Nearest Neighbor Analysis of Compaction in Basin Margin Sediments of the Capitan Depositional System

by

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Class of 2010
Acknowledgements

I could not have created this project without the funding of the Hughes Scholarship Program and a number of people to whom I am greatly indebted. First, I would like to thank Phil Resor for his patience, generosity, and grace in helping me through the process of creating this work. Second, I would like to thank my readers, Peter LeTourneau and Suzanne O’Connell, for devoting their time and care to reading this document. Special thanks goes out to: Gene Shinn, for providing us with the samples, Dana Royer for assisting me during the writing process, Lexi Malouta for her help in the field, and George Bennum for introducing me to the Guadalupe Mountains National Park. Thanks to Gus Seixas for his companionship and his help every time I needed advice from a fellow student. Finally, I would like to thank my family for supporting me in all my endeavors and always knowing when to offer a kind word of encouragement.
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I. Preface:

Motivated by the exquisite nature of its outcrop exposures and the potential for hydrocarbon recovery, carbonate geologists have devoted countless hours to studying the high-resolution stratigraphy of the Capitan shelf margin in the Guadalupe Mountains National Park. Despite the extensive amount of work that has been conducted in this area, the cryptic nature of upper platform geometries has impeded the development of a unified theory describing the evolution of the Capitan reef and its associated strata.

Hunt et al. (1995) introduced a conceptual model illustrating that compaction of basin margin sediments may have a profound effect upon the evolution of prograding margins like the Capitan. Although studies of the upper platform have documented features consistent with this model, extensive erosion has inhibited the observation of unequivocal evidence to prove its significance. Additionally, traditional methods for quantifying compaction in carbonate sediments require prior knowledge of original burial depth or original porosity, which are both uncertain in the Capitan depositional system. In order to determine the significance of compaction in basin margin sediments, and the role it has played in driving the evolution of the Capitan shelf margin, new methods must be developed. In the first portion of this work, we have provided a detailed description of the Capitan shelf margin, and the controversies related to its development. In the second portion of this work, we introduce a new method that can be used quantify compaction in fossiliferous limestones whose original burial depth is uncertain. We apply this method to basin
margin sediments of the Lamar Member and demonstrate that differential compaction has likely played a role in driving the evolution of the Capitan shelf margin.
Chapter 1. Geologic Setting of the Capitan Depositional System

In the Guadalupian time of the Upper Permian Era, marine transgression, basinal subsidence, crustal mobility, and tropical climate conditions acted in conjunction to promote the growth of algal reef structures around the boundaries of the Delaware (Fig. 1.1) and the Midland Basins (King, 1948; Hill, 1996). Reef growth began during the lower to middle Guadalupian age, when species of frame building...
organisms like calcareous sponge and algal species (e.g. *Tubiphytes* and *Archealothoporella*) began to flourish (Hill, 1996). By the time of the Upper Guadalupian, the reef had developed into a massive, self-supporting microbial boundstone factory that supplied large amounts debris toward the shelf and the underlying basin (Brown and Loucks, 1993A; Kenter et al., 2005). In the modern day, the lithified remains of the reef structure are preserved in the form of the Guadalupe (Fig. 1.2), Delaware, Apache, Davis, and Glass Mountains (Hill, 1996).

Figure 1.2: Map of Canyons within the Guadalupe Mountains, New Mexico and Texas (Saller, 1999).
Although the Capitan Reef surrounded nearly the entire Delaware Basin during the time of its growth, progradation and aggradation of the Capitan Reef was most pronounced on the north central area of the Delaware Basin, in what is now the Guadalupe Mountains (Harris and Saller, 1999). In this area, progradation of the Capitan reef extended for about 5 km into the underlying basin, while aggradation occurred on the order of 150-200 meters (Brown and Loucks, 1993B; Hill, 1996; Tinker, 1998).

Most of the reefs progradation occurred within the first 1.5 million years of the Capitan time during the deposition of the Seven Rivers Formation. Throughout the remainder of the Guadalupian time, the reef experienced a dominantly aggradational phase that steepened the slope to an angle of 30-35 degrees and created about 600 meters of paleobathymetric relief (Longley, 1999). The steep inclination of the shelf margin that was established during this phase is preserved in the basinward dip of upper slope deposits within the Guadalupe Mountains (Tinker, 1998). The last period of Capitan Reef progradation is preserved in the succeeding Tansill formation. Geochemical studies of this formation noted increasing salinity during late Tansill times, which many workers have attributed to increasing aridity and an end to normal marine circulation around the Capitan Reef (Given and Lohmann, 1985; Kirkland and Chapman, 1990).
1.2 Reef Strata

Massive reef strata near the Capitan Shelf margin are comprised of the Goat Seep and the Capitan Reef formations (Fig. 1.3). The Goat Seep Formation accumulated during the middle Guadalupian age and contains the first reef structures to develop in the Guadalupian time (Crawford, 1979). The Goat Seep outcrops along the Western Escarpment of the Guadalupe Mountains in the form of highly dolomitized limestones, and is equivalent to the Cherry Canyon formation in the underlying basin. Although few fossils can be found within the Goat Seep, some framebuilding and encrusting organisms like *Tubiphytes* and *Archeolithporella*
present within its youngest sections represent species from the colonization stage of reef development that occurred prior to massive aggradation during the time of the Capitan Reef formation (Crawford, 1979).

The massive reef unit of the Capitan Formation was formed during the late Guadalupian time. It is a fine-grained, locally dolomitic limestone that is comparatively more massive and more fossiliferous than the underlying Goat Seep Formation. Reef units are equivalent to shelf strata of the Yates, Seven Rivers, and Tansill Formations. The boundstone lithology is highly cemented, and predominantly non-porous. Faunal populations within the Capitan are diverse, containing fusilinids algae, sponges, coral, crinoids, bryozoans, brachiopods, cephalopods, pelecypods, echinoderms, ammonoids, mollusks, and trilobites. Frame building and encrusting organisms like *Tubiphytes* and *Archeolithoporella* are abundant within this unit, and are considered to be the primary colonizers of the Capitan’s massive reef structures (Hill, 1996)

Figure 1.4: Photomicrograph represents reef colonizing *Tubiphytes* encrusted by *Archeolithoporella*. Modified from Kirkland et al. (1993).
1.3 Forereef Strata

Basinward of the reef, the slope of the Capitan Shelf margin dips at bedding angles between 30-35 degrees and drops 400 vertical meters to the underlying basin. Much of the Capitan Limestone is found upon the weathered slope in the form of talus slopes, turbidites, and debris flows. These deposits have been supplied by carbonate production on the Capitan Reef that was shoaled and shed basinward by marine currents (Kendall, 1969). The upper foreslope consists of bedded to non-bedded intraclastic packstones and rudstones that contain large intraclasts of meter scale diameter (Melim and Scholle, 1995). Lower slope and toe of slope sediments are mostly wackestone, packstones, grainstones, and boundstones that have been deposited by turbidites and debris flows (Brown Loucks 1993B; Melim and Scholle 1995). Fossils found within the lower slope sediments were supplied by “death assemblages” of previously disassociated fossil remains derived from the reef, along with some in situ formation. Deposition of these sediments over the slope contributed to filling of the underlying basin, stimulating lateral basinward progradation of the reef over it’s own debris (Hill, 1996).

1.4. Shelf Strata

Shelf equivalent strata of the Capitan Reef Formation include the Seven Rivers, the Yates, and the Tansill formations. The Seven Rivers conformably underlies the Yates Formation, while the Yates formation conformably underlies the
Tansill formations (Fig. 1.3). All three formations contain a similar array of facies types including: fossiliferous limetones, dolostones, pisolites, and evaporates (Hill, 1996). These formations consist of meter scale siliclastic-carbonate clastic cycles that accentuate periods of reef progradation and aggradation in the stratigraphic record (Longley, 1999). A number of workers have suggested that orbitally forced eustatic sea level changes are the reason for small cycles within the sedimentary record (e.g. Jacka et al., 1985, Meissner, 1985). More recently, studies by Longley (1999) have attributed the origin of these cycles to localized sea level changes on the outer shelf that occurred as a result of differential compaction induced subsidence.

![Figure 1.5: Fall in bed geometry exposed on the northeast wall of Slaughter Canyon. Bedded units represent outer shelf facies of the Seven Rivers and Yates Formations. Basinward splay and steepening of beds is demonstrated from left to right of this cartoon (Longley 1999).](image)

Although differences in lithology are limited among these three formations, facies types within the formations vary dramatically across the shelf. The inner shelf facies, otherwise known as the tidal flat or lagoonal facies extends for approximately 10-13 km across the shelf. This unit consists primarily of dolomitic mudstones with
evaporite nodules, crystal clasts and some interbedded sand layers (Hill, 1996). Further basinward, pisolite and teepee structures form a marginal mound that extends parallel to the reef front (Dunham, 1972). The adjacent outer shelf strata consist of bioclastic packstones, but are recognized mainly for their observable increases in thickness and bedding angle toward the reef. King (1948) and Newell et al. (1953) were some of the first workers to recognize the unique geometry of these beds. Esteban and Pray (1977) gave them the informal name the “fall in beds,” (Fig. 1.5) which has been widely used by workers studying the architecture of the Capitan Shelf margin. Beds increase in thickness from 1-2 meters to about 8-9 meters directly behind the reef and demonstrate progressively increasing basinward dips of 5-15 degrees toward the reef (Saller, 1999). Areas of maximum vertical relief are associated with periods of reef aggradation and have been measured in the Seven Rivers, Yates, and Tansill formations (Tinker, 1998; Longley, 1999). Preservations of maximum relief in the present day are 20-30 meters in the Tansill Formation, and 40-50 meters in both the Yates formation Seven Rivers Formation (Harris and Saller, 1999).

1.5. Basinal Sediments

Basinal sediments associated with the Capitan Reef complex include the Brushy Canyon, Cherry Canyon, and Bell Canyon Formations (Hill, 1996). The Brushy Canyon formation consists of arkosic sands that were deposited prior to the formation of the Capitan Reef complex. The Cherry Canyon and the Bell Canyon Formations overly this unit, and are correlatable to strata of the Goat Seep and
Capitan Formations respectively (Crawford, 1979). Both of these formations are characterized by fine-grained sandstone and coarse-grained siltstones, interbedded with thin limestone tongues that thicken toward the slope (Crow and Bell 2000, Fig. 1.6). Altogether, these units make up the Delaware Mountain group which is present over almost the entire Delaware Basin, reaching a maximum thickness of 1200-1600 meters near its center (Hill, 1996).

![Stratigraphic column demonstrating the manner in which limestone tongues of the Cherry Canyon and Bell Canyon Formations thin to termination in the basin (Crow and Bell 2000).](image)

Figure 1.6. Stratigraphic column demonstrating the manner in which limestone tongues of the Cherry Canyon and Bell Canyon Formations thin to termination in the basin (Crow and Bell 2000).

