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FACIES CHANGES ACROSS A DEFORMING SALT SHOULDER, CHINLE FORMATION, GYPSUM VALLEY, COLORADO

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Abstract—A unique exposure of the Chinle Formation in Gypsum Valley in the southern Paradox Basin in Western Colorado documents its interaction with a deforming salt diapir as it partially buried a salt diapir to form a salt shoulder. The Chinle Formation forms a southeastward tapering wedge that thins from 160 to 50 m across the shoulder. Thinning of shales is accommodated by pinchout of braided stream channel sands, and truncation by and onlap onto four unconformities of 2 to 6°. Eight facies associations represent deposition in braided and meandering streams, marshes, ponds, and overbank settings as well as deposition from debris flows shed from the adjacent topographically high diapir. Trends in facies result from syndepositional deformation of the adjacent and subjacent salt. Shales thin, and laterally stacked braided stream channel sandstones pinch out toward the diapir. Paleosols are replaced by marsh and pond deposits reflecting isolation from the main fluvial system. The location of marsh and pond deposits in local basins reflects syndepositional subsidence on the inboard parts of the salt shoulder. Salt shoulders are a previously unrecognized, but common element in the diapirs of the Paradox Basin. The unique setting allows preservation of both fluvial facies and deforming salt, and as far as we know, this is the first study relating sedimentation and deformation in this setting.

INTRODUCTION

Current models of diapirism indicate depositional changes in response to continued deformation until the burial and inactivation of the diapir. Most models describe diapirs that rise episodically until they are completely and uniformly buried. Such models have been applied to the breached salt diapirs of the Paradox Basin (Ge et al., 1996; Ge and Jackson, 1998; Guerrero et al., 2015). In this paper, we present field observations of the nature of fluvial facies deposited on a salt shoulder, which represents an important, and hitherto undescribed structural element of salt diapirs. A salt shoulder is a zone at the margin of a salt diapir where the margin steps relatively abruptly inboard. This creates a "step", above which the rising salt diapir is narrower. While the salt shoulders we describe are from an elongate salt wall, similar features may occur on other salt structures, wherever burial changes the behavior of the rising salt. They have been identified on seismic and in well logs, from offshore Brazil (e.g., Campos and Santos Basins), the Pricaspian Basin, the North Sea, as well as deep-water Gulf of Mexico (Demercian et al., 1993).

Diapirs of the Paradox Basin

Gypsum Valley is the southernmost of the salt diapirs in the Paradox basin (Fig. 1). The Paradox basin formed during the Ancestral Rocky Mountain (ARM) orogeny. The Paradox basin contains 16 elongate salt diapirs, eight of which have been breached and expose the salt and flanking strata (Joesting and Byerly, 1958; Cater and Craig, 1970; Hite et al., 1972; Baars and Stevenson, 1981). The diapirs have been described as northwest elongate "salt walls" that range from 1 to 5 km wide and up to 30 km long (Fig. 1). The diapirs and basin are flanked on the northeast by the Uncompahgre uplift (Hanshaw and Hill, 1969; Matthews et al., 2007). Subsidence began during the Pennsylvanian when salt was deposited as the Paradox Formation (Elston et al., 1962). As the proximal minibasins filled with sediment, the depositional locus and the youngest diapirs migrated to the southwest. Sediment shed from the Uncompahgre uplift loaded the salt, driving the formation of the salt anticlines (Elston et al., 1962; Ohlen and McIntyre, 1965; Banham and Mountney, 2013). Beginning in the Pennsylvanian to Permian, salt diapirs rose as the Paradox basin was rapidly filled with sediment derived from evolving stacked thrust faults of the Uncompahgre uplift (Cater and Craig, 1970; Mack and Rasmussen, 1984; Kluth and DuChene, 2009). These salt anticlines and accompanying synclines trend northwest, paralleling the Uncompahgre plateau's orientation and probably parallel the orientation of basement faults that began after salt deposition (Shawe et al., 1968; Kluth and DuChene, 2009; Trudgill and Paz, 2009). Permian through late Jurassic strata rotated into the adjacent synclines, while the strata adjacent to the diapir remained near the surface and were episodically thinned by erosion (Cater and Craig, 1970; Mack and Rasmussen, 1984). Synclines adjacent to Gypsum and Paradox Valleys provided salt for the cores of these anticlines (Cater and Craig, 1970). In the latest Jurassic, most of the diapirs were buried by the Morrison Formation, and diapiric rise ceased (Elston et al., 1962; Cater and Craig, 1970; Trudgill and Paz, 2009).

In the Late Cretaceous, the Sevier orogeny in central Utah resulted in subsidence in the area associated with the Sevier foreland basin (Trudgill and Paz, 2009). The Gypsum Valley was buried under approximately 1.5 km of strata consisting of the Cretaceous Dakota Sandstone, Mancos Shale and Mesa

