Tectonostratigraphic Evolution Of A Suprasalt Minibasin, Oligocene - Miocene Western Slope Of The Gulf Of Mexico

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TECTONOSTRATIGRAPHIC EVOLUTION OF A SUPRASALT MINIBASIN, OLIGOCENE – MIOCENE WESTERN SLOPE OF THE GULF OF MEXICO

FELIX ALAN RAMIREZ
Master’s Program in Geological Sciences

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by

Félix Alan Ramírez

2016
Dedication

To my lovely wife Karen Garcia

To my caring parents Maria Cuellar and Santiago Ramirez

To my family
TECTONOSTRATIGRAPHIC EVOLUTION OF A SUPRASALT MINIBASIN,
OLIGOCENE – MIOCENE WESTERN SLOPE OF THE
GULF OF MEXICO

by

FELIX ALAN RAMIREZ, B.S.

THESIS

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The University of Texas at El Paso
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of the Requirements
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Abstract

The petroleum industry has focused Gulf of Mexico exploration efforts on salt-related geologic provinces in search of major hydrocarbon accumulations. This work has led to recognition of the influence that salt tectonics has had on play elements of the petroleum system in the producing salt basins. In particular, the complex depositional patterns of the reservoir rocks controlled by shifting depocenters resulting from the accommodation space generated by salt withdrawal from minibasins.

The focus of this research is to understand the structural and stratigraphic components that controlled the evolution of an Oligocene – Miocene suprasalt minibasin located in the upper continental slope of the western Gulf of Mexico. A methodology based on sequence stratigraphic, and structural restoration concepts was used to delineate the influence of these elements. Thus, this research will provide a more predictive model for further exploration of upper slope salt-hosted minibasins in the western Gulf of Mexico.

The study area contains three depositional sequences (Sequence 1, Sequence 2, and Sequence 3) consisting of fine to very fine-grained deep-water turbidites. These sequences were stratigraphically controlled by different rates of accommodation space creation and sedimentation. The depositional sequences are dominated by low stand system tract deposition in the lower parts, and minor transgressive and / or high stand system tract deposition in the upper parts. The studied suprasalt minibasin developed in four evolutionary phases: (1) prethrusting stage, in which the stratal thickness was relatively isopachous, (2) shortening stage, in which the sediments were preferentially confined to synclines between compressional thrusts, (3) compressional folding stage, in which the underlying ductile material (salt or shale) is inflated and the overlying minibasin thin margins are folded, (4) gravitational subsidence stage, in which the minibasin fill is thick and
dense enough to gravitationally sink. The minibasin became elongate to the northeast - southwest, perpendicular to the northwest – southeast regional depositional trend, which was sourced from the Rio Bravo, Rio Grande and Guadalupe rivers.
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Chapter 1: Introduction and Background

1.1 Importance

A general survey of play elements and trap types in producing fields in the Gulf of Mexico (GoM), both onshore and offshore, shows the strong association of major traps and salt tectonics. Salt structures are responsible for approximately 60% of hydrocarbon fields in the GoM basin (Halbouty, 1979), and most of the largest oil and gas producing fields are directly influenced by salt tectonics (D. M Worrall, 1989; T. H. Nelson, 1989; Shengyu Wu, 1990; Diegel, 1993; F. J. Peel, 1996; Schuster, 1996; Rowan and Trudgill, 1999). The petroleum industry has focused GoM exploration efforts in salt-related geologic provinces in search of major hydrocarbon accumulations that can contribute to counteracting the declining production in the main oil/gas producing fields of the world. In particular, the complex depositional patterns of the reservoir rocks controlled by the accommodation space generated, shifting depocenters, and differential of salt uplifting rates created by salt withdrawal minibasins (REF).

On the continental slope of the GoM a large number of salt-withdrawal minibasins have been identified and described in terms of their seismic stratigraphic architecture (Winker, 1996; Prather et al., 1998; Madof et al., 2009), correlation between sea level and facies assemblages (Weimer, 1990; Posamentier and Kolla, 2003), style of sediment dispersal characteristics (Lamers and Carmichael, 1999; Cains et al., 2000; Sinclair and Tomasso, 2002; Marini et al., 2016) as well as structural and stratigraphic relationship and interaction through time (Montoya, 2006) in order to obtain a better understanding of their development.

1.2 Project Goal.

This project improves the understanding of upper slope minibasin architecture. By restoring the basin tectonostratigraphically, its evolution and the timing of the minibasin formation
can be estimated. The study improves the ability to interpret the stratigraphic location, areal distribution, and nature of potential reservoir facies. Results from this study provide a more predictive pathway forward for further exploration upper slope salt-hosted minibasin in the western GoM.