At the slope, limestone tongues of the Cherry Canyon and Bell Canyon Formations (Fig. 1.6) reach thicknesses of about 80 meters and dip at inclinations between 20-25 degrees. Further down the basin, at a distance of about 8-16 km away
from the slope, these units thin to only a few meters and assume bedding angles of less than 1 degree (Garber et al., 1989). The Cherry Canyon formation is predominantly dolomitic limestone and is 300-400 meters thick below the upper slope of the Capitan Limestone (Hill, 1996). Limestone tongues of the Bell Canyon are predominantly calcitic and merge with the Capitan Limestone toward the reef. Fossils are abundant within the carbonate tongues of the Bell Canyon, but rapidly diminish in the basinward direction (Brown and Loucks, 1993A). Limestone Members of the Bell Canyon Formation include the McCombs, the Rader, the Pinery, the Hegler and the Lamar Limestones (Fig. 1.6). The Lamar Limestone is the only member that is continuous across the basin. Due to the lack of fossils and pronounced basinward thinning of the sandstone units, the other limestone tongues on the basin margin cannot be correlated to deeper basin strata (Hill, 1996).
2. Evidence for Differential Compaction

Prograding high relief margins like the Capitan have the potential to exert high levels of overburden pressure on underlying basin margin sediments and induce significant compaction by way of grain reorientation, dewatering, grain breaking, or pressure solution at grain-to-grain contacts (Hunt et al., 1995). The onset of these processes can vary across carbonate shelf margins as facies with different compaction potentials are frequently juxtaposed in the lateral direction. Evidence for differential compaction on shelf margins is most clearly demonstrated as originally flat lying strata overlying a diverse arrangement of facies types are warped down (Fig. 2.1.) over a more compactable unit (Hunt et al. 1995). Synsedimentary deformation of
tilted strata may contribute to the creation of accommodation space, the modification of original depositional geometries, and the generation of faults and joints (Devaney et al., 1986; Hunt et al. 1995; Resor and Flodin, 2009). A number of workers have attributed the basinward tilting and thickening of outer shelf strata on the Capitan Shelf Margin to the onset of differential compaction induced subsidence (e.g. Saller, 1996; Hardwood and Kendall, 1999; Kosa and Hunt, 2005). In contrast, others have suggested that fall in bed geometry has been formed by the preservation of primary depositional dip (e.g. Hurley 1989; Osleger 1998; Tinker, 1998).

Controversy surrounding the origin of fall in bed geometry has inhibited the development of a unified theory concerning the original height of the water column above the upper platform. Workers who have interpreted the geometry of the outer shelf beds as the preservation of primary depositional dip have suggested that the current distribution of the pisolite facies belt across the upper platform provides an indication of water depth above the reef (e.g. Hurley, 1989; Osleger, 1998; Tinker; 1998). Proponents of differential compaction induced subsidence have contended that facies distribution across the upper platform has been substantially modified by secondary rotation, and that the evolution of the Capitan's bathymetric profile was in fact, much more complex (e.g. Saller; 1996; Hunt et al., 2002). Although the correct interpretation of fall in bed geometry is still hotly debated among workers studying the Capitan depositional system, a number of studies have documented several features on the upper platform that provide strong evidence supporting the model of differential compaction induced subsidence (e.g. Saller, 1996; Longley, 1999; Hardwood and Kendall, 1999; Kosa and Hunt, 2005).
2.1. Stratigraphy

Longely (1999) demonstrated that outer shelf strata on the Capitan Shelf margin are separated into distinct “packages” that shallow and thin up-section (Fig. 2.2). This geometry is unique to profiles influenced by syndepositional compaction, owing to the episodic creation of antecedent topography, and subsequent filling of accommodation space by succeeding deposits (Hunt et al., 1995). Points of maximum subsidence coincide with periods of reef aggradation, and are marked by maximum flood stages in shelf strata (Longley, 1999).

2.2. Geopetal Fabrics

Geopetals, otherwise known as “way up” structures, are distributed throughout the upper platform, and serve to indicate the original orientation (Fig 2.3) of their encasing strata (Hunt et al., 2002). These structures are found most often in
the Capitan Reef system in the form of cavity fills that were filled with horizontally stacked wind blown deposits.

Studies by Saller (1996) and Hardwood and Kendall (1999) employed a three-dimensional field analysis of geopetal fabrics that accurately corrected for apparent dip within outer shelf strata. Geopetals demonstrated a mean dip between 10-10.3
degrees within the Yates Formation and Reef system tracks, similar to the current depositional dip of outer shelf strata. Later studies by Hunt et al. (2002) demonstrated that geopetals and beds of the lower Yates strata dipped at angles of 12 degrees, while beds of the youngest Yates and Tansill Formations dipped at near horizontal bedding angles. Extensive erosion has limited the identification of these features in uppermost strata of the Tansill Formation and given ground to work conducted by Hurley (1989), who measured a number of near-horizontal geopetal fabrics in Yates equivalent strata. Although some have cited the work by Hurley (1989) as evidence for primary depositional dip on the outer shelf, this work has been largely discredited, due to it’s use of poorly oriented outcrop photographs to perform measurements on centimeter scale geopetal fabrics (Hunt et al. 2002).

2.3. Faults

![Figure 2.4: East wall of Slaughter Canyon in the Guadalupe Mountains National Park, showing synsedimentary faults and fractures (Hunt et al. 2005).](image)

Studies conducted by Hunt et al. (2002) and Kosa and Hunt (2005) have observed several faults within the upper platform of the Capitan shelf margin.
(Fig. 2.4), that many workers have attributed to the onset of syndepositional deformation. Faults dip at a high angles parallel to the shelf margin for about 35 kilometers along strike (Hayes, 1964; Hunt et al., 2002). These studies have observed that the growth of faults and fractures on the platform occurs where back reef strata exhibit basinward thickening and steepening. Additionally the predominance and consistently steeper dip of basinward-dipping faults compared to down to shelf faults provides evidence that faults were rotated in the basinward direction with the platform (Kosa and Hunt, 2005). These observations suggest that fault propagation, deposition, and tilting of the outer shelf beds occurred around the same time, as a result of differential compaction induced subsidence (Kosa and Hunt, 2005).

Supporters of the primary depositional dip model have attributed the formation of fault structures to the onset of post-depositional collapse and refuted the influence of secondary rotation as the cause of their formation (e.g. Osleger 1998, Tinker, 1998)

2.4. Need For Future Work

Although the appearance of synsedimentary features provides strong evidence that differential compaction has affected the architecture of the upper platform, very little is known about how severely compaction has affected facies of the underlying basin margin. Unfortunately, traditional methods for quantifying compaction cannot be applied to basin margin deposits of the Capitan depositional system due to the extensive erosion of original burial depth above the lower slope. In order to quantify compaction within basin margin sediments on the Capitan depositional system and
thereby, understand its significance to the development of outer shelf geometries, new methods for quantifying compaction must be developed. This has served as the motivation for the study described in the next portion of this work.
3. Abstract

Prograding carbonate shelf margins are capable of exerting high levels of overburden pressure upon underlying sediments that may induce processes of physical and chemical compaction in facies with different compaction potentials. During the Guadalupian time of the Upper Permian Period, the Capitan shelf margin prograded over compactable wackestones, packstones, and mudstones of the basin margin. A number of workers studying the Capitan depositional system have adopted the model of differential compaction to explain the tilting and thickening of outer shelf strata. Although workers have observed synsedimentary features on the upper platform that have likely formed as a result of differential compaction induced subsidence, little is known about the significance of compaction in the underlying basin margin. The inability of traditional methods to circumvent uncertainties of burial depth and initial porosity has created a high demand for a new method that can be used to quantify compaction within these sediments.

In this study, we develop a new method that can be used to quantify compaction in basin margin facies of the Capitan depositional system. Our methods involve the use of nearest neighbor techniques that quantify compaction based upon the index of grain repackaging. Although our methods are affected by material heterogeneities, they consistently provide a minimum estimate for compaction in fossiliferous limestones that can be used to broaden the understanding of carbonate diagenesis and reef bathymetry. We apply our methods to samples of the Lamar
Limestone Member and demonstrate within this work, that the compaction of basin margin sediments has likely contributed to the subsidence of outer shelf strata on the Capitan Shelf Margin.
4. Methods For Quantifying Compaction

Compaction is defined as the volume reduction that occurs in buried sediment in response to applied pressure (Shinn and Robbin, 1983). Compaction can occur through purely physical processes (e.g. grain rearrangement), or through chemical processes (pressure solution of solid minerals) (Goldhammer, 1997). The possibility that compaction may have influenced the evolution of the Capitan Shelf Margin provides strong incentive to study the process of compaction and the methods by which it can be quantified.

4.1. Physical Compaction:

During physical compaction, fluids are forced out of pore spaces, leading to pore volume reduction and grain repackaging. Physical compaction can produce a number of features within sediment including: plastic deformation of grains, interpenetration of grains, high ratios of linear grain contacts to point contacts, deformed burrows, and the formation of isolated grain embayments (Fig. 4.1) (Budd, 2002). Physical compaction occurs within the early stages of burial and is controlled by the initial porosity of the buried sediment (Goldhammer, 1997).
Highly porous (65-80%) matrix-supported sediments are capable of experiencing significant compaction within the first tens of meters of burial. Grain supported sediments with comparatively lower initial porosities (40-60%), are more resistant to detwatering and tend to compact at a lower rate (Goldhammer, 1997). Experiments performed by Fruth et al., (1966) demonstrated that packstones and grainstones require approximately 4500 meters of burial depth in order to experience volume losses between 50-75%, compared to only a few hundred meters in fine-grained mudstones (Choquette and James, 1986).

Burial compaction can be permanently inhibited in carbonate sediments by the onset of early cementation. Cementation occurs as pore spaces are filled with rigid
aragonite, dolomite, or calcite cements (Choquette and James, 1986). Carbonates can be cemented locally or uniformly, depending upon the conditions of sub-aerial exposure, dolomitization, or microbial activity (Pratt, 1979). The destruction of porosity and the augmentation of bearing strength that occurs as a result of cementation removes the potential for dewatering in initially porous sediment, thereby inhibiting compaction in the host rock (Pratt, 1979; Choquette and James, 1986).