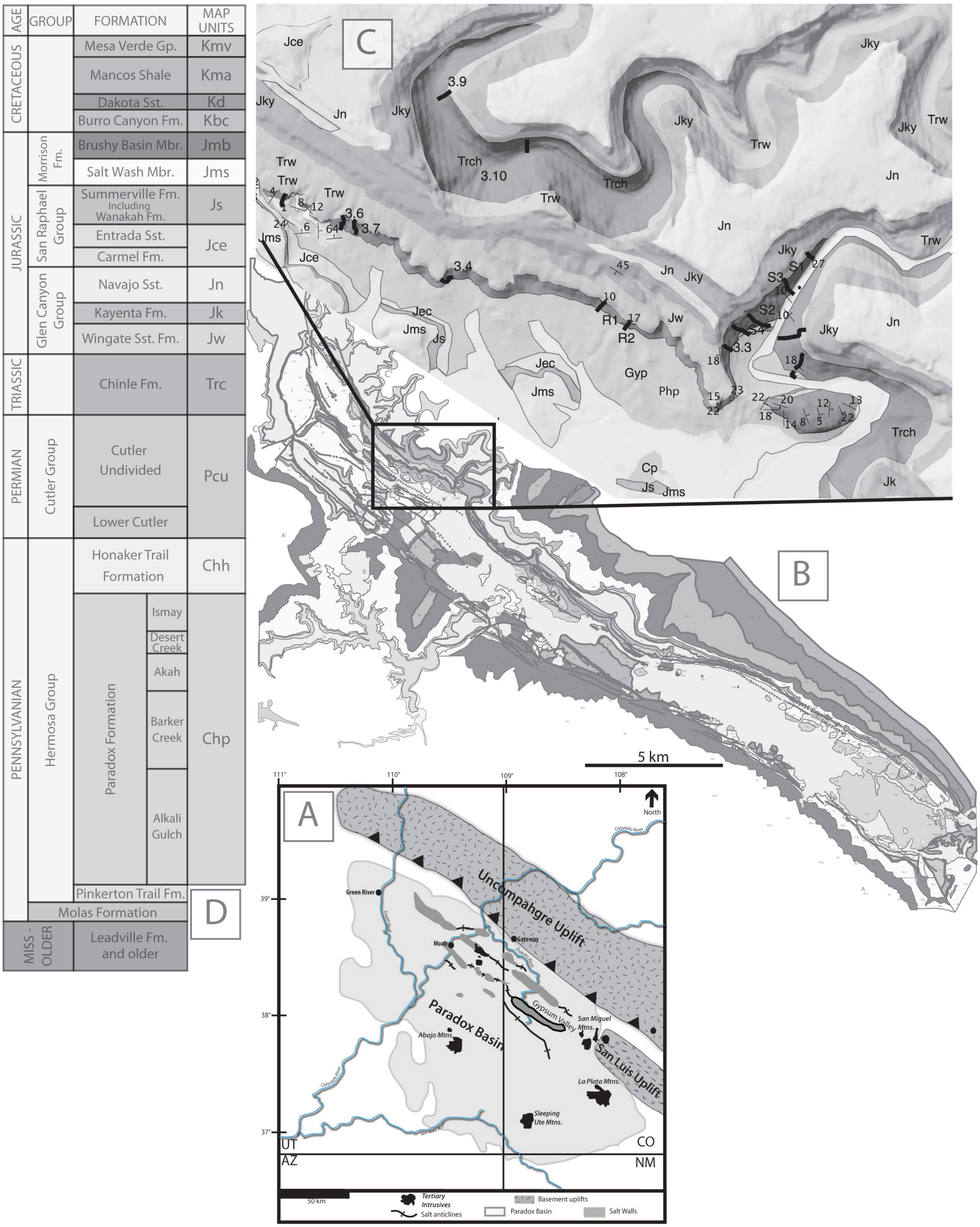

FIGURE 1. Stratigraphy and geologic maps of the study area. **A.** map of the Paradox basin, showing limit of salt, and the exposed salt diapirs, and major buried salt anticlines. Gypsum Valley is outlined. **B.** Geologic map of Gypsum Valley showing the location of the study area. The large yellow, Qal covered area in the center of the map is largely buried diapir gypsum and carbonate caprock. **C.** Geologic shaded relief map of the study area showing the geometry of the salt shoulder, the extent of Chinle Formation outcrop and the locations of measured sections used in the study Stratigraphic sections identified by notations S1, R1, ect.). **D.** Stratigraphic section of southern Paradox basin.

Verde Group. During the latest Cretaceous to Paleogene, Laramide orogeny deformation of the Colorado Plateau may have squeezed the salt walls as the region underwent shortening (Ohlen and McIntyre, 1965). Reactivation of basement faults may also have occurred during Laramide shortening, which tilted Mesozoic strata (Kluth and Coney, 1981). Laramide folds bound the Paradox Basin, the Monument upwarp to the southwest, the San Rafael swell and Henry basin to the west, and to the south and southeast margins are the Defiance plateau and San Juan Basin (Baars and Stevenson, 1981; Nuccio and Condon, 1996; Bump and Davis, 2003). Little Laramide deformation is noted in the Paradox Basin itself, with the possible exception of the Glade and Dolores fault zones (Shawe, 1970; Stevenson and Baars, 1985). During the late Tertiary, San Juan volcanism occurred along with uplift of the Colorado Plateau, which resulted in unroofing, incision, and collapse of the diapirs (Ohlen and McIntyre, 1965; Cater and Craig, 1970; Nuccio and Condon, 1996).

