pine Ridge diapir. The isochron varies 120 ms between the thinnest and thickest sections.

Time 9: Chinle to Wingate isochron, see Figure 4.8. Deposition centers in the Wingate remained close to the Pine Ridge diapir to the northwest. However, there was a slight shift to the north compared to the previous isochron in time 8. The isochron changes about 140 ms from the west to the east.

Time 10: Wingate to Navajo isochron, see Figure 4.8. The depocenter again shifted during the Navajo, and once again was the thickest to the northwest near the present day Pine Ridge diapir. The isochrons range about 140 ms across the northern minibasin.

4.2 Pine Ridge Restoration

Restorations were constructed across the Pine Ridge diapir and its associated minibasins in order to understand the evolution of the salt body and minibasin deposits as well as validate seismic interpretations. The northern Pine Ridge minibasin has good data constraint from the 3-D seismic dataset, where as the southern Pine Ridge minibasin is only controlled by a
projected 2-D line drawing interpreted from Kluth & DuChene (2009), corroborated by 2-D seismic line observations. Completed restorations give a good indication of evolution of the Pine Ridge minibasin and diapir geometries throughout the basin history.

In order to establish the best orientation for the restoration, a stereoplot was generated using dip data obtained from field measurements; and a traverse (T) orientation was calculated. Figure 4.9 displays the dip data from the field, and the resulting stereoplots used in calculations. The best fit line through the dip data was measured at 041 degrees, and this orientation will therefore display the most accurate cross section representation of the structure. Therefore, the restorations are completed at this orientation. It should be noted that 3-D inlines are oriented at 040 degrees, and are a therefore very accurate representation of the structure (see Figure 4.9).

During each restoration stage, decompaction values were calculated according to the Sclater & Christie (1980) equation, the decompaction parameters are presented in Table 4.4. As restorations were completed, the mean airy isostatic response was also noted, as seen in Table 4.4. This was done in order to see how much influence compaction has had on original deposition thicknesses as well as understand what the salt geometries were at the time of deposition.

Restorations began at present day geometries, and work backward through time to remove influences of deposition on the resultant salt structure. The present day elevation between 2,190 and 2,520 meters (variation due to topography) was used for the first step in the restoration process. However, the next restoration involves the Mancos shale, which was deposited in a marine environment indicating 0 to -200 meters of elevation. This drastic change in elevation occurred because of the Colorado Plateau uplift around 5-10 Ma (Trudgill pers. comm 2017), long after any depositional record exists in the area. Restorations are impossible for this uplift without the depositional record, and therefore, it is assumed that the uplift did not influence the internal structural geometries or thicknesses, but instead bulk shifted the entire structure up in elevation. Therefore, starting with these restorations, the Mancos was bulk
Figure 4.9: Optimal cross section orientation across the study area using dip data acquired from the field. The above map shows the study area with a 2X vertically exaggerated digital elevation model from the Utah Geological Society. Dips are displayed in formation color, which is the same as the rose plot. Draped on top is the La Sal Quad 30’ x 60’ Quadrangle by Doelling, (2004). Well paths are displayed, and the data boundary is in red. Rose plots show datapoints as well as optimal orientation of cross sections.
### Table 4.4: Decomposition values used in the restorations

Decomposition was calculated using Sclater and Christie’s (1980) equation. The mean isostatic response was calculated for each of the units during restorations.