4.2. Chemical Compaction

Chemical compaction, or pressure solution compaction, occurs as grain-to-grain pressure solution leads to volume reduction in lithified sediment or partially lithified rock as a result of overburden pressure (Goldhammer, 1997). Pressure solution (Fig.4.2) occurs as high levels of overburden pressure contribute to the dissolution of solid minerals into aqueous pore fluid (Goldhammer, 1997). Carbonate rocks typically undergo chemical compaction at depths greater than 1000 meters, though Shinn and Robbin (1983) have suggested that given periods of geologic time, chemical compaction can occur at shallower depths. Chemical compaction is responsible for the production of burial cements as mobile ions of aqueous minerals reprecipitate into empty pore spaces (Budd, 2002). Unlike physical compaction, chemical compaction is influenced by pore fluid composition, burial temperature, the presence of clays, and relative mineral solubilities (Goldhammer, 1997). Although organic material within carbonate sediment does not dissolve like solid minerals, such
as aragonite or calcite, experiments conducted by Shinn and Robbin (1983) provide strong evidence that consolidated “seams” of organic material represent precursors to stylolite formation in carbonate sediment.

Stylolites are sutured seams that are commonly produced in carbonate rock by chemical compaction. The heights of stylolite “teeth” represent an interpenetration of grains at discrete solution surfaces within the host rock, and are related to the severity of contractional strain during compaction (Park and Schot, 1968). Like anticracks and compaction bands, stylolites are anti-mode I fractures that propagate perpendicular to the direction of maximum compressive stress. In vertically compacted carbonates, stylolites form parallel to bedding (Fig. 4.3), and can be used to estimate the amount of material dissolved by pressure solution during chemical compaction (Benedicto and Shultz, 2009).

Figure 4.2. Pressure Solution at grain-to-grain contacts and the transport of aqueous minerals to areas of lower pore fluid pressure (Choquette and James, 1986).
When stylolites are smaller than the average grain size of the host rock, their propagation is stimulated mainly by points of textural heterogeneity within the rock material (Park and Schot, 1968). Once the length of a stylolite grows beyond the average grain size, peak development becomes a mainly a function of contractional strain (Ebner et al., 2009).

4.3. Previous Work Quantifying The Effects Of Compaction

Motivated to understand the processes involved in transforming porous carbonate sediment to non-porous rock, a number of workers have developed
methods that can be used to quantify compaction in carbonate sediments. In one of the first studies to pursue this charge, Schmoker and Halley (1982) studied the affects of porosity loss with depth by applying borehole gravity measurements to shallow water grainstones in the South Florida Basin. Through the use of “porosity depth curves” (Fig. 4.4) Schmoker and Halley (1982) demonstrated that porosity levels generally decrease in marine carbonates at an exponential rate, though a number of diagenetic processes may be involved.

Fig 4.4. Porosity depth curve for a number of carbonate facies developed by Schmoker and Halley (1982), and modified by Choquette and James (1986).
Shinn and Robbin (1983) further explored the role of compaction in reducing porosity in carbonate sediment by artificially compacting carbonate sediment cores obtained from the South Florida Bay. Shinn and Robbin (1983) demonstrated that the application of artificial overburden pressure reduced the volume of compactable carbonate muds by as much as 50%, while destroying initial porosity by 35-75%.

Figure 4.5. Decompaction curve for carbonate sand, relating present day thickness and initial thickness through graphical curves (“Z”) that represent varying burial depths. (Goldhammer, 1997).

Using results generated by Shinn and Robbin (1983) and a number of other studies, Goldhammer (1997) expanded the understanding of porosity loss and compaction in his development of “decompaction curves” (Fig.4.5). These curves illustrate the volume reduction that has occurred in a layer of carbonate mud or carbonate sand by relating the initial thickness of the unit to the present day burial depth. Although these
methods have addressed the general relationship between compaction and burial depth, they fail to address the potential for porosity loss by other diagenetic processes (e.g. cementation).

While porosity loss by compaction (COPL) is controlled primarily by burial depth, porosity loss by cementation (CEPL) can occur extensively even at shallow burial depths. By analyzing indices specific to both compaction and cementation in grainstones from west-central Florida, Budd (2002) was able to integrate the relationship of porosity and depth for processes of both compaction and cementation. Budd (2002) demonstrated that sediments affected predominantly by physical and chemical compaction experienced no trend in CEPL indices with depth. Budd (2002) also demonstrated that samples that had been cemented prior to burial were unaffected by processes of mechanical and chemical compaction at greater burial depths. Although point count analysis quantifies the relative influence of compaction and cementation in carbonate sediments, it requires an assumption of initial porosity in order to assess levels of COPL and CEPL. Assuming initial porosity is especially difficult in naturally occurring carbonate rocks that have been substantially altered by diagenesis (Lundegard, 1992). For this reason, accurate quantification of compaction in naturally occurring carbonates may require the use of methods that derive their results from compaction-specific indices, with no prior knowledge of the initial sedimentary fabric.
4.4. Quantifying Compaction As Strain

Compaction is a form of contractional strain where volume is not conserved (Sanderson, 1976). In the absence of tectonic deformation, the accumulation of overburden pressure acts as the primary impetus for the development of strain deformation in sedimentary rocks. The record of strain in carbonate sediment may serve as a better index for compaction than porosity loss because unlike pore volume, it exclusively reflects vertical loading. Quantifying the level of strain in carbonate sediment may be particularly useful for understanding the level of compaction in basin margin sediments of the Capitan Shelf margin, that appear to have experienced little to no tectonic deformation following their deposition.

Ramsay (1967) first developed nearest neighbor methods for quantifying 2-dimensional strain in rock fabric by analyzing the distances \( r \) and orientations \( \theta \) of adjacent grains. When this analysis is applied to anticlustered objects, these statistics can be used as polar coordinates to generate a 2-dimensional plot that forms a central vacancy field around the origin. The axes of the central void define the ratio of a strain ellipse, which describes the severity and orientation of strain in the sample. These methods can be used to quantify the severity of compaction in fossiliferous limestones by analyzing the extent to which grains have been repackaged within the sediment. In this way, nearest neighbor methods may alleviate the problem many studies have encountered when attempting to isolate the effects of compaction from other diagenetic processes (Budd, 2002), and provide a means of exclusively
quantifying compaction based upon the compaction-specific index of grain repackaging.

4.5. Application Of Nearest Neighbor Analysis To Carbonate Sediments

One goal of our study was to develop a method that could be used to quantify compaction in basin margin sediments based solely upon analyzing the index of strain with little potential for interpretative bias. In order to accomplish this goal, we have combined center-to-center methods that improve upon Ramsay’s (1967) method (Erlsev and Ge, 1990; Mulchrone, 2002) with computational ellipse fitting techniques (Gray et al., 2003; Waldron and Wallace, 2007) in a Matlab® script (Appendix A). This script can be used to generate a polar plot from raw (x,y) coordinate data of grain-to-grain centroids, and fit a strain ellipse to the resulting central void that can be interpreted to estimate compaction.

Figure 4.6. DTNNM applied to a computer-generated sample. A) object centroid. B) Line connecting nearest neighbor C) Points on the convex hull. Figure modified from Waldron and Wallace (2007).
The delaunay triangulation method improves upon Ramsay’s (1967) method by efficiently identifying nearest neighbors. The delaunay triangulation nearest neighbor method (DTNNM) accomplishes this by superimposing a voronoi diagram over a set of object centroids and generating lines that connect points in adjacent regions (Fig. 4.6). DTNNM derives statistics for $r$ and $\theta$ from the resulting lines and following Ramsay’s (1967) method generates a polar plot. Points on the edge, or “convex hull,” of the sample are removed when performing this analysis in order to reduce edge effects.

![Figure 4.7: Application of Normalization Techniques to the polar plot](image)

Figure 4.7: Application of Normalization Techniques to the polar plot A) non-normalized polar plot B) normalized polar plot. Figure modified from Erslev and Ge (1990).

Normalization techniques developed by Erselv and Ge (1990) alleviate “clouding” of the central void that occurs in poorly sorted samples due to variations in grain radii (Fig 4.7). Normalization increases the resolution of the central void by normalizing the distances between touching objects to variations in grain size. This technique creates a rim of maximum point density around the rim of the central void.
that is defined by statistics for touching grains (Fig. 4.7). The resulting rim can easily be interpreted even in samples containing packed aggregates.

A strain ellipse may be fit to the resulting void and rim using the continuous function method presented by Waldron and Wallace (2007). This method seeks to minimize the distance \( x \) between the radius of a trial ellipse \( r_e \) and the distance of the nearest neighbor points \( r \) for the orientation \( \theta \) of each point (Eq. i).

\[
x = (r - r_e)/r_e = r/r_e - 1
\]

\textbf{Eq. i}

Waldron and Wallace chose an arbitrary function (eq. ii) to weight the resulting ellipse fit values \( x \). The function gives high weight to the edge of the void and a negative weight to points within the central void (Fig. 4.8) (Eq. ii). This equation thus places a high weight on points represented by touching neighbors or nearly touching neighbors within the plot (Fig. 4.8).

\[
z = \left( \frac{x}{Fr} \right) e^{-\left(kx\right)^2}
\]

\textbf{Eq. ii}

Figure 4.8. Plot of the uncorrected weight function, \( y = x \exp\left(-(kx)^2\right) \), from which Eq. ii is derived (Waldron and Wallace, 2007).

The constant \( k \) is arbitrary based upon the desired sensitivity of the ellipse to the outer part of the polar plot. High values of \( K \) tend to exclude more points on the outer portion of the polar plot, while low values will make \( Z \) more sensitive to those points.
Waldron and Wallace (2007) found a value of 3 to be generally effective in their application of the continuous function ellipse to polar plots.

Rather than relying solely upon the polar coordinates of touching or nearly touching grains, the covariance method developed by Gray et al. (2003) analyzes the distances ($r$) and orientations ($\theta$) of all neighbors within the sample to estimate the ratio of the strain ellipse. These statistics are used to calculate a covariance matrix (Eq. iii).

$$M = \frac{1}{n-1} \sum_{\beta=1}^{n} d_{\beta}d_{\beta}'$$  \hspace{1cm} \text{Eq. iii}

The eigenvalues of $M$ are an approximation of the principal stretches of the elliptical axes, while the eigenvectors represent their orientations.

Within our script (Appendix A), we used DTNNM to generate a polar plot from statistics of ($r$) and ($\theta$) between adjacent matrix-supported grains. We combined DTNNM with the continuous function method (DTCF) in order to quantify compaction by fitting a strain ellipse to the boundary of the central void. In DTCF, the trial ellipse is defined using a perfect circle with a radius of $1.0=\tau$ on the polar plot. We have also devoted a portion of our script to combining DTNNM with the covariance method (DTCov), which estimates a strain ellipse by analyzing the position of all grains within the sample.
4.6. Evaluation Of Our Methods

We tested the efficacy of our script in replicating nearest neighbor and ellipse fitting methods by applying them to computer-generated samples obtained from Waldron and Wallace (2007). These samples include circular shaped groups of artificial “grains” (Fig. 4.9.) that were deformed to varying levels of strain and sorted at 0.4 phi units.