Gypsum Valley is 23 km long by an average of 4.5 km wide (Fig. 1.) The southeast end of the basin exposes salt caprock composed of gypsum, folded shales, and dolomite in contact with sediments that range from Permian to Late Cretaceous (Mast, 2016). It has previously been suggested that sediments older than the Latest Jurassic Morrison Formation were rotated away from the diapir as salt was withdrawn from the adjacent minibasins (Nuccio and Condon, 1996; Trudgill, 2011). To the northwest, along the southwestern rim of the anticline, the Pennsylvanian and Permian strata are overlapped and buried by Jurassic Morrison strata beneath an angular unconformity. Pennsylvanian strata dip away from the anticline by 45 to 90° and are truncated by a syn-Morrison unconformity that exposes strata dipping less than 40°. Along the opposite, northeast side of the salt wall, older strata are exposed, including Permian Cutler and Triassic Chinle Formations (Fig. 1). This implies different depositional and deformational histories across the diapir. Part of this story involves the formation of salt shoulders during deposition of the Triassic Chinle Formation through the Glen Canyon Group (Fig. 1). The relationship and history of the salt shoulders is exposed around the northwestern end of the diapir, but is best exposed from where the Dolores River enters and exits Gypsum Valley to its northwestern end (Fig. 1).

Chinle Stratigraphy and Environments

The Chinle Formation is divided into numerous members in different parts of the Colorado Plateau and western Great Plains (Akers et al., 1958; Stewart et al., 1972; Dubiel et al., 1996; Lucas et al., 1997). However, these can be simply grouped into two subsets (Stewart et al., 1972). The lower units in the Chinle Formation contain bentonitic clayey red beds, full of volcanic detritus from the Mogollon Highlands to the south. The upper unit contains feldspathic red beds that coarsen to silty-sand and pebble conglomerates (Riggs et al., 2013). The lower unit contains the Monitor Butte, Petrified Forest, Shinarump and the Moss Back members. The upper unit includes Owl Rock, Church Rock and related members (Fig. 2) (Stewart et al., 1972). The lower members of the Chinle Formation are not present in southwestern Colorado (Stewart et al., 1972; Hazel, 1994; Lucas et al., 1997). Recent studies correlate the Gypsum Valley stratigraphy with members of the upper unit with the Dolores Formation in western Colorado (Fig. 2; Lucas et al., 1997; Martz et al., 2014).

Shawe et al. (1968) identified the Moss Back, Petrified Forest and Church Rock members in the Gypsum Valley area (Fig. 2). A normal Chinle section, uninfluenced by deformation on the salt anticline, consists of two sandstone and conglomerate channel fill units, the Mossback and the unit equivalent to the Black Ledge (Shawe et al., 1968; Fig. 2). Each is overlain by a thick shale-rich unit. The Mossback and Black Ledge are separated by the slope forming, mud-rich Petrified Forest Member. The middle member of the Church Rock separates the Black Ledge and the upper member of the Church Rock (Fig. 2).

Study Area

The study area is an exposure of the northeast margin of the Gypsum Valley salt wall. The Dolores River flows across the floor of Gypsum Valley and exits into a canyon with salt caprock and Triassic Chinle Formation through Jurassic Navajo Sandstone on the walls (Fig. 1). Salt caprock crops out in the base of the canyon and in the base of the cliff, and underlies the Chinle Formation beneath an irregular unconformity. The Chinle Formation is folded into an open anticline, with strata adjacent to Gypsum valley almost flat, and strata within the canyon dipping progressively more steeply, reaching 27° (Figs. 1, 2). Salt caprock underlies the Chinle Formation for a distance of 400 m into Dolores Canyon. Northwest of the canyon, the Chinle Formation crops out for 4 km to where it is truncated by the Wingate Sandstone across an angular unconformity (Fig. 1).

METHODS

A total of 15 stratigraphic sections ranging from 40-200 m in thickness were measured and correlated using a Jacob staff and located using handheld GPS (Fig. 1). Units were physically correlated across each outcrop panel by tracing marker horizons. These were then transferred to photomosaic panoramas to create cross-sections for analysis. In the Dolores River Canyon, the salt shoulder and angular unconformities within the Chinle Formation were correlated and mapped in three dimensions using GPS-located photographs and Agisoft Photoscan photogrammetric modeling software. Geology was added using the software Vulcan Maptek to create polygons, which were then imported into Midland Valley Move software and combined with orientation data from outcrop measurements with the Brunton Compass.