<table>
<thead>
<tr>
<th>Values used for decomposition</th>
<th>Unit</th>
<th>Sandstone (%)</th>
<th>Shale (%)</th>
<th>Limestone (%)</th>
<th>Porosity</th>
<th>Vshale Depth Coefficient (km(^{-1}))</th>
<th>Grain size (cm)</th>
<th>Density (kg/m(^3))</th>
<th>Youngs Modulus (Mpa)</th>
<th>Poissons Ratio</th>
<th>Decompaction (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mancos</td>
<td>25</td>
<td>75</td>
<td>0</td>
<td>0.59</td>
<td>0.45</td>
<td>0.75</td>
<td>0.0129</td>
<td>2702</td>
<td>28125</td>
<td>0.3</td>
<td>7.8</td>
</tr>
<tr>
<td>Dakota</td>
<td>25</td>
<td>75</td>
<td>0</td>
<td>0.59</td>
<td>0.45</td>
<td>0.75</td>
<td>0.0129</td>
<td>2702</td>
<td>28125</td>
<td>0.3</td>
<td>8.5</td>
</tr>
<tr>
<td>Burro Canyon</td>
<td>25</td>
<td>75</td>
<td>0</td>
<td>0.59</td>
<td>0.45</td>
<td>0.75</td>
<td>0.0129</td>
<td>2702</td>
<td>28125</td>
<td>0.3</td>
<td>8.8</td>
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<tr>
<td>Brushy Basin</td>
<td>60</td>
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<tr>
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<td>40</td>
<td>0</td>
<td>0.55</td>
<td>0.37</td>
<td>0.4</td>
<td>0.0244</td>
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<td>22000</td>
<td>0.3</td>
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<tr>
<td>Summerville</td>
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<td>40</td>
<td>0</td>
<td>0.55</td>
<td>0.37</td>
<td>0.4</td>
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<td>2678</td>
<td>22000</td>
<td>0.3</td>
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<td>Entrada</td>
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<td>40</td>
<td>0</td>
<td>0.55</td>
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<td>0.55</td>
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<td>2664</td>
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<td>27.7</td>
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</table>

### Airy isostasy effect

<table>
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<th>Mean Isostatic response (m)</th>
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</thead>
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<tr>
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<td>190</td>
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<tr>
<td>Elephant Canyon</td>
<td>290</td>
<td>61.2</td>
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</table>
shifted downward to match the interpreted depositional paleo-elevation.

Restoring some units had a minimal affect on the resultant salt and minibasin geometry, therefore, only key restorations are shown in Figure 4.11 and Figure 4.12. The last restoration completed is to the top Honaker Trail. Faulting is interpreted to occur before salt deposition and the Laramide Orogeny did not reactivate these surfaces. Therefore, the faults are older than the salt, and are not restored.

Present Day (Figure 4.10)-The present day Pine Ridge salt diapir shows asymmetry between the northern and southern minibasins. Although both the northern and southern minibasins show the same general salt geometry trends, minibasin formation thicknesses differ as well as the details of the salt geometry. Most notable thickness differences between the minibasins occur in the White Rim, Organ Rock, Elephant Canyon, and Honaker Trail formations. The White Rim is much thicker in the southern minibasin, however, the Organ Rock, Elephant Canyon and Honaker Trail formations are much thicker in the northern minibasin.

Salt geometries show similar histories, however, the details in the diapir shape are unique to the minibasins. The southern and northern minibasins both show a broad salt cusp in the Navajo, Entrada, and Summeville. However, the southern minibasin has three additional cusps, whereas the northern minibasin only has one additional cusp. The three cusps in the southern minibasin are more sharply defined, and although not fully developed, appear to be the beginnings of salt wings or christmas tree structures, where extrusion of salt occurs in the former and stacking of these patterns occurs in the later (Ladzekpo et al., 1988; Yielding & Travis, 1997; Jackson & Hudec, 2017). This is indicative of salt rise rate greater than sediment accumulation rate. The three cusps in the Pine Ridge southern minibasin occur in the Elephant Canyon, Organ Rock and White Rim sandstones. The Pine Ridge northern minibasin broad cusp occurs mostly in the White Rim, however it starts in the Organ rock Formation. These differences between the minibasins would suggest that the southern minibasin experienced two seperate pulses of high
Fig. 4.10: Present day minibasin and salt geometries used in restorations. Formations are labeled with corresponding seismic package numbers. However, some formations were interpreted using well logs and/or outcrop data, and cannot be seen, or were not characterisic enough to be picked on the seismic datase. Therefore, these units do not have a corresponding seismic package number. The northern minibasin was interpreted using the 3-D seismic cube, the southern minibasin was interpreted using 2-D regional seismic lines, well logs and Kluth & DuChene’s (2009) interpretation projected onto the cross section.
salt rise rate (compared to sediment accumulation rate) whereas the northern minibasin only experienced one longer time span of higher salt rise rate.