Figure 4.9: Computer-generated samples obtained from Waldron and Wallace (2007). A) Strain ratio of 1.3. B) Strain ratio of 1.8. C) Strain ratio of 2.5

Levels of strain in these samples can best be described by strain ratios of 1.3:1, 1.8:1, and 2.5:1, which translate to 23%, 44%, and 60% compaction respectively. High levels of anticlustering and homogenous strain make these samples...
ideal for testing the accuracy of our script in replicating nearest neighbor methods and ellipse fitting techniques (Fry, 1979; Waldron and Wallace, 2007).

The next step was to evaluate the capabilities of our method in quantifying compaction in carbonate sediment. Understanding how DTCF and DTCov quantify compaction in samples of carbonate sediment can provide an analogue for interpreting results obtained from our analysis of compaction in basin margin sediments from the Capitan Shelf Margin. Our first test was designed to estimate compaction in granular carbonate wackestones that were artificially compacted to known levels of strain by Shinn and Robbin (1983), images provided by Shinn (pers. comm. September, 2009). These samples (Appendix B) contain an array of distinguishable grains that can be used to generate raw data for analysis by DTCF and DTCov. Although no level of compaction was provided for the “uncompacted” Rodriguez Key Sample, we assume it represents levels of compaction near 0%. The compacted Biscayne Sample and Rodriguez Key Samples have been compacted to levels of 72% and 60% respectively (Shinn and Robbin, 1983).

Within our tests, we also devoted a portion of our script (Appendix A) to analyzing the statistical and spatial distribution of compaction in carbonate sediment. We randomly selected one hundred grain centroids from each sediment sample (Fig. 4.10), and generated a set of circular sub-samples, using the selected centroid as their centers. We chose to use one hundred sub-samples for each individual sample in the interest of finding an appropriate balance between sample-size and computing cost (Fry, 1979). We estimated values for compaction within each sub-set using DTCF
and DTCov methods, and then calculated a statistical mean and standard deviation for the compaction values of every sub-sample.

Figure 4.10. Random generation of object centroids that were chosen as sub-sample centers for statistical analysis. Red dots represent the centers of sub-samples, green “+” symbols represent surrounding centroids.

4.7. Preparation Of Samples

In order to prepare samples into raw data for analysis, we carried out a number of processing steps in Adobe Photoshop® and ESRI ArcMap® (Appendix C). First, we scanned previously published samples obtained from Waldron and Wallace (2007) and Shinn (pers. comm. September, 2009) and converted them into 2-dimensional images. We then imported them into Adobe Photoshop® and converted them into black and white images. We assigned white color to grains and black color to the surrounding matrix. Next, we used “raster tools” in ESRI ArcMap® to improve the size and location of grains at the pixel scale and remove any extraneous noise from the image. We then “vectorized” these grains into polygons that could be analyzed as raw (x,y) coordinate data.
We determined the location of each grains centroid, using “Feature To Point” tools in ESRI ArcMap® that computationally define the centers of any arbitrary shape, inside or outside the grain itself. Because the accuracy of the central void in center-to-center methods is directly related the degree of anticlustering (Ramsay, 1967), we analyzed the level of clustering for grain centroids of each sample using “Average Nearest Neighbor” tools in ArcMap®. Our analysis determined that only the Uncompacted Rodriguez Key Sample demonstrated clustering.

![Figure 4.11](image.png)

Figure 4.11. Output generated from “Average Nearest Neighbor” in Spatial Statistics Tools form Arc Toolbox (ESRI ArcMap®). The top window shows the quality of clustering within the sample. The bottom window displays a z-score and p value for the distribution pattern.

We induced anticlustering in this sample by gradually increasing a grain threshold in single pixel increments. This process was conducted meticulously to provide for maximum resolution of the central void while minimizing the number of grains deleted from the original sample. We continued this process until the “Average
Nearest Neighbor” spatial statistics tool qualified a high level of anticlustering at a threshold of 0.07 mm$^2$. In order to determine how clustering and the adjustment of grain-size thresholds affected our results for the uncompacted sample, we decided to include both clustered and anticlustered samples in our analysis. Lastly, we exported the (x,y) coordinates for the centroids of each sample into Matlab® to be analyzed by DTCF, DTCov, and statistical analysis methods.
Table 1: Results from methodology tests on computer-generated samples from Waldron and Wallace (2007).

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Predicted Strain Ratio</th>
<th>Estimated Compaction From Strain Ratio (%)</th>
<th>Waldron and Wallace (2007) Calculation For Strain Ratio</th>
<th>DTCF Results</th>
<th>DTCov Results</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Strain Ratio</td>
<td>Estimated Compaction From Strain Ratio (%)</td>
</tr>
<tr>
<td>Sample A</td>
<td>1.30:1</td>
<td>-23</td>
<td>1.25:1</td>
<td>1.25:1</td>
<td>19.92</td>
</tr>
<tr>
<td>Sample B</td>
<td>1.80:1</td>
<td>-44</td>
<td>1.74:1</td>
<td>1.56:1</td>
<td>44.43</td>
</tr>
<tr>
<td>Sample C</td>
<td>2.50:1</td>
<td>-60</td>
<td>2.36:1</td>
<td>2.40:1</td>
<td>58.38</td>
</tr>
</tbody>
</table>
Figure 4.12. DTCF Strain ellipses generated from methodology tests on computer-generated samples from Waldron and Wallace (2007). Red dots represent polar coordinates, ellipses are slightly underfit to the central void. A) Results for sample with a strain ratio of 1.3. B) Results for the sample with a strain ratio of 1.8. C) Results for the sample with a strain ratio of 2.5.
<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Predicted Compaction (%)</th>
<th>Estimated Compaction From DTCF (%)</th>
<th>Deviation From Predicted Compaction (%)</th>
<th>Estimated Compaction From Statistical Analysis (%)</th>
<th>Deviation From Predicted Compaction (%)</th>
<th>Standard Deviation For Statistics</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biscayne Compacted</td>
<td>-72.0</td>
<td>0.7</td>
<td>-71.3</td>
<td>36.24</td>
<td>-35.76</td>
<td>36.71</td>
</tr>
<tr>
<td>Rodriguez Compacted</td>
<td>-60.0</td>
<td>3.75</td>
<td>-56.25</td>
<td>31.33</td>
<td>-28.67</td>
<td>35.09</td>
</tr>
<tr>
<td>Rodriguez Uncompacted Clustered</td>
<td>0</td>
<td>3.75</td>
<td>12.77</td>
<td>9.14</td>
<td>18.47</td>
<td>18.47</td>
</tr>
<tr>
<td>Rodriguez Uncompacted Anticluustered</td>
<td>0</td>
<td>7.1</td>
<td>7.1</td>
<td>6.7</td>
<td>6.7</td>
<td>19.18</td>
</tr>
<tr>
<td>Difference</td>
<td>0</td>
<td>-5.67</td>
<td>-5.67</td>
<td>-2.44</td>
<td>-2.44</td>
<td>0.71</td>
</tr>
</tbody>
</table>

Table 2: Results for methodology tests on carbonate sediment samples
Figure 4.13: DTCF and statistical analysis results for uncompacted Rodriguez Key samples. A) Results for the clustered sample. Left, continuous function strain ellipse fit slightly outside the central void. Red dots represent polar coordinates. Right, statistical distribution of compaction values in randomly generated sub-samples. B) Results for the anticlustered sample. Left, red dots represent polar coordinates. Continuous function strain ellipse fit slightly outside the central void. Right, statistical distribution of compaction values in randomly generated sub-samples. Applying a grain threshold increased the resolution of the central void.
Figure 4.14. DTCF and Statistical analysis results for compacted Biscayne Sample. Top, polar plot with continuous function strain ellipse. Red dots represent polar coordinates. The strain ellipse appears to surround the central void. Bottom, statistical distribution of compaction values in randomly generated sub-samples.
Figure 4.15. DTCF and statistical analysis results for compacted Rodriguez Key sample. Top, polar plot with continuous function strain ellipse. Red dots represent polar coordinates. The strain ellipse is fit outside the central void. Bottom, statistical distribution of compaction values in randomly generated sub-samples.
Figure 4.16. Spatial distribution of compaction values in uncompacted Rodriguez Key sample. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated subsample. A) The clustered sample. B) The anticlustered sample.
Figure 4.17. Spatial Distribution of compaction values in the compacted Biscayne sample. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. The yellow arrow points in the direction that pressure was applied.
Figure 4.18. Spatial Distribution of compaction values in the compacted Rodriguez Key sample. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. The yellow arrow points in the direction that pressure was applied.
4.8. Results From Methodology Tests

Results from our tests on computer-generated samples reveal that DTCF consistently produced good estimates for strain, while DTCov consistently underestimated strain by a substantial amount (Table 1). Both methods underfit the strain ellipse to the central void, though results for DTCF (Fig. 4.12) were all within 6% of known values and consistently better than DTCov. DTCov underestimated known levels of strain by a range ~14-54%, and produced similar estimates for Sample B and Sample C, despite substantial differences in strain. The accuracy of DTCF appeared to increase with strain, with the most accurate estimates for Sample C. In comparison of results to Waldron and Wallace (2007), our approach generated results that were similar for sample (A), less accurate for sample (B), and more accurate for sample (C).

DTCF, DTCov, and statistical analysis all tended to underestimate the level of compaction when applied to carbonate sediments, though deviations from predicted values were much more substantial compared to results for computer-generated samples. DTCF generated good estimates for compaction in the uncompacted sample (Fig.4.13), deviating from the presumed value of 0% compaction by 9% (Table 2). A grain area threshold (0.07 mm²) slightly increased the accuracy of DTCF, decreasing the estimated value from ~9% to ~6%. DTCF results for compacted Biscayne (Fig. 4.14) and compacted Rodriguez Key (Fig. 4.15) samples consistently underestimated the known level of compaction. DTCF underestimated compaction in compacted samples by ~30% but unlike DTCov, successfully generated higher values for more severely compacted samples.
Calculating the statistical mean among sample sub-sets generated results similar to those produced by DTCF (Table 2).