### 1.3 Specific Objectives

1. **Document the detailed structural framework of the minibasin.** The structures were evaluated for relationship to salt tectonic evolution of the minibasin. Additionally, the synsedimentary control on depositional facies and postdepositional partitioning of reservoir facies within the minibasin was analyzed.

2. **Document the stratigraphic framework of depositional facies within the minibasin.** Determining the age and distribution of the most likely reservoir facies, and relating them to their corresponding depocenters in order to develop a predictive sediment dispersal model.

3. **Perform a structural and stratigraphic restoration of minibasin.** Based on results from the first two specific objectives, the tectonostratigraphic evolution of the minibasin was reconstructed, and the features containing possible reservoir facies evaluated.

### 1.4 Regional structural framework

The GoM passive margin, developed as a result of the breakup of supercontinent Pangea during the Early-Middle Jurassic, about 175 Ma (Buffler and Sawyer, 1985; Salvador, 1987; Feng et al., 1994; Bird et al., 2005). The passive margin evolved in several deformation events, which are linked to each other, and overlap in time (Cruz et al., 2010). Central GoM oceanic crust transitions toward the margins into continental crust, which is overlain by several kilometers of salt, known as the Louann Salt that was deposited from Middle to Late Jurassic time (Diegel, 1993;
Deposition of an overlying thick sequence of Mesozoic and Cenozoic sediments resulted in loading of the underlying autochthonous salt and initiation of diapirism and allochthonous salt sheets and canopies (Diegel, 1993).

In the western GoM (Figure 1-1 A), four evolutionary stages have been distinguished (Figure 1-1 B) (Cruz et al., 2010):

- **D0-N event of Middle Jurassic age**: Rifting to form the GoM basin.
- **D1-N event of Late Jurassic – Late Paleocene age**: A linked system with updip gravitational extension and downdip gravitational contraction.
- **D2-N event Eocene age**: Gravitational extension and halokinesis.
- **D3-N event of Late Oligocene to Recent age**: Subdivided into four linked systems, each exhibiting updip gravitational extension and downdip gravitational contraction:
  - Early Oligocene age linked system.
  - Late Oligocene – Recent age linked system.
  - Early Miocene – Recent age linked system.
  - Late Miocene – Recent age linked system.

The D0-N event has been interpreted as a regional event marking the opening of the GoM during the Middle Jurassic (Cruz et al., 2010) (Figure 1-1 B). The D1-N, D2-N, and D3-N events are considered to be caused by post-rift partially localized tectonic and thermal subsidence events, in addition to halokinesis in the basin (Cruz et al., 2010) (Figure 1-1 B).

The Early Miocene – Recent, and Late Miocene – Recent linked systems (D3-N) correspond to an gravitational updip extension – downdip contraction structural schemes that deformed Neogene stratigraphy (Cruz et al., 2010) (Figure 1-1 B). The gravitational extension in
these events is accommodated by normal growth faults on the shelf platform. The Burgos geological province (Figure 1-1 B) was originated by the continuous space created due to the underneath salt withdrawal that resulted in a complex system of normal growing fault (PEMEX E&P, 2013). The eastern contractional feature of this system is the Perdido Fold Belt geological province (Figure 1-1 B). This formed as a consequence of the western down-building process in Burgos, which generated an overburden load enough to transfer the deformation through the detachment surfaces left by the expulsion of the salt in the Salina del Bravo province, and deform the downdip Perdido Fold Belt (PEMEX E&P, 2013) (Figure 1-1 B).
Figure 1-1. Evolutionary stages of western GoM basin formation. A. Map showing position of cross-section TSN-1 across the western GoM. B, Depth cross-section of the western GoM, TSN-1, showing the four evolutionary stages of the GoM. Modified from Cruz et al., 2010.
1.5 Regional stratigraphic framework.

Slope and basinal siliciclastic turbidite sequences deposited in deepwater settings represent the general sedimentological depositional system in GoM during the Tertiary Period. Paleogene turbidite deposits were sourced by the ancestral rivers of Colorado, Mississippi and Rio Grande, (Galloway et al., 2011) (Figure 1-3B). However, during the Neogene period, the Mississippi River was the primary, and largest source of detrital sediments with some secondary inputs such as the Rio Bravo, Brazos, Red, and Tennessee rivers (Galloway et al., 2011) (Figure 1-3B). The western GoM shelf was sourced primarily from the Rio Bravo, Rio Grande, and to a lesser degree the Guadalupe River except during the Middle Miocene when The Guadalupe River became an important source and the Rio Bravo and Rio Grande were inactive (Figure 1-3B).

Figure 1-2 provides the key