The salt diapir is close to the surface and only buried about 200 m below ground surface. Stratigraphic thickness changes and some dip changes occur across both minibasins. The southern minibasin shows a slight anticline at the surface, while the northern minibasin defines a slight syncline. The anticline is interpreted to be generated by the rapid evacuation of underlying salt further to the southwest and out of the cross section causing subsidence on the southwestern end of the crosssection.

Honaker Trail (Figure 4.11)- As sediment was shed from the Uncompahgre uplift, differential loading began forming what would become the Pine Ridge diapir. Differences in sediment supply existed, and the Pine Ridge northern minibasin received more sediment than the Pine Ridge southern minibasin. This generated a greater differential load in the north, and salt began to flow and evacuate in response. The evacuating salt generated accomodation, allowing for additional sediment deposition. Because the northern minibasin experienced more differential loading and hence accomodation, sediment thickness in the northern minibasin is thicker than that in the south. As salt evacuated out from under the sediment load, initiation of the diapir began. Sediment accumulation rate was high, and formed a contact relationship where the diapir narrows, and sediment onlaps onto the diapir. In the early stages of the diapir growth, it was still quite broad and was near or at the surface.

Elephant Canyon (Figure 4.11)- The Elephant Canyon restoration continues to show asymmetry between the Pine Ridge northern and southern minibasins. The northern minibasin has a much thicker accumulation of Elephant Canyon than the southern minibasin. The northern minibasin also shows a thickening towards the north, away from the salt diapir. This suggests that the depocenter generated was to the north, presumably from autochthonous salt evacuation, and that the Pine Ridge diapir was a subaerial feature. The northern minibasin had higher sediment vs diapir rise rate as can be seen by the gentle salt slope inward. The southern
minibasin shows a vertical salt wall contact with the sediment, meaning that the diapir rise rate and sediment accumulation rate were equal in this minibasin during Elephant Canyon deposition. The diapir is beginning to be more pronounced than the broad structure seen during the Honaker Trail restoration.

Organ Rock (Figure 4.11)- Organ Rock sedimentation continues to be thicker in the northern minibasin with thickest isopachs to the north. The southern minibasin has a thickening sediment package to the south and away from the diapir. Suggesting that the southern minibasin began to experience some allochthonous salt evacuation beginning during Organ Rock deposition. The southern minibasin also had a pulse of very high sedimentation rate, causing the salt diapir to step abruptly inward. However, the remainder of the Organ rock deposition in the southern minibasin matched the rate of salt rise, as can be seen by the mostly vertical salt walls. The northern minibasin had nearly equal salt rise rates to sediment accumulation rate during the Organ rock deposition, although gradually the salt rise rate increased. The diapir is becoming even more prominent than what was seen in the Elephant Canyon restoration, and the autochthonous salt has thinned dramatically beneath the northern minibasin. At the base of the Organ Rock is what has been described as a regional unconformity. This “regional” unconformity is very prominent within the northern minibasin, however the formation appears to be conformable with the Elephant Canyon and Honaker Trail within the southern minibasin. This can be explained using two different interpretations. The first explanation uses Kluth & DuChene’s (2009) “heel and toe model” (Figure 2.2). The northern minibasin is developed as salt evacuates toward the northern diapir (Paradox Valley salt wall), this generates accommodation, and the depocenter is closer to the Paradox Valley salt wall. Then autochthonous salt moves into the Pine Ridge diapir, causing a depocenter shift towards the Pine Ridge diapir. During the same time the Pine Ridge southern minibasin has not had much influence from the autochthonous salt movement, so any deposition is flat lying and hence conformable with units below. During a regional unconformity event, the northern Pine Ridge minibasin displays an angular unconformity, easily recognizable in seismic, whereas the southern Pine Ridge minibasin is still
Figure 4.11: Key restorations across the Pine Ridge Minibasin with the resultant salt geometries. Salt was at or near the surface for much of the diapir’s growth.
flat lying, and can be missed on seismic. Yet another explanation can describe the discrepancies between minibasins, and suggests that the unconformity is not regional. The second interpretation is that during a pulse of Uncompahgre Uplift, additional elevation generated an erosional environment. However, the Pine Ridge diapir could have been a structural feature, protecting the southern minibasin from any erosional effects created from the Uncompahgre Uplift. This would mean that the northern minibasin experienced erosion while the southern minibasin was free from erosional processes.