![Figure 4.19: Compaction Values associated with grain shape. A) High values near packages of flat grains. B) Lower values controlled by the presence of larger grains.](image)

Values for compaction in both uncompacted samples (Fig. 4.13) are uniformly distributed in a relatively small range of values, while in compacted samples (Fig.4.14, Fig. 4.15) the distribution of values is highly variable, including multiple peaks. Although the range of compaction values in our compacted samples is broad, standard deviations are consistently below 0.2 in all samples. High values for compaction appear to be associated with dense packages of flat grains (Fig. 4.19A), while areas with large, or less densely packaged grains appear to control low compaction values (Fig.4.19B). These observations are consistent with the problem of using the central void to estimate the strain ellipse. The black box within the compacted Biscayne sample does not appear to substantially affect the spatial distribution of compaction (Fig. 4.17). Both low and high values are positioned on all sides of the box. Crossing into the box does not appear to affect compaction
values in circular sub-samples, aside from those that appear to be influenced the shape and density of grain packing.
Figure 4.20. Map of study area at the mouth of McKittrick Canyon, in the Guadalupe Mountains National Park, Texas. Stippled line represents projected orientation of reef front.
Figure 4.21. Aerial view of transects 1 and 3 on the northern cliff at the mouth of McKittrick Canyon, Guadalupe Mountains National Park, Texas. Locations of GH005, GH023, and GH026 may be affected by horizontal or vertical error.
Transect 2

<table>
<thead>
<tr>
<th>Location</th>
<th>Height</th>
<th>Strike</th>
<th>Dip</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH010</td>
<td>1600.448794</td>
<td>10</td>
<td>8</td>
</tr>
<tr>
<td>GH012</td>
<td>1614.523117</td>
<td>336</td>
<td>19</td>
</tr>
<tr>
<td>GH014</td>
<td>1629.356640</td>
<td>2</td>
<td>19</td>
</tr>
<tr>
<td>GH015</td>
<td>1636.675412</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>GH016</td>
<td>1640.550289</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>GH017</td>
<td>1646.864559</td>
<td>25</td>
<td>16</td>
</tr>
</tbody>
</table>

Figure 4.21. Aerial view of transect 2 on the northern cliff at the Mouth of McKittrick Canyon, Guadalupe Mountains National Park, Texas. The location of GH017 may be affected by vertical or horizontal error.
4.9. Field Methods

The ultimate goal of our study was to use DTCF and DTcov methods to quantify the level of compaction within samples of basin margin sediment from the Capitan depositional system, and thereby assess whether or not compaction in these sediments has played a significant role in the development of fall in bed geometry. At the mouth of McKittrick Canyon in the Guadalupe Mountains National Park (Fig. 4.20), we identified the distribution and relative position of facies types of the Lamar Limestone Member and collected samples from a variety of different facies types for analysis.

We made our interpretations of stratigraphy within the Lamar Limestone along three transects. We determined the locations for our transects based upon two criteria:

1) Accessibility
2) Exposure of sedimentology and stratigraphy

All transects spanned approximately the entire section of the Lamar Member (Appendix D) from the Bell Canyon sandstone to the uppermost identifiable Lamar beds. Transects 1 and 3 (Fig. 4.21) are positioned on the southern cliff of McKittrick Canyon, while Transect 2 (Fig. 4.22) is positioned about 0.5 kilometers north, on the northern cliff. Transect 1 is positioned at the lowest elevation, between 1548-1620 meters above sea level, and stretches the furthest into the basin, perpendicular to the
shelf margin. Transect 2 is ~300 meters closer to the shelf margin than Transect 1, lying between 1600 and 1646 meters above sea level, parallel to the reef front. Transect 3 also extends parallel to the shelf margin, and spans an elevation of 1600-1646 meters above sea level. Facies types generally transition from silty limestones, to coarser packstones, and back to bioturbated or burrowed wackestones. We did not observe a burrowed wackestone facies in Transect 3, although we did sample a silty-breccia facies, which we identified as the top Lamar beds based upon descriptions by Brown and Loucks (1993B).

For each measured section (Fig. 24) we identified facies using Dunham’s (1964) classification for carbonate sediments. We measured the vertical extent for each stratigraphic layer using a tape measurer, and a field compass to correct for angle of inclination. We also made strike and dip angle measurements on each bedding plane using a field compass. In order to provide for sub-meter spatial orientation, we used a handheld differential GPS unit to mark the location of the sections we walked. We also collected samples for a number of facies types along each transect (Fig. 4.21, Fig. 4.22). We processed four of these samples (Appendix E) for analysis by DTCF and DTCoV methods including:

4.10. Sample Description:

GH005: GH005 represents an intraclast packstone obtained from the southern cliff on Transect 1. It is the lowest sample within the Lamar Limestone, positioned at an elevation of ~1588 meters above sea level. Very few touching grains are observed in
thin section, aside from a few point contacts at isolated locations. Intraclasts are comprised of blocky or elliptical algal and calcareous sponge fossils that are poorly sorted and between 2-4 millimeters in length.

**GH010**: GH010 is a skeletal packstone obtained from the bottom of Transect 2 on the northern cliff of McKittrick Canyon. It is positioned about ten meters above the Bell Canyon sandstone at 1600 meters above sea level. Linear grain contacts are present in isolated locations within the sample, and are most noticeable on the left and right sides. Matrix-suspended grains are flat and relatively well sorted at sub-millimeter scale.

**GH014**: GH014 is positioned near the middle of the Lamar Member on the northern cliff at an elevation of ~1628 meters above sea level. This sample was obtained from an outcrop of packstone with algal boundstone intraclasts that are all sub-millimeter in size. Fossils and shell fragments are abundant within this sample in the form of mollusks, platy algae, shell fragments, and fusilinids. Linear grain contacts are visible within this sample in isolated packages, and all grains appear to be oriented horizontal with respect to burial.

**GHTop**: GHTop represents a silty megabreccia with ~mm size intraclasts obtained from the top of the Lamar Limestone Member. Out of all our samples GHTop is the coarsest, and is positioned the highest within the Lamar Member at an elevation of
1646 meters above sea level. While point contacts between grains are abundant within this sample, there are no visible linear grain contacts.

4.11. Processing Of Lamar Limestone Samples:

Before analyzing our samples using DTCF and DTCoV, we processed them into thin sections and photographed them using a Leica® high-speed digital color camera, at a magnification of 3.56x. We then processed thin section photographs into raw coordinate data using the same image adjustment procedures in Adobe Photoshop® and ESRI ArcMap® that we used to prepare previously published samples (Appendix C). Next, we repeated methods for determining the level of clustering in our samples, and applied grain thresholds as needed (Table 3). Finally, we exported the centroid data for each our samples to Matlab® for analysis by DTCF, DTCoV, and statistical analysis methods.

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Grain size threshold (mm²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH010</td>
<td>0.002</td>
</tr>
<tr>
<td>GH014</td>
<td>0.004</td>
</tr>
<tr>
<td>GHTop</td>
<td>0.15</td>
</tr>
</tbody>
</table>

Table 3: Grain threshold for initially clustered samples.
4.12. Stylolite Density Measurements

To qualify our analysis of compaction in basin margin sediments of the Capitan depositional system, we collected bed-perpendicular scanlines of stylolite density in facies of the Lamar Limestone. Our methods are similar to those that have been described by Benedicto and Shultz (2009), and involve estimating the amount of material lost along a stylolite surface by measuring the tooth height and stylolite density within an outcrop.

Figure 4.23. Stylolite scanline methods. A) Scanlines are designated by draping a tape measurer on a vertical face. B) The average amplitude of stylolite teeth is recorded in order to calculate the amount of volume loss along the stylolite surface and the total stylolite density for the facies. Figure 4.23B modified from Benedicto and Schultz (2009).
We extended a tape measurer across a number of outcrops in order to create a vertical scan line of stylolite density for a diverse selection of litho-facies (Fig. 4.23). The outcrops we chose for stylolite measurements displayed a continuously exposed vertical height greater than 0.7 meters, allowing for an accurate estimate of stylolite density. Along transects 1, 2, and 3, stylolite scanlines measured stylolite density for ~90%, ~24%, and ~30% of the Lamar Limestone Member respectively. We documented the location of each stylolite along the scanline and the average amplitude of stylolite teeth within each identifiable solution surface. Using this data, we calculated the contribution of stylolite-compaction to volume loss within each facies outcrop (Table 4).
Figure 4.24. Measured sections for outcrops of the Lamar Limestones Member created from transects. Samples A) Transect 1, B) Transect 2, C) Transect 3,
Table 4: Stylolite Density and Compaction In Lamar Limestone Facies

### Transect 1

<table>
<thead>
<tr>
<th>Location Name</th>
<th>Description</th>
<th>Average Tooth Height (cm)</th>
<th>Present Thickness (m)</th>
<th>Stylolites Per meter</th>
<th>Material Lost At Stylolite (m)</th>
<th>Stylolite Compaction (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH001</td>
<td>Sandstone</td>
<td>0.23</td>
<td>2.05</td>
<td>4</td>
<td>0.018</td>
<td>0.8</td>
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<tr>
<td>GH002</td>
<td>Silty Packstone</td>
<td>0.23</td>
<td>2.26</td>
<td>5</td>
<td>0.023</td>
<td>1</td>
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<tr>
<td>GH003</td>
<td>Silty Wackestone</td>
<td>0.06</td>
<td>1.36</td>
<td>6</td>
<td>0.035</td>
<td>2.50</td>
</tr>
<tr>
<td>GH004</td>
<td>Wackestone</td>
<td>0.53</td>
<td>3.24</td>
<td>8</td>
<td>0.126</td>
<td>3.70</td>
</tr>
<tr>
<td>GH005</td>
<td>Intralast Packstone</td>
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<td>5</td>
<td>0.052</td>
<td>1.38</td>
</tr>
<tr>
<td>GH007</td>
<td>Burrowed Wackestone</td>
<td>0.59</td>
<td>3.07</td>
<td>15</td>
<td>0.256</td>
<td>7.8</td>
</tr>
</tbody>
</table>

### Table 4A) Transect 1 stylolite density measurements.

### Transect 2

<table>
<thead>
<tr>
<th>Location Name</th>
<th>Description</th>
<th>Average Tooth Height (cm)</th>
<th>Present Thickness (m)</th>
<th>Stylolites Per meter</th>
<th>Material Lost At Stylolite (m)</th>
<th>Stylolite Compaction (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH010</td>
<td>Thin Wackestone</td>
<td>0.36</td>
<td>8.6</td>
<td>3</td>
<td>0.084</td>
<td>1.2</td>
</tr>
<tr>
<td>GH012</td>
<td>Thick wackestone</td>
<td>0.29</td>
<td>3.56</td>
<td>7</td>
<td>0.049</td>
<td>1.55</td>
</tr>
<tr>
<td>GH014</td>
<td>Skeletal Packstone</td>
<td>0.7</td>
<td>8.4</td>
<td>8</td>
<td>0.167</td>
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<tr>
<td>GH015</td>
<td>Wavy Wackestone</td>
<td>0.29</td>
<td>1.04</td>
<td>14</td>
<td>0.04</td>
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<tr>
<td>GH016</td>
<td>Thick Packstone</td>
<td>0.31</td>
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<td>GH017</td>
<td>Burrowed Wackestone</td>
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<td>1.64</td>
<td>13</td>
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</table>