RESULTS

Shoulder Geometry

Figure 3 illustrates the geometry of the Chinle Formation in

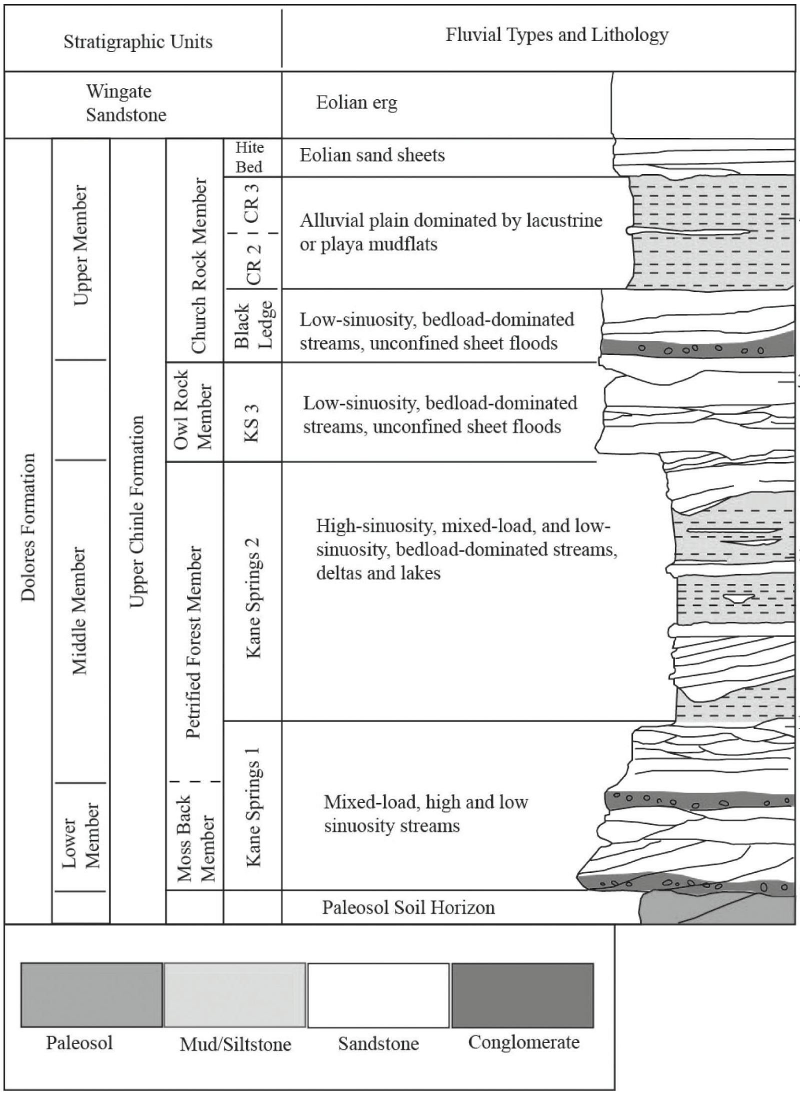

FIGURE 2. Generalized stratigraphic column of the Chinle Formation in the study area showing the alternating coarse and fine units and correlation with other Chinle and Dolores Formation members (stratigraphic correlations from Shawe et al., 1968; Stewart et al., 1972; Dubiel et al., 1989; Hazel, 1994; Lucas et al., 1997).

the study area, based on isopachs derived from the stratigraphic sections and additional measured thicknesses. The maximum thickness is 160 m, thicker than that measured to the northeast by Dubiel et al. (1989). Along the inboard, southeast margin of the outcrop, the Chinle Formation varies from 44 to 86 m thick. The thickest sections fill syndepositional structural basins, ranging from 1 to 2 km long parallel to the shoulder outcrop and 300 m wide across the shoulder (Fig. 3). The salt-Chinle Formation contact forms an inclined anticline, trending N60°W (Figs. 3, 4). Strata dip up to 25° along the northeast edge of the outcrop, where the Chinle Formation dives into the subsurface. (Figs. 3, 4).

Normal faults are found inboard of the steepest curvature of the fold (Fig. 4). The faults extend through the Navajo Sandstone at the top of the outcrop. Thickness and facies do not change across the faults indicating that the faults were not active during Chinle time and post-date deposition. Most faults do not offset the Morrison Formation, indicating they formed prior to the Latest Jurassic. Other faults to the southeast of the study area seem to mark gravity driven subsidence into the salt and are probably Neogene.

Chinle Facies

The Chinle Formation in the study area contains many of the facies described in other locations, but also contains facies unique to the diapir margin. Eleven lithofacies were identified (Table 1). Miall's (1978) facies codes were applied for ease in correlation with other studies. These were grouped into eight facies associations (Fig. 5). Facies in the Chinle Formation form upward-fining cycles that grade from ledge forming conglomerate/sandstone to siltstones and mudstones.

The eight facies associations were defined by lithofacies groupings, bedding geometry and sedimentary structures (Fig. 5). These facies associations include, 1) Caprock-bearing stratified conglomerate and sandstone channel fill (FA1), 2) Noncaprock-bearing stratified conglomerate and sandstone channel fill (FA2), 3) Overbank deposits (FA3), 4) Paleosols (FA4), 5) Ponds/marshes (FA5), 6) Lacustrine deposits (FA6), 7) Heterolithic channel fill (FA7) and, 8) Disorganized conglomerate (FA8).

Facies Association 1 – Caprock-Bearing Stratified Conglomerate and Sandstone Channel Fill.

FA1 consists of channel forms filled with sandstone and conglomerate (Fig. 5). Sandstone and conglomerate ranges in color from tan to reddish purple. Clast compositions of conglomerate is predominately composed of limestone and dolomite, similar to the clasts exposed in the subjacent salt caprock. Caprock clasts are subrounded to rounded, and well to moderately-sorted. Conglomerates are matrix-supported and in some instances contain lenses of imbricated clasts. The matrix is composed of subrounded, well-sorted sand with abundant feldspar and carbonate grains, with compositions ranging from lithic arkose to feldspathic litharenite.

FA1 crops out as medium to thick ledge-forming single, laterally stacked, or vertically and laterally stacked channels (Fig. 4). FA1 channels fine upward into FA3 (overbank deposits), or are overlain by another FA1 channel fill.