White Rim (Figure 4.11)-during the White Rim deposition, the depositional patterns shift, with the thickest sedimentation occurring in the southern minibasin, and only a thin White Rim package depositing in the northern minibasin. The White Rim does not show much of an isopach thickness change across the northern minibasin, whereas there are noticeable changes in the southern minibasin. The autochthonous salt has thinned considerably under the White Rim isopach thick in the southern minibasin, with minimal thinning further to the south. The diapir is continuing to be more prominent. The diapir rise rate increased in the southern minibasin through the White Rim deposition, and decreased with respect to sediment accumulation rate in the northern minibasin.

Moenkopi (Figure 4.12)- The Moenkopi still had greater deposition in the southern minibasin and also showed isopach variations, with a thickening toward the diapir. The isopach thicks did not vary strongly in the northern minibasin, and autochthonous salt steadily thinned as salt evacuation occurred. During Moenkopi deposition, the salt diapir was still at or near the surface. The sediment accumulation rate was greater than the salt rise rate for the first part of Moenkopi deposition in the both the northern and southern minibasins. The salt rise rate then became greater than the sediment accumulation rate in the later stages of Moenkopi deposition, although this occurred later in the northern minibasin than in the southern minibasin.

Several restorations are left out due to minimal changes in diapir geometry and depositional changes across the minibasins. These restorations are the Chinle, Wingate, Navajo,
Figure 4.12: Key restorations across the Pine Ridge Minibasin with the resultant salt geometries.
Entrada, Summerville and Salt Wash. This suggests that autochthonous salt was relatively stable from the Chinle onward, although subtle shifts can be seen in the isochrons.

Brushy Basin Restoration (Figure 4.12)- During the Brushy Basin deposition the autochthonous salt welded out beneath the southern minibasin. Isopach thicknesses remained constant across both minibasins, suggesting that not much additional salt evacuation was occurring. The sediment accumulation outpaced the salt rise rate, and completely buried the diapir at this point. The change was gradual in the northern minibasin, but more abrupt in the southern minibasin.

Burro Canyon (Figure 4.12)-Restorating the Mancos back to the original deposition was not possible due to extent of erosion and lack of outcrop. It was determined too many assumptions would be made and the integrity of the data compromised. The actual extent and thickness of the unit could not be ascertained based on the small fragment of the unit left. However, the Burro Canyon was restored back to what was believed to be the depositional location and extent. The salt geometries and the minibasin appear very similar to that seen in the present day geometries. The southern minibasin shows a slight anticlinal structure, whereas the northern minibasin shows a slight synclinal structure within the minibasin. This is interpreted to be a thicker unit of Burro Canyon that blanketed the southern minibasin at a constant formation thickness. In the later stages of the Burro Canyon deposition, evacuation of autochthonous salt to the south (off of the cross section restoration) occurred, bringing the overlying units down to the south. The diapir resembles the present day salt geometry in the Pine Ridge area.
CHAPTER 5
DISCUSSION

5.1 Correlation with Salt Valley Salt Wall

The development of the Pine Ridge diapir can be better understood by comparing it with the Salt Valley Salt Wall (Figure 5.1). These comparisons help describe why salt geometries, stratigraphic architecture, diapir rise rate versus aggradation rates, faulting architecture and the timing of events differ across the basin, and give context for the Pine Ridge Diapir evolution.