### Table 4b) Transect 2 stylolite density measurements.

### Transect 3

<table>
<thead>
<tr>
<th>Location Name</th>
<th>Description</th>
<th>Average Tooth Height (cm)</th>
<th>Present Thickness (m)</th>
<th>Stylolites Per meter</th>
<th>Material Lost At Stylolite (m)</th>
<th>Stylolite Compaction (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH023</td>
<td>Silty Limestone</td>
<td>0.41</td>
<td>9.39</td>
<td>7</td>
<td>0.264</td>
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<td>GH024</td>
<td>Thin Skeletal Wackestone</td>
<td>0.36</td>
<td>5.3</td>
<td>6</td>
<td>0.111</td>
<td>2</td>
</tr>
<tr>
<td>GHTop</td>
<td>Sny Megabrecia</td>
<td>0.3</td>
<td>3.34</td>
<td>8</td>
<td>0.077</td>
<td>2.2</td>
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### Table 4c) Transect 3 stylolite density measurements
<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Facies Type</th>
<th>Transect Number</th>
<th>Styolite Compaction (%)</th>
<th>DTCov Compaction (%)</th>
<th>DTCF Compaction (%)</th>
<th>Statistical Mean Compaction (%)</th>
<th>DTCF and Stylolite Compaction (%)</th>
<th>Standard Deviation From Statistical Mean</th>
<th>Projected Height To Shelf Margin (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH005</td>
<td>Intraclast Packstone</td>
<td>1</td>
<td>1.38</td>
<td>3.88</td>
<td>4.55</td>
<td>10.94</td>
<td>5.93</td>
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<td>545</td>
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<td>GH010 Clustered</td>
<td>Skeletal Wackstone/ Packstone</td>
<td>2</td>
<td>1</td>
<td>3.2</td>
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<td>13.71</td>
<td>11.84</td>
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<td>Skeletal Wackstone/ Packstone</td>
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<td>18.63</td>
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</tr>
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<td>Intraclast Packstone</td>
<td>2</td>
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<td>19.26</td>
<td>10.32</td>
<td>10.46</td>
<td>13.32</td>
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<td>3</td>
<td>21.73</td>
<td>13.14</td>
<td>17.42</td>
<td>16.14</td>
<td>0.0932</td>
<td>504</td>
</tr>
<tr>
<td>GHTop Clustered</td>
<td>Silty Megabreccia</td>
<td>3</td>
<td>2.2</td>
<td>18.93</td>
<td>2.66</td>
<td>19.11</td>
<td>4.86</td>
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<td>489.6</td>
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<tr>
<td>GHTop Anticlustered</td>
<td>Silty Megabreccia</td>
<td>3</td>
<td>2.2</td>
<td>13.26</td>
<td>5.34</td>
<td>-</td>
<td>7.54</td>
<td>-</td>
<td>489.6</td>
</tr>
</tbody>
</table>

Table 5: All results generated from Lamar Limestone samples. The "-" symbol indicates that analysis was not performed. Projected height to present day shelf margin is an approximation obtained from Brown and Loucks (1993B).
Figure 5.1. GH005, DTCF and statistical analysis results. Red points represent polar coordinates. A) Polar plot with continuous function strain ellipse (4.55% compaction) B) Statistical distribution of sub-sample compaction values.
Figure 5.2. GH005, Spatial distribution of compaction values. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. Updip is marked by the corner on the upper right side of the sample. The arrow points down-section from the top of the bedding plane.
Figure 5.3. GH010, DTCF results and statistical analysis results for clustered and anticlustered samples. Red points represent polar coordinates. A) Results for clustered sample. Left, continuous function strain ellipse (10.84% compaction). Right, statistical distribution of sub-sample compaction values. B) Results for anticlustered sample. Left, continuous function strain ellipse (15.38% compaction). Right, statistical distribution of sub-sample compaction values.
Figure 5.4. Spatial distribution of compaction values in GH010. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. Updip is marked by the corner on the upper right side of the sample. The arrow points down-section from the top of the bedding plane. A) The clustered sample. B) The anticlustered sample.
Figure 5.5. GH014, DTCF and statistical analysis results for clustered and anticlustered samples. Red dots represent polar coordinates. A) The clustered sample results. Left, polar plot with continuous function strain ellipse (10.32% compaction). Right, the statistical distribution of sub-sample compaction values. B) The anticlustered sample results. Left, strain ellipse (13.14% compaction). Right, statistical distribution of sub-sample compaction values.
Figure 5.6. Spatial distribution of compaction values in GH014. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. Updip is marked by the corner on the upper left side of the sample. The arrow points down-section from the top of the bedding plane. A) The clustered sample. B) The anticlustered sample.
Figure 5.7. GHTop, DTCF results and statistical analysis for clustered and anticlustered samples. Red dots represent polar coordinates. A) The clustered sample results. Left, continuous function strain ellipse (2.66\% compaction). Right, the statistical distribution of sub-sample compaction values. B) The anticlustered sample results. Left, Continuous function strain ellipse (5.34\% compaction).
Figure 5.8. Spatial distribution of compaction values in the clustered sample of GHTop. Circles represent the center of sub-samples that we generated during statistical analysis. The color of each circle represents the quantity of compaction within each associated sub-sample. This sample was not oriented due to it’s massive appearance in the field.
5. Results For Samples Of The Lamar Limestone

Statistical analysis yielded lower standard deviations in samples from the Lamar Limestone (Table 5) compared to artificially compacted sediment samples (Table 2) obtained from Shinn (2010). Like the results generated during our methodology tests, DTCF and DTCov produced inconsistent results for nearly all of our Lamar Limestone samples (Table 5). DTCov produced values that were both higher and lower than DTCF results, demonstrating no clear pattern relative to DTCF. The statistical mean rarely produced similar values to DTCov, with the exception of the clustered GHTop sample (Table 5). Like we observed in samples obtained from Shinn (2010), spatial analysis of compaction values demonstrates no clear vertical or lateral trend in any sample, aside from those that are induced by the density of grain packaging.

GH005 represents an intraclast packstone with millimeter-sized grains and was the only sample for which DTCF and DTCov produced relatively consistent results. DTCF estimated a compaction of ~4.5% (Fig. 5.1), while DTCov estimated a compaction value of 3.88% (Table 5). Statistical analysis demonstrates that compaction values are uniformly distributed around a single peak between 5-15% compaction (Fig. 5.1), with a standard deviation of 0.06 (Table 5). Sub-samples containing millimeter sized grains are commonly associated with low compaction values (Fig 5.2).

GH010 is a finer grained packstone facies than GH005, and consists primarily of sub-millimeter sized grains. DTCF estimated ~11% compaction in the clustered
sample (Fig. 5.3). Applying a grain threshold increased DTCF estimates for compaction to ~15% (Table 5), and shifted the angle of the polar plot ~32 degrees oblique to the central axis (Fig. 5.3). DTCov estimated ~3% compaction in both the clustered and anticlustered samples (Table 5). Statistical analysis generated a statistical mean of ~14% in the clustered sample and ~19% in the anticlustered sample (Table 5). Standard deviations were below 0.1 for both clustered and anticlustered samples. Both the anticlustered and the clustered sample display variable statistical distributions, though more peaks are observed in the anticlustered sample (Fig 5.3). High compaction values appear to coincide with areas where flat grains are packed into tight groups. The lowest compaction values are positioned around larger, less tightly packed grains (Fig. 5.4).

GH014 is similar in facies type to GH010, and produced similar results from DTCF analysis. DTCF estimated ~10% compaction in the clustered sample and ~13% compaction in the anticlustered sample (Fig. 5.5). DTCov methods estimated larger values of ~19% and 22% in both clustered and anticlustered samples respectively (Table 5). Statistical analysis estimated a statistical mean of ~10.5% compaction in the clustered sample and ~17.5% in the anticlustered sample (Table 5). Standard deviations are less than 0.1 for both clustered and anticlustered samples. Both samples also demonstrate variable statistical distributions with multiple peaks (Fig. 5.5). High compaction values are associated with linear grain contacts, while low compaction values appear to be associated with round grains (Fig. 5.6).

GHTop represents a silty megabreccia with ~mm size intraclasts obtained from the top of the Lamar Limestone Member. Out of all our samples GHTop is the
coarsest, and is positioned highest within the Lamar. DTCF produced low estimates for compaction in both clustered and anticlustered samples (Fig. 5.7), generating results of ~3% and 5% compaction respectively (Table 5). DTCov estimated ~19% compaction in the clustered sample (Table 5). Applying a grain threshold decreased the DTCov estimate to ~13% compaction (Table 5). Statistical analysis estimated ~19% compaction in the clustered sample. Applying a grain threshold inhibited the application of statistical analysis in this sample. High compaction values in the clustered sample (Fig. 5.8) are primarily associated with grain centroids that were removed by applying a grain threshold (Appendix E). This suggests that the statistical mean for compaction in the clustered sample may be affected by “noise” in the polar plot.

5.1. Results For Stylolite Compaction Analysis

Measurements for stylolite compaction throughout the Lamar Member (Table 4) appear to be controlled by facies type. All of the samples analyzed by center-to-center methods represent granular packstone to breccia facies that tend to display levels of stylolite compaction between 1-3%. Wackestone facies on the Lamar display values for stylolite compaction between 2.5-8% that are in some cases, comparable to levels of intergranular compaction estimated by DTCF and DTCov methods (Table 5). Stylolite density generally increases up section and appears to be more influential on results for stylolite compaction than average tooth height. We observed the lowest levels of stylolite compaction on the bottom of Transect 1 in
samples GH001, GH002, GH003, GH004, and GH005. We observed the highest values of stylolite compaction in burrowed and bioturbated wackestone facies, which at times exceeded estimated values for intergranular compaction recorded for packstone samples of the Lamar Limestone (e.g. GH005, GHTop).
6. Discussion

6.1. Interpretations Of Methodology

Although DTCF produced relatively accurate estimates for strain in computer-generated samples, both DTCov and DTCF consistently underfit the strain ellipse to the central void (Fig. 4.12). Underfitting of the strain ellipse may be related to the weight of $\theta$ statistics within the polar plot. By assigning an equal weight to all $\theta$ between nearest neighbors, DTNNM places greater emphasis on fitting the short axis of the ellipse than the long axis, leading to an underestimation of the central void’s elliptical ratio. Although this only slightly affected the accuracy of DTCF within the computer-generated samples, it may pose a problem to future studies attempting to quantify 2-dimensional strain in samples that are more clustered and heterogeneously strained.

The greater accuracy of the DTCF method compared to the DTCov method in calculating strain can be attributed to the manner in which statistics from Delaunay Triangulation are used to calculate the strain ratio. In a polar plot, touching or nearly touching grains define the boundaries of the central void and its associated strain ellipse. DTCF determines the ratio of the strain ellipse by analyzing data solely from closely spaced grains near the center of the polar plot, while DTCov includes statistics from both touching and distant grains to fit the strain ellipse. Using statistics from distant neighbors may “dilute” the resolution of the central void, and subsequently lead to substantial underestimations or overestimations of strain. The inherent differences between DTCF and DTCov may also explain the large
discrepancy between the results from these two methods in sediment samples from both Shinn (2010) and the Lamar Limestone.