The caprock-bearing channel fill is interpreted as forming in a high-energy fluvial system that transported the clasts that form the basal lags of channels. High width-to-depth ratios indicate broad braided streams with intermittent, probably seasonal flow (Blakey and Gubitosa, 1983, 1984; Dubiel, 1987a; Dubiel et al., 1996). The caprock clasts are not the

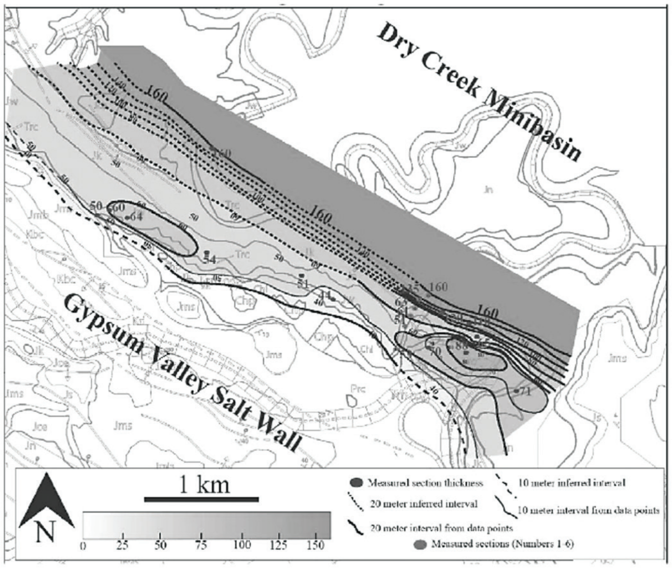

FIGURE 3. Isopach map of the Chinle Formation in the study area showing the thinning onto the shoulder and the small structural basins on the inboard side of the shoulder. The Dry Creek minibasin is the salt withdrawal basin northeast of Gypsum Valley.

carbonate clasts of Dubiel et al. (1991), which were eroded from finer-grained paleosols and lacustrine nodules. This facies association is similar to other descriptions proximal to salt anticlines including that of Matthew et al. (2007) fluvial channel-fill sandstone in its geometry, erosive base and lithologic characteristics. However, those channels are floored with mud clast lags rather than caprock-derived carbonates. Shock (2012) documented similar carbonate caprock incorporated into fluvial channels of the Cutler and Moenkopi Formations.

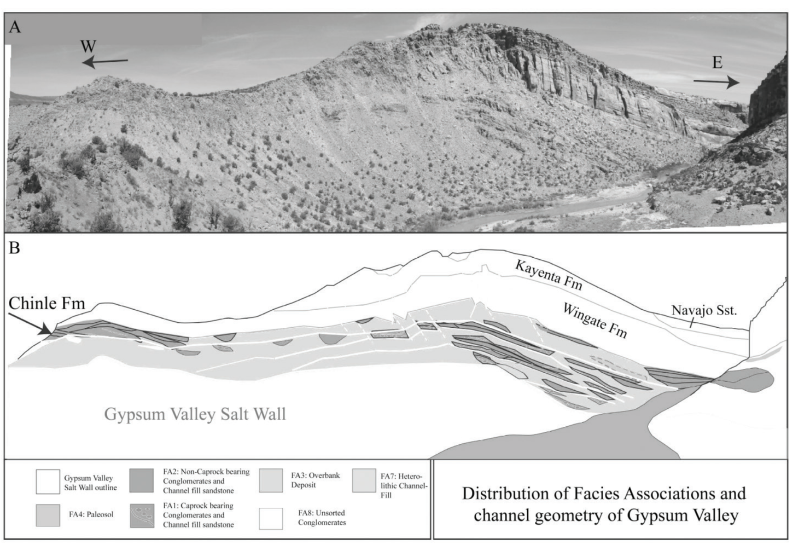

FIGURE 4. Geometry of the Chinle Shoulder. A. outcrop panorama of the north side of the Dolores River Canyon. Sections 3.1 to 3.5 with correlation of key marker horizons and the salt-Chinle contact. Note the faults to the left of the crest of the anticline. B. Surface model created in Midland Valley Move software showing the geometry of the top of the salt shoulder. Correlations of stratigraphic sections showing the onlap and pinchout of units onto the shoulder. Note the thinning of the shales toward the diapir. Also note the thick sandstones and conglomerates adjacent to the diapir that contain caprock clast conglomerates.

Facies Association 2 – Non-caprock bearing stratified conglomerate and sandstone channel fill.

FA2 consists of channel forms filled with sandstone and conglomerate (Fig. 5). Sandstones and conglomerates range in color from tan to reddish brown. The conglomerate is predominately composed of rounded chert, mud rip-up clasts, sandstone concretions and rare septarian carbonate nodules. Conglomerate beds are matrix supported and in some instances appear imbricated. The matrix ranges from lithic arkose to feldspathic litharenite and is similar in composition to FA1. FA2 crops out in thick ledges as laterally and vertically stacked channels (Fig. 5). A typical channel has an erosional contact at the base and consists of conglomerate containing chert, mud rip up clasts (20% of the channel fill), and trough cross-bedded sandstones (St) or planar tabular cross-bedded sandstones (Sp) (40% of the channel fill). These channels are then capped-by ripple cross-laminated or climbing ripple sandstone (Sr) (40% of the channel fill).

FA2 is interpreted as deposits of a braided stream. High width-to-depth ratios indicate broad streams with intermittent seasonal flow and flooding. Chert and volcanic fragments were probably shed off the Uncompahgre Highlands, whereas the nodule- and concretion-derived clasts were most likely eroded locally from FA3, FA4 or FA5 (Dubiel et al., 1996; Dubiel, 1987b; Martz et al., 2014). Martz et al. (2014) also interpreted similar conglomerates as being a braided fluvial system.