5.1.1 Stratigraphic Architecture

As was the case in the Pine Ridge diapir, the Salt Valley salt wall shows a regional tilt in the underlying basement towards the northeast (see Figure 5.2). This supports the argument made by Barbeau, (2003) that the Paradox Basin is a flexural basin caused by the Uncompahgre Uplift. The northeast area of the seismic cube, closer to the Uncompahgre is deeper than the southwest zone located further from the Uncompahgre uplift.

The salt geometry of the Salt Valley salt wall lacks any Christmas-tree or salt cusp structures (see Figure 5.2), and instead has a gentle tapering salt geometry with a slope becomes shallower with time. Minibasin sediment is slightly upturned at the salt-minibasin contact. Using both of these relationships, it is interpreted that the Salt Valley salt wall had a generally higher sediment accumulation versus diapir rise, and gradually the discrepancy became even larger, hence the inward tapering structure and change of slope. The upturned minibasin sediments suggest that there is some drape folding at the minibasin-salt contact. As downbuilding occurs, the center of the depocenter is compacting whereas the sediment adjacent to the diapir is rising.

The Pine Ridge diapir also shows a gradual waning in salt rise rate, as can be seen by the broader salt base tapering towards the surface. After the intra-Cutler unconformity, the sediment accumulation versus salt rise rate was consistent in both the northeastern Pine Ridge minibasin and southwestern minibasin (based on Giles & Lawton, 2002). This is interpreted because the
Figure 5.1: Location of the Salt Valley salt wall dataset in comparison to the Pine Ridge diapir dataset.
Figure 5.2: Interpretation across the Salt Valley Salt wall A.) Composite section across structurally smoothed 3-D seismic cube. Inline 239, crossline 280. B.) Interpretation by Trudgill and Paz (2009) of the Salt Valley Salt Wall. The downlapping surfaces can be seen on the 2D regional lines. C.) Location map of A and B.
minibasin sediment has horizontal contacts with the salt wall. A few pulses in the Cutler exist where the salt extends into the minibasin stratigraphy, and during these smaller pulses, salt rise rate increased, and extended further into the minibasin as denoted by red stars in Figure 4.12. One halokinetic sequence boundary is potentially present during the Elephant Canyon formation that indicates a change in accumulation and sedimentation rates. Halokinetic Sequences, as it was originally described by Giles and Lawton (2002) are “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.” The definition of a halokinetic sequence is dependent on scale, and defined as a feature that resides within 1 km of the salt body. The potential halokinetic sequence seen at the Pine Ridge diapir is 1.1 km away from the salt wall, and therefore not strictly defined as a halokinetic sequence. However, these rapid changes in isopach thicknesses are still generated by salt flow from beneath the flanking minibasin into the growing salt wall, and therefore the processes that formed the Pine Ridge minibasin geometries are the same.

5.1.2 Subsalt and Suprasalt Faulting

The Salt Valley Salt Wall shows a very similar subsalt fault framework to the Pine Ridge diapir. The faults strike northwest to southeast, paralleling the salt wall orientation (see Figure 5.3). The majority of subsalt offset across the entire Paradox Basin is generated by preexisting normal faults which controlled the deposition of evaporites. These faults exist beneath the distal portion of the salt body, near where the autochthonous salt pinches and welds out. Extension was minor after the Permian, and there is no evidence of reactivation in the study area. The faulting geometries suggest that both the Salt Valley Salt Wall and the Pine Ridge diapir began their evolution above presalt “trigger” faults.