Inherent problems within our methodology may be involved, but the influence of material heterogeneities (Fig. 6.1) is the most likely reason for substantial underestimations of compaction in samples from Shinn (2010). In both the compacted Biscayne and the compacted Rodriguez Key samples, the most obvious effect of compaction is the horizontal consolidation of organic material into black “wispy” seams (Shinn and Robbin, 1983; Appendix B). Because lime muds and organic material tend to compact more readily than granular carbonate sediments, it is likely that these features formed prior to the repackaging of coarse grains within the core. If this is the case, the record of compaction that is recorded by matrix suspended grains may be limited to the amount of compaction that occurred after the consolidation of

Figure 6.1. Compaction of heterogenous carbonate material through time. T1) Presence of mud material and matrix suspended grains with no compaction. T2) Burial of the sedimentary layer initially induces compaction and consolidation of highly compactable muds. T3) Overburden pressure contributes to repackaging of grains after muds have been compacted and consolidated.
organic matter and other highly compactable materials. A limited record of “intergranular” compaction would have a pronounced effect upon our quantification of compaction, due to our reliance on grain-to-grain centroids in developing a strain ellipse, and may influence the spatial and statistical distribution of compaction values in our samples.

Figure 6.2. Distance between centroids for two grains of equal area. A) Flat grains with an area of 3 cm. Each grains respective centroid is separated by a distance of 0.6 cm. B) Circular grains with an area of 3 cm. Each grains centroid is separated by a distance of 1.94 cm.

The statistical and spatial distribution of compaction values is affected profoundly by grain size in carbonate sediment. As flat grains are compacted into dense groups, large numbers of grain centroids are packaged together. As a strain ellipse is calculated for sub-samples of these areas, closely packaged centroids have the potential to induce substantial shortening of the central void. Round grains are also capable of being closely packaged as a result of compaction, but the closeness of their centroids is limited by the size and shape of the grains (Fig. 6.2). Variability of
grain size, shape, and spacing throughout each sample may provide an explanation for varying statistical distributions in compacted samples. We believe that the sporadic nature of packaging, and the diverse shape of matrix-suspended grains cause values of compaction to vary from high values within dense packages to relatively low values in areas where packaging is less dense. In contrast, a lack of secondary grain packaging in uncompacted samples may explain their relatively uniform distributions of compaction values.

6.2. Lamar Limestone Interpretations

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Facies Type</th>
<th>Transect Number</th>
<th>Stylolite Compaction (%)</th>
<th>DTCF Compaction (%)</th>
<th>DTCF and Stylolite Compaction (%)</th>
<th>Projected Height To Shelf Margin (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>GH005</td>
<td>Intraclast Packstone</td>
<td>1</td>
<td>1.38</td>
<td>4.55</td>
<td>5.93</td>
<td>545</td>
</tr>
<tr>
<td>GH010</td>
<td>Skeletal Wackstone/Packstone</td>
<td>2</td>
<td>1</td>
<td>15.38</td>
<td>16.38</td>
<td>533</td>
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<tr>
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<td>Intraclast Packstone</td>
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<td>13.14</td>
<td>16.14</td>
<td>504</td>
</tr>
<tr>
<td>GHTop</td>
<td>Silty Megabrecina</td>
<td>3</td>
<td>2.2</td>
<td>5.34</td>
<td>7.54</td>
<td>489.6</td>
</tr>
</tbody>
</table>

Table 6: DTCF results from Lamar samples used for interpretations.

Due to the greater accuracy of DTCF compared to DTCov, we elected to include only DTCF results in our interpretations of compaction in the Lamar Limestone. Additionally, because the application of grain thresholds increased the accuracy of DTCF results during our methodology tests, we have only included the anticlustered samples. We have combined estimates for compaction by DTCF with
calculations of stylolite compaction to develop a minimum estimate for compaction in basin margin sediments of the Capitan Reef System, and demonstrate that compaction has at least partially controlled the development of the “fall in beds” on the outer shelf.

We believe that facies type is the primary control on intergranular compaction across Lamar Limestone exposures on the northern and southern cliffs of McKittrick Canyon. All samples are positioned within 60 meters of each other in elevation, and have likely undergone similar levels of vertical loading during burial. Porosity-depth curves for carbonate sands developed by Goldhammer (1997) demonstrate that differences of 60 meters in burial produce minimal differences in the porosity of granular facies, and likely would not contribute to substantially different levels of compaction in our samples. Based upon these observations and the spatial distribution of compaction values within our samples, we believe that levels of compaction within the Lamar decrease with the size of matrix-suspended grains. Our most substantially compacted samples, GH010 and GH014, are skeletal packstone facies that consist predominantly of sub-millimeter sized grains, and display some evidence of dense grain packaging. Our least compacted samples, GH005 and GHTop, contain a number of large blocky intraclasts that are up to 2-4 millimeters in length and are virtually devoid of linear grain contacts or dense packages. Because mud-rich facies tend to compact more readily than granular facies, it is reasonable to assume that layers of mudstone and fine wackestone facies within the Lamar experienced higher levels of compaction than our samples have demonstrated. The relationship between fine-
grained sediments and increasing levels of compaction is visible in patterns of stylolite compaction across the Lamar Member.

The most evident discrepancy in stylolite compaction between fine-grained sediments and coarse sediments is noticeable within Transect 1, where the fine-grained wackestone facies at GH007A is ~7% more compacted than underlying layers of packstone at GH002 and GH005 (Table 4). The observation of this trend throughout the Lamar can be attributed to the higher quantity of fine-grained sediment in carbonate wackestones compared to packstones (Ebner et al., 2009). While fine-grained sediment allows stylolite-forming stress concentrations to be distributed across a large surface area, coarse grains suspended within packstone facies limit the accumulation of stress to isolated grain contacts, forcing pressure solution to occur only at isolated locations. The independence of stylolite formation from grain packaging within our samples suggests that our quantification of stylolite compaction can be interpreted as an independent process, justifying the combination of DTCF results with stylolite compaction values to generate a minimum estimate for total compaction (Table 6).
Based upon studies of subsurface expression conducted by Harris and Saller (1999), we believe that our estimates for compaction in samples of the Lamar Limestone provide an indication of differential compaction induced subsidence in the shelf equivalent Tansill Formation. Using reconstructions of the Capitan Shelf Margin produced by Harris and Saller (1999), we estimate that basinward tilting of outer shelf Tansill strata has produced 20 meters of vertical relief between the point

Figure 6.3: Vertical relief generated by subsidence on the outer shelf of the Capitan Shelf Margin for the Tansill, Yates, and Seven Rivers Formations. Figure modified from Harris and Saller (1999).
of maximum subsidence and the horizontal bed located closer to the shelf (Fig. 6.3). In order to account for this relief, strata of the Bell Canyon Formation, including the Lamar Limestone, would most likely need to compact beneath the shelf margin. Currently, the Bell Canyon Formation is ~200 meters thick (Hill, 1996). If the generation of vertical relief had originated solely from compaction of the Bell Canyon Formation, the entire wedge of sediment would need to compact by 19%. This value is only 3% higher than the values estimated for our most highly compacted samples (e.g. GH010 and GH014), suggesting that compaction in the Lamar Limestone may have been significant enough to induce ~20 meters of subsidence. Although our two others samples demonstrate ~11% lower estimates for compaction than the required amount, it is possible that highly compactable mud-rich facies within the Lamar Member and the other limestone tongues of the Bell Canyon Formation have compacted even more severely than the relatively granular facies we analyzed. Additionally, compaction in organic rich facies of the underlying Brushy Canyon and Cherry Canyon Formations may have also contributed to the development of present day relief in the Tansill Formation.

Estimates for the combined vertical relief of all outer shelf strata within the Yates, Seven Rivers, and Tansill formations is approximately 175-180 meters (Fig.6.3). The combined thickness of the basin equivalent Bell Canyon and Cherry Formations is ~600 meters on the distal slope of the shelf margin. In order for these two formations to account for the relief that is presently observed across all outer shelf strata, a compaction of 23% would be required. Although this amount is higher than our estimates for compaction by ~5% in GH010 and GH014, and ~14% in
GH005 and GHTop, it is likely that compaction in fine grained facies within the Bell Canyon and Cherry Canyon Formations accounts for much of this vertical relief.

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Facies Type</th>
<th>DTCF and Stylolite Compaction (%)</th>
<th>Projected Height To Shelf Margin (m)</th>
<th>Thickness</th>
<th>Estimate</th>
<th>Burial Depth</th>
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<td>Intraelast Packstone</td>
<td>5.93</td>
<td>545</td>
<td>3.7</td>
<td>3.93</td>
<td>25</td>
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<tr>
<td>GH010</td>
<td>Skeletal Wackstone/Packstone</td>
<td>16.38</td>
<td>533</td>
<td>9.8</td>
<td>11.71</td>
<td>50</td>
</tr>
<tr>
<td>GH014</td>
<td>Intraelast Packstone</td>
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<td>504</td>
<td>17.5</td>
<td>21.19</td>
<td>25*</td>
</tr>
<tr>
<td>GHTop</td>
<td>Silty Megabrecia</td>
<td>7.54</td>
<td>489.6</td>
<td>5.15</td>
<td>5.56</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 7. Interpretations of estimated burial depth for Lamar Limestone facies based upon compaction results obtained from DTCF. The “+” symbols indicate that more burial depth would be expected to achieve the compaction values we observe in our results. The “*” symbols indicate extrapolation of burial depth values.

Using “decompaction curves” developed by Goldhammer (1997, Appendix F) we have estimated the amount of burial depth required to produce levels of compaction observed in our samples (Table 7). We have interpreted these estimates as a range of burial depths because our samples contain a combination of the “end member facies” represented in these curves. Additionally, we consider our interpretations to only be rough estimates due to the need for extrapolation when applying these curves to layers of sediment thicker than 10 meters. Although the facies types represented by our samples are mud-matrix supported, the abundance of carbonate grains encased within them leads us to believe that the decompaction curve
for carbonate sands is a better representation of the burial depths required to produce our estimates for compaction (Appendix F).