Facies Association 3: Overbank deposits

FA3 is composed of red, purple, green, and gray siltstone, mudstone, claystone, and silty-sandstone that exhibits patchy mottling (Fig. 5). FA3 units crop out as poorly exposed slopes. Sedimentary structures include horizontal lamination and climbing ripples. Individual beds are cm-scale and form units ~16 meters thick that extend several hundred meters. Individual sandstone intervals are 2 to 5 meters thick and laterally grade into siltstone. Overbank deposits form contemporaneously, and are interbedded with FA4 paleosols, FA5 ponding deposits, and FA6 lacustrine deposits.

FA3 is interpreted as overbank mudstones deposited in low energy and suspension settings. Sandstone deposition would have occurred as splays or flood deposits. As overbank deposits became more widespread, paleosols formed. Similar strata have been documented at other diapirs (Dubiel et al., 1989; Matthews et al., 2004, 2007; Martz et al., 2014). Matthews et al. (2007) interpreted these beds as splays or sheets associated with flooding events and noted that root traces and mud cracks indicate subaerial exposure.

Facies Association 4 - Paleosols

FA4 consists of pink, red, green and gray clay/siltstone with very fine-grained sandstone which forms beds 0.5–5.0 meters thick (Fig. 5). Root casts, mottling, sand-filled mud cracks, root traces, concretions, and trace fossils are observed. In

TABLE 1: Facies Associations of the Chinle Formation in the study area.

Lithofacies	Interpretation	Grain/clast size	Abundance/ Distribution	Geometry
Siltstone	overbank deposit	dominantly silt	common throughout	laterally continuous (up to 30 meters thick)
Climbing ripple cross-stratified sandstone	fluvial channel fill, overbank deposit	medium- to very fine-grained sandstone	common throughout	laterally continuous
Ripple cross-stratified sandstone	fluvial channel fill	fine to medium-grained sandstone	common throughout	laterally continuous
Trough cross-stratified sandstone	fluvial channel fill	medium-grained sandstone	common throughout	laterally continuous
Planar-tabular cross stratified sandstone	fluvial channel fill	mine to medium-grained sandstone	common throughout	laterally continuous
Massive sandstone	fluvial channel fill	fine to medium-grained sandstone	sparse throughout	laterally continuous
Unsorted angular clast conglomerate	debris flow	granules to boulders in silt and sand matrix	locally on edge of diapir	laterally discontinuous, lenticular
Chert-bearing conglomerate	fluvial channels	cobbles to granules	bases of channels throughout	lenticular
Caprock clast conglomerate	fluvial channels	cobbles to granules	bases of channels throughout	lenticular
Rip-up clast comglomerate	fluvial channels	cobbles to granules	bases of channels throughout	lenticular
Mudstone	overbank deposits	silt and clay with sand	common throughout	laterally continuous

the study area, lower units of the Chinle Formation form silty mud-rich localized blocky peds, with few reduced root halos. These beds grade into overbank deposits. In the uppermost unit of the Chinle Formation, trace fossils identified as beetle traces (*Scoyenia*) and insect larva burrows (*Fuersichnus*) occur in very fine-grained sandstone.

FA4 is interpreted as paleosols based on root traces and pedogenic fabrics that indicate probable vertisol development with mounded curved fractures (Fig. 5). Prochnow et al. (2006) interpreted similar prismatic peds as vertisols. The clay rich horizons, 0.5 meters in thickness, contain extensive lateral root horizons and lesser-burrowed surfaces that indicate a moisture-rich environment. The types of burrows indicate the presence of a near-surface fresh water table (Dubiel and Hasiotis, 2011). Mud cracks indicate subaerial exposure, whereas mottling of some horizons indicates a fluctuating water table (Dubiel and Hasiotis, 2011). Chinle Formation paleosols exhibiting mottling, desiccation, and burrows have been classified as aridisols or calcisols (Dubiel, 1989; Dubiel et al., 1996; Dubiel and Hasiotis, 2011; Martz et al., 2014)

Facies Association 5 – Marsh and ponds

FA5 are found along the inboard salt shoulder margin contact in the exposures northeast of the Dolores River Canyon (Fig. 1). Red to brown, poorly lithified mudstone, siltstone, and sandstone contain abundant septarian nodules, sandstone concretions, petrified wood fragments, carbon films of plant remains, thin– shelled unionid bivalve fossils (*Triasmanicola*) (Dubiel, 1987a; Parrish and Good, 1987; Dubiel et al., 1991), and rare crayfish burrows (*Camborygma*) 6 to 12 cm in width (Fig. 5). The fossils, concretions and nodules form in a matrix of massive siltstone with minor sandstone. Fabric is rarely preserved as bioturbation disrupts original primary sedimentary structures. FA5 forms lenses ~1 meter thick and ~20-50 meters wide, with gradational lateral margins merging with overbank deposits. Topographically, FA5 is unique to basins forming on the inboard part of the shoulder where beds have low dips that create along-strike thickness variations of >30 m.

FA5 is interpreted as local ponds formed in floodplains. The unionid bivalves have been interpreted as a transported assemblage in crevasse-splay deposits, occurring from high-discharge flood events with disarticulation of shells (Dubiel, 1987a; Parrish and Good, 1987; Dubiel et al., 1991). Inundation by flood events is also evident from different petrified wood samples (i.e., branches and stumps), which would have been deposited during the seasonal flooding events. The high concentration of septarian nodules and concretions indicate a locally high water table (Dubiel et al., 1991). Vertical crayfish burrows and mottling indicate a fluctuating water table. Concretions have been incorporated into the basal lag of channels, indicating they formed near the sediment surface. Carbonized leaf and tree imprints indicates a suitable habitat for vegetation (Dubiel, 1987a) and high concentration of hematitic cement indicates partially oxygenated waters. These ponds were likely the product of localized subsidence and probably fed by seasonal flooding of crevasse splays from channels.