However, the later evolution of the salt bodies diverges and they have different suprasalt faulting histories. Suprasalt faults are present within the Salt Valley salt wall (see Figure 5.3), and absent at the Pine Ridge Diapir. Suprasalt faulting within the Paradox Basin is generated
Figure 5.3: Salt Valley salt geometries and subsalt and suprasalt faulting. A.) Dissolution and brittle collapse model used to describe suprasalt faulting at the Salt Valley Salt wall (modified from Gutierrez, 2004). B.) Top of Paradox salt at Salt Valley salt wall colored on elevation (twt. ms) C.) Suprasalt fault polygons displayed on the slick rock formation defined from well logs D.) Subsalt fault polygons displayed on the Base of salt
through dissolution and subsequent collapse of overburden sediment. The dissolution mechanism has normal, near-vertical faults that dip in towards the fold axis, are focused over the anticline of the salt body, and run parallel to the fold axis (Gutierrez, 2004), which closely aligns with the Salt Valley salt wall geometries (see Figure 5.2). Subsidence in the Salt Valley salt wall occurs through brittle collapse, the most common subsidence mechanism (see Figure 5.3) (Christiansen, 1967, 1971; Baars, 2000; as referenced by Gutierrez, 2004). Gutierrez (2004) interpreted the suprasalt faulting in the Paradox Basin as occurring in the Paleogene, Neogene as well as the Quaternary whereas Ge et al. (2008) interpreted the timing of the faults as Eocene to Oligocene.

The Pine Ridge diapir overburden sediments did not experience major extension or extensive salt dissolution and collapse, as seen by the lack of suprasalt faulting; suggesting the diapir was in an area where dissolution was minimal or that the diapir crest was far enough below the surface not to be affected by groundwater circulation. Reconstructions indicate that the salt was near the surface for much of the diapir’s history, but the lack of dissolution can be explained by additional sediment accumulation over the top of the diapir as it was rising, preventing dissolution. Even though the timing of any suprasalt faulting is not well constrained, faulting did not occur until at least the Paleogene (Gutierrez, 2004; Ge et al., 2008), well after sediment records exist within the Pine Ridge diapir area. Therefore, a possibility exists that evidence of suprasalt faulting has since been eroded, and some earlier dissolution did occur.

5.1.3 Timing of Development and Regional Unconformities

The detailed timing of the growth of the Salt Valley salt wall and Pine Ridge diapir flanking minibasins is unclear, as there are no biostratigraphic markers within the Cutler group sediments. However, as salt walls and diapirs developed across the basin, they become progressively younger away from the Uncompahgre Uplift (Kluth & Duchene, 2009). In addition, the Salt Valley salt wall shows a clear difference in the timing of minibasin development to the west (Paz, 2006). Paz describes that during the Cutler, progradational packages show a 6 m.y. period of nondeposition in the Salt Valley Salt Wall minibasin, whereas
the eastern area near the Uncompahgre shows continual deposition. This means that “regional unconformities” may be located in structurally similar positions within a minibasin, but may have vastly different ages from minibasin to minibasin.

### 5.2 Hydrocarbon Potential

Four total petroleum systems exist within the Pine Ridge area (Whidden et al., 2012), however, many of the wells drilled in the area are dry holes. This may be due to the influence of the La Sal intrusives, which intruded the area during the Paleogene. As hydrocarbons are heated up, the molecules begin to break down into smaller and smaller chains. As the hydrocarbon chains become smaller, the hydrocarbons become easier to migrate (Jackson & Hudec, 2017). In addition, during heating, salt expands and becomes easier to flow (Jackson & Hudec, 2017). The flowing salt will encourage evacuation of supporting salt bodies. As the salt flows, the overburden collapses into resultant void, generating fractures and migration pathways. Although overburden faulting is not seen within the seismic, there is evidence of faulting within the well Huber Federal #1-15 in the northeast minibasin. Faulting in the area is most likely below the seismic resolution. With smaller chain gas hydrocarbons and migration pathways through the faulting system, hydrocarbons have the potential to have migrated out of the Pine Ridge structure and its associated minibasins.
CHAPTER 6
CONCLUSIONS