With the exception of GH005, “decompacted” burial depths are lower than estimates for vertical relief between the toe of slope and upper platform during the Guadalupian time (Harris and Saller, 1999). This discrepancy is not surprising, considering that our estimates likely represent a minimum estimate for compaction in the facies we sampled. Even so, our estimates for burial depth are capable of inducing substantial porosity loss in mud-rich carbonate facies on the order 40-50%. The potential for high compaction values in fine-grained sediments is most clearly illustrated by comparing our results for intergranular compaction to measurements of stylolite compaction. Fine-grained wackestone facies of the Lamar consistently demonstrate higher levels of stylolite compaction than coarse-grained packstones. The most extreme example is the burrowed wackestone facies in GH007, which demonstrates higher levels of stylolite compaction than combined estimates for compaction in both GH005 and GHTop. Although stylolite propagation is only partially controlled by mechanical compaction in carbonate sediments, the high magnitude of stylolite compaction in GH007 provides strong evidence for extensive dewatering and subsequent volume reduction (Shinn and Robbin, 1983).
7. Conclusions

Combining methods of Delaunay Triangulation with the continuous function method (DTCF) provides a means for quantifying strain in an anticlustered set of objects with a relatively high level of accuracy. DTCov methods are influenced profoundly by the most distant neighbors in a set of objects and tend to underestimate levels of compaction too substantially to generate accurate results using the methods described in our study.

When applied to fossiliferous carbonate sediments, DTCF can be used to generate a minimum estimate for compaction by analyzing the distances and orientations of adjacent matrix-suspended grains. DTCFs estimates for compaction in carbonate sediment quantify the level of compaction that has been recorded by matrix-suspended grains, but likely exclude the amount of compaction that has occurred in carbonate muds and organic material prior to grain packing.

Statistical analysis of compaction in carbonate sediments reveals that high compaction values are distributed variably throughout compacted samples (>11%), while uncompacted samples (<10%) typically demonstrate uniformly low compaction values. High values of compaction are associated with dense packages of flat grains while lower values are associated primarily with circular or blocky grains and large (millimeter-centimeter) sized grains.

Quantifying compaction in samples of the Lamar Limestone Member using DTCF methods reveals that the compaction of basin margin sediments has likely contributed to the tilting of outer shelf strata on the Capitan shelf margin. Our estimates for compaction
in our most compacted samples of the Lamar Limestone are similar to the level required to produce present day relief in the shelf equivalent Tansill Formation. Although our estimates for compaction are lower than the required level to account for the combined relief observed across outer shelf strata in the Yates, Seven Rivers, and Tansill Formations, the granular nature of our samples suggests that compaction in facies with higher mud/grain ratios may account for most of the compaction that has occurred in the basin margin. Additionally, our estimates of stylolite and intergranular compaction likely represent minimum estimates of compaction in samples of the Lamar, which are within an order of magnitude of the required level to produce present day subsidence of Yates, Seven Rivers, and Tansill strata on the outer shelf.

The demonstration of compaction in basin margin sediments of the Capitan depositional system provides further evidence that differential compaction has profoundly influenced the architecture of the Capitan reef system and that reconstructions of original reef bathymetry must include the role of differential compaction induced subsidence in order to be considered accurate. Additionally the use of center-to-center methods can effectively be used to quantify a minimum estimate for compaction in fossiliferous carbonate sediments whose diagenetic history and original porosity are unknown.
8. Future Work

Future research could apply our methods to basin margin facies positioned closer to the shelf margin, or investigate the significance of compaction in samples from limestone tongues underlying the Lamar. Research could also improve upon our use of the covariance ellipse fitting method by applying a weight function to the most distant neighbors defined by DTNNM. As a final suggestion, the authors of this work propose that improved statistical analysis of compaction could conduct a more extensive study of the relationship between compaction and grain size and/or grain shape, and thereby broaden the understanding of the potential for compaction in a diverse range of facies types.
9. References


Dunham, R.J., 1972, Capitan reef, New Mexico and Texas—Facts and Questions to Aid Interpretation and Group Discussion: Permian Basin Section. SEPM Publication. 72–142, 297 p.


Appendix A: Matlab scripts for DTCF, DTCov, and statistical analysis methods

Please Note: The "%" Symbol indicates the step being carried out to perform the described by italicized and underlined titles.

1. Script for DTCF and DTCov Methods

% commands for DTNNM
% load data from ArcGIS
% datafile ID, area, X, Y
[data]=load('Sample.txt');
ID=data(:,1);
area=data(:,2);
x=data(:,3);
y=data(:,4);
figure;
plot(x,y,'g+)

DT=delaunay(x,y);

% calculate mean radius
rbar=sqrt(area/pi);

% Establish orientation and length variables
th=zeros(size(DT));
r=zeros(size(DT));
dn=zeros(size(DT));

% find convex hull and remove points, iterate several times to clean up % edges
bad=[];
IDt=ID; xt=x; yt=y;
for k=1:3
    qh=convhull(xt,yt);
    bad=[bad; IDt(qh)];
    xt(qh)=NaN;
    j=find(isfinite(xt));
    IDt=IDt(j); xt=xt(j); yt=yt(j);
end

% Set outer points to NaN for analysis
x(bad)=NaN; y(bad)=NaN;
size(find(isfinite(x)));

% plot(x(qh), y(qh), 'r-', x, y, 'b+');
hold on
triplot(DT,x,y,'r-');
axis equal
plot(x,y,'b+');

% find the length and direction for the 3 sides of each triangle
% taking difference of x and y values between each pair of columns
% converting this to angle and distance
for k=1:size(DT,1)
    % xs# = three sides of triangles differences between points on DT
    xs1=x(DT(k,2))-x(DT(k,1));
    ys1=y(DT(k,2))-y(DT(k,1));
Appendix B) Artificially Compacted Carbonate Sediment Samples.

A) Compacted Biscayne Sample (72% compaction)

B) Uncompacted Rodriguez Key Sample (Presumed 0% compaction)

C) Compacted Rodriguez Key Sample (~60% Compaction)
Sample uncompacted Rodriguez Key clustered polygons
Sample uncompacted Rodriguez Key anticlustered polygons
Sample compacted Biscayne polygons
Sample compacted Rodriguez Key polygons
Appendix C: Sample Preparation Procedure

In this appendix we describe the procedure we followed in order to convert our samples into raw coordinate data for analysis using DTCF and DTCoV methods. We carried out these steps after photographs had been scanned and imported into our Microsoft Windows® XP Professional, 64 bit desktop P.C. The programs we used for image processing include: Adobe Photoshop® CS3 version 10.0.1. and ESRI ArcMap® version 2009. These steps can be used in future studies to convert sample photographs to 2-dimensional samples for nearest neighbor analysis.

Image Adjustment Procedure

We began by importing each image into Adobe Photoshop® and converting the images from JPG to Tagged Image File Format (TIFF) files. We converted the files to the TIFF format because TIFF is the preferred format for analyzing and editing photographs in ESRI ArcMap. Second, we adjusted the “brightness,” and the “contrast” of each photograph in order to sharpen the distinction between the grains and the surrounding matrix. We then changed the “mode” of the image from the color spectrum of “Red Green Blue” to the black and white spectrum of “Grayscale.” We chose “Grayscale” mode because it is more effective at displaying color differences between grains and the surrounding matrix than the color spectrum. Next, we adjusted the “threshold” of each photograph in order to display the grains as discrete black objects and eliminate any “noise” clouding their shape or location.
Image adjustment methods in Adobe Photoshop®. A) Adjustment of brightness and contrast. B) Application of Grayscale C) Adjustment of black and white threshold
Vectorization Of Samples

After we finished processing every image in Adobe Photoshop®, we imported each black and white image as a “TIFF” file into ESRI ArcMap®. In ArcMap, we converted the “white” color of the picture to “no color” so that only the location and the size of the grains were represented in our image, without any weight given to the surrounding matrix. Next, we imported an original copy of each processed photograph into the ArcMap file to serve as a template for further editing of the processed image. Using the “draw” and “erase” tools in the “raster editing” tool bar, we continued to edit the processed images on pixel scale to ensure that the tube structures were accurately drawn and the “noise” was accurately removed.

Raster Painting in ESRI ArcMap. Grains can be painted in using the Raster “Paint” tool. Extraneous noise can also be removed using the “erase” tool in the Raster Toolbar.
Processing of GHTop sample (A) into anticlustered vectors (B) and anticlustered centroids (C).
In order to convert the locations and the size of the grains into two-dimensional coordinate data, we “vectorized” the grains represented in each of the processed images using the vectorization tab from the “raster editing” toolbar. This process converted the pixels of the image into “polygons” of quantifiable coordinate data. We then calculated the centroid for each polygon using “spatial statistics tools” in ArcMap. Polygons were chosen to represent the center of arbitrary shapes, inside or outside the grain. We then exported the raw x and y coordinates of each grain to text files, for analysis by DTCF, DTCov, and statistical analysis methods in Matlab® version 2009 (Appendix A).
Appendix D: Transects In Outcrop View

Transect 1 with facies interpretations and stylolite scanlines. A) A view of the cliff from the northern side of the canyon facing south. B) Outcrop view with list of samples. Samples with "*" symbols represent samples that were used to quantify compaction using nearest neighbor and ellipse fitting techniques. No scale bar is provided because the vertical extent of each facies is skewed by the shifting perspective of the photograph. Vertical heights for each facies layer are based upon the observable location of beds rather than the actual height observed in the field.
Transect 2, facies descriptions and associated samples on the northern cliff of McKittrick Canyon. This photograph is a taken facing the northern Cliff from the southern side of the canyon. Stop 6 and Stop 7 represent stops along the Geology Trail in McKittrick Canyon. The cluster of cliff bands to the right of our transect represents the Grotto. "**" symbols represent samples that we used for quantitative analysis using DTCF and DTCov methods. No scale bar is provided because the vertical extent of each facies is skewed by the shifting perspective of the photograph. Vertical heights for each facies layer are based upon the observable location of beds rather than the actual height observed in the field.
Transect 3, with facies descriptions and associated samples on the southern cliff of McKittrick Canyon. This photograph was taken at the foot of the cliff, facing south. Samples obtained from this transect are displayed in yellow text, with an asterisk symbol indicating that compaction was quantified for that sample. No scale bar is provided because the vertical extent of each facies is skewed by the shifting perspective of the photograph. Vertical heights for each facies layer are based upon the observable location of beds rather than the actual height observed in the field.
Appendix E: Samples Of Lamar Limestone
Sample GH005
Sample GH005 Polgyons
Sample GH010 clustered polygons
Sample GH010 anticlustered polygons
Sample GH014 clustered polygons
Sample GH014 anticlustered polygons
Sample GHTop
Sample GH0Top clustered polygons
Sample GHTop anticlustered polygons
Appendix F: Decompaction curves for endmember carbonate facies (Goldhammer, 1997).

Decompaction curve for layers of carbonate mud with present day thicknesses of less than 10 meters. Z represents different burial depths from 25 m to 4000 m.

Decompaction curve for layers of carbonate sand with present day thicknesses of less than 10 meters. Z represents different burial depths from 25 m to 4000 m.