Facies Association 6 –Lacustrine deposits

FA6 is made up of thin isolated lenses extending for 0.5 km in outcrop, and are found in sections on the northwest part of the shoulder (Fig. 1). FA6 is dominated by mudstone and siltstone with few sedimentary structures and lacks plant material

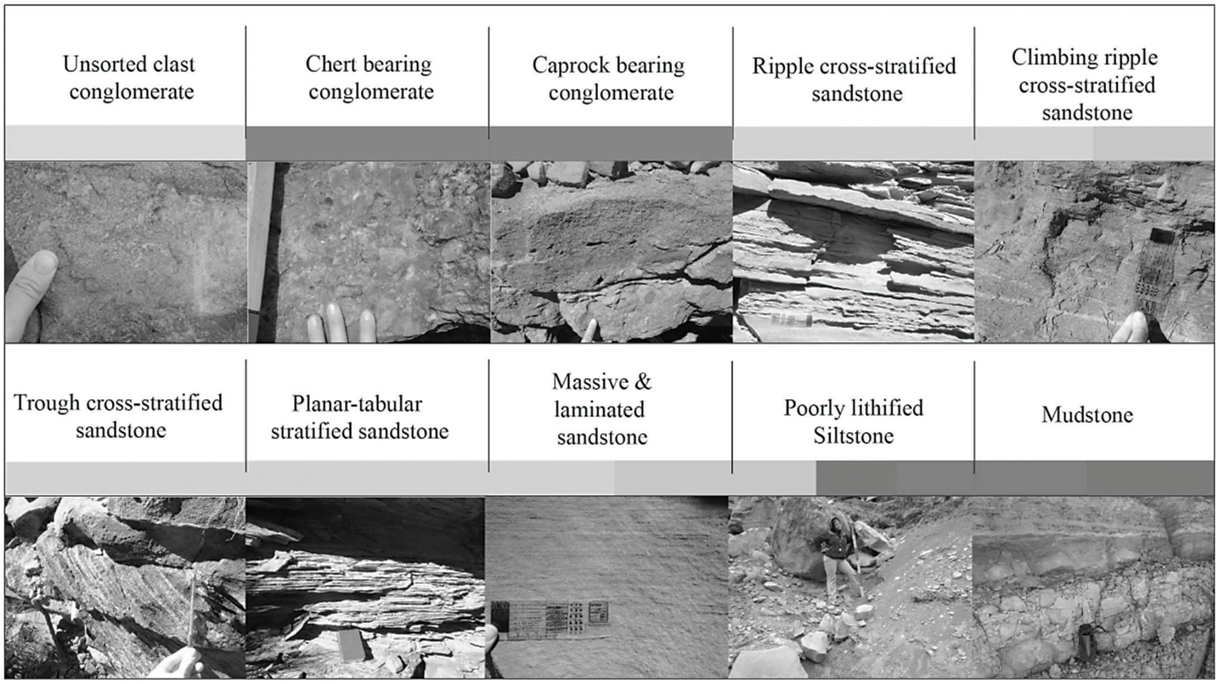

FIGURE 5. Facies of the Chinle at Gypsum Valley. Color bars above images indicate which facies associations contain the facies. Orange is disorganized conglomerate (FA8), purple and yellow apply to both caprock and non-caprock bearing conglomerates and sandstone (FA1-2&7), Red is overbank deposits (FA3), and blue and grey represent wetlands, lacustrine and paleosols (FA 4- 6). facies association 7, heterolithic channel fill can be represented by a combination of all lithofacies. Rip-up clast conglomerate not shown.

and roots. Rare horizontally laminated red and gray siltstone and mudstone are present, but most of these beds are overprinted by mottling and bioturbation.

FA6 is interpreted as the deposits of seasonal floodplain lakes. The fine-grained size indicates suspension deposition. The lacustrine deposits are thought to be low productivity, possibly the result of large scale, episodic flooding events (Dubiel et al., 1991; Matthews et al., 2007). The absence of sedimentary structures may indicate that productivity was episodically high. The lack of plant material could indicate the presence of oxic conditions (Dubiel et al., 1991).

Facies Association 7 - Heterolithic Channel Fill

FA7 consists of one exposed channel form filled with alternating beds of tan and brown sandstone, conglomerate and red claystone. Lithofacies include trough and planar cross-stratified sandstone, stratified sandstone and mudstone/siltstone. Conglomerate beds are similar to those in FA1 and are ~0.5 meters in thickness. Sandstone beds are upper fine- to lower medium-grained and are 0.1-0.6 meters thick. These are arranged in 7-m-tall sigmoidal cross beds with alternating layers of conglomeratic lags, sandstone, and silty shale. The silty shales are up to 1 m thick and thicken at the bottom of the lens. The sandstone sometimes contains interbedded clay drapes, 1-10 cm in thickness. The overlying mudstone contains mudcracks in some places.