• Correlation between minibasins separated by salt structures should be undertaken with great care. The Pine Ridge diapir shows a similar growth history across the associated minibasins, however, there are changes that occur in one minibasin that are not present in the associated minibasin. Likewise, (as can be seen in the temporal depocenter shifts), sediment thicknesses may change even on the same side of the minibasin. These systems do not act in the manner predicted by simple 2D models (behaving all at once in a uniform manner), instead they respond through subtle spatial and temporal shifts within the depocenters as salt evacuates from the autochthonous layers.

• Correlation between associated minibasins is particularly prone to error where the sediment source is not distributed equally across the minibasins. This can generate drastic changes in thickness of sediment packages, such as seen in the Pine Ridge diapir. This is due to differential loading in one minibasin generating a salt withdrawal response and additional accommodation that is not seen in the associated minibasin further away from the sediment load.

• Timing of the associated salt wall and diapir minibasins in the northern Paradox Basin region cannot be exactly age equivalent due to diapiric structures becoming younger as they move away from the Uncompahgre Uplift.

• The timing of the uplift along the Uncompahgre Front affects when the Pine Ridge minibasin and Salt Valley salt wall minibasins receive a differential load through sediment supply, and therefore salt withdrawal, affecting when and where the associated minibasins develop.

• The growth of the Uncompahgre Uplift and associated sediment flux into the basin affected the structural style of the salt bodies within the Paradox Basin. The Pine
Ridge diapir has minor evidence of halokinetic sequences, and no supra-salt faults, whereas the closer Salt Valley salt wall has extensive halokinetic sequences and suprasalt faults. This is assumed to be due to larger vertical offset on the bounding thrust in the northern Uncompahgre and less in the southern portion. The closer Salt Valley salt wall is exposed to a more steady sediment supply from the Uncompahgre uplift, and therefore has both greater sediment accumulation as well as greater salt withdrawal due to differential loading. The increased salt withdrawal activity leads to the generation of halokinetic sequences.

- Field data are consistent with the main thickness changes and halokinetic sequence packages being developed in the Elephant Canyon/lower Cutler beds. This can be seen with the rose diagram of the field measurement dips. The Elephant Canyon has the greatest dips, due to the highest sediment accumulation and accommodation generation. This is the oldest sediment package that can be seen in the study area, so although older sediment packages were also experiencing strong salt modification, there is no field evidence directly over the Pine Ridge area.

- Sediment accumulation versus salt rise rate in the Pine Ridge system was fairly consistent after the Cutler unconformity, with only minor fluctuations (based on Giles & Lawton 2002)

- There is one major “halokinetic sequence boundary” within the northern Pine Ridge minibasin (however, it lies at 1.1 km, just outside the explicit range of definition)

- Extension was not significant across the Pine Ridge diapir after the Permian (as indicated by the lack of basement fault reactivation). The overburden sediments did not experience extension or salt dissolution and collapse, suggesting that the diapir was not near the surface (or close enough to be dissolved during its growth) However, the Salt Valley salt wall experienced dissolution and collapse.
REFERENCES


## APPENDICES

### Table A.1: Table showing Utah and Colorado wells used near the Pine Ridge Diapir dataset. Highlighted blue represents wells that are located within the seismic cube.

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102
Table A.2: Table showing wells used in Salt Valley salt wall.

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Table A.3: Table showing wells used near the Salt Valley salt wall continued.
Figure A.1: Map of the Pine Ridge area. Map modified from Doelling (2004) updated with field data and aerial photographs.
Figure A.2: Wells containing LAS files used in the Pine Ridge area. The caliper log (black) shows borehole conditions. The gamma ray log (green with a fill pattern showing low gamma ray in yellow and high gamma ray in black) is an indication of lithology. The DRHO is shown in red, and RHOB curve shown in blue.