FA7 is interpreted as a preserved point bar in a muddy meandering stream with fluctuating seasonal changes in flow. The presence of lateral accretion sets and fining-upward bundles indicate a meandering stream system. Alternations in lithology and grain size indicate cyclic changes in stream energy. Clay drapes and mudcracks represent flooding of bar forms followed by exposure. The presence of caprock-derived clasts indicates exposure of the diapir during deposition or cannibalization of nearby debris flows. Hazel (1994), Dubiel (1987b) and Matthews et al. (2007) documented similar channels within the Chinle Formation and interpreted the deposits as partly confined suspended and mixed-load meandering-stream deposits.

Facies Association 8 - Disorganized conglomerate

FA8 forms brown, tan or purple lenticular cliffs and ledges 2-15 meters thick that extend no more than 10 m in outcrop (Fig. 4). FA8 is found exclusively along the contact between the Chinle Formation and underlying caprock and is restricted to within 200 meters of the southeast margin of the shoulder. The beds contain massive conglomerate. Clasts are angular to subrounded pebbles and boulders composed of sandstone, claystone, limestone, and dolomite. FA8 is usually clast-supported with a matrix of poorly-sorted, fine- to coarse-grained sandstone.

The disorganized conglomerate's clast composition, poor sorting and lack of sedimentary structures are interpreted to represent deposition adjacent to the diapir by debris flows derived from a topographically high diapir. The proximity of deposits to the diapir, isolated nature of the lenses, and presence of caprock clasts necessitates a localized source. Similar debris flows are found adjacent to salt walls in the Triassic Moenkopi Formation in Castle Valley (Lawton and Buck, 2006). Debris flows form part of halokinetic sequences that consists of reworking older deformed strata and the non-evaporite part of the diapir (Giles and Lawton, 2002).

DISCUSSION

The stratigraphy of the Chinle Formation reflects the deformation that occurred during deposition. Figure 4 illustrates the stratigraphy and syndepositional deformation that formed during Chinle time. The Chinle Formation thins from 160 to 54 m across the shoulder within a distance of approximately 1 km (Figs. 3, 4). This is accommodated by thinning of the shale units and pinchout of sand bodies, but mostly by truncation beneath and onlap onto unconformities. The Chinle Formation can be divided into four wedge shaped, unconformity-bound sequences with 2-6° of angular discordance (Figs. 3, 4). While some of these sequences may correlate regionally, they are mostly shaped by local salt tectonism and are thus best termed "halokinetic sequences" (Giles and Lawton, 2002; Andrie et al., 2012). Each sequence was rotated to the northeast toward the Dry Creek Basin Syncline during and after deposition by salt tectonism. Increased subsidence or reduced sedimentation resulted in relative base-level fall and created the unconformities.

In addition to the deformation on the shoulder, the salt diapir continued to at least episodically form a topographic high that shed debris into the diapir. This was partly in the form of debris flows (FA8) adjacent to the diapiric high (Fig. 4). Partly this was as debris was reworked into fluvial channels (FA1). The caprock clasts decrease laterally in abundance and are absent more than 700 m from the edge of outcrop (Fig. 4). This, along with the concentration of debris flow conglomerates along the present edge of outcrop (Fig. 4) indicate that the actively rising, topographically high diapir was very near to the modern edge of outcrop of the Chinle Formation.

The marsh, pond, and lacustrine deposit (FA5 and FA6) are found along the inboard margin of the salt shoulder and are interbedded with debris flows from the diapir. Additionally, the heterogeneity channel fill (FA7) is also found here, suggesting that adjacent to the diapir a marshy, low gradient landscape predominated. Conglomeratic lenses disappear 0.7 km away from the diapir and ripple cross-stratified sandstones thicken away from the diapir. Near the inboard, southeast edge of the shoulder, ribbon geometry and channels with low width to depth ratios are more common (McFarland, 2016). Farther from the margin, channels are laterally stacked with high width-to-depth ratios. Paleocurrent estimates from crossbedding indicate a predominant south to southeast flow. This suggests that the diapir interfered with fluvial transport so that the main axes of transport was farther from the diapir in the salt-withdrawal basins, where subsidence rates were greater.

The facies, fossils and trace fossils are similar to those described in exposures near Bedrock Colorado, 10 km north of the study area. These deposits have been correlated with the Petrified Forest and Church Rock members that compose the Chinle Formation in the study area (Dubiel et al., 1989). Although the bedrock exposure is more distal from the adjoining Paradox salt diapir, Dubiel et al. (1989) found similar facies to those in the study area and made environmental interpretations.

CONCLUSIONS

The Chinle Formation strata reflect active salt tectonism during deposition. Braided streams are replaced by meandering streams near the diapir. Facies are similar to those in other upper Chinle Formation studies (Dubiel, 1987b; Dubiel et al., 1989; Dubiel and Hasiotis, 2011). However, three facies including diapir derived debris flows, diapir-clast conglomerates, and heterolithic channel fills are only found near diapirs. The distribution of facies is shaped by the deformation on the diapir shoulder. Braided stream channels are more common in the distal outcrops, and meandering streams are common in the proximal outcrops. Ponds and lakes form in actively subsiding small basins on the shoulder, and adjacent to the rising diapir and fill syndepositional basins on the shoulder. Internal angular unconformities separate halokinectic sequences formed by rotation of the strata away from the rising diapir and into the adjacent minibasin. Laterally extensive braided stream channel fills pinch out toward the diapir and are replaced with isolated channel braided and meandering stream channel fills adjacent to the diapir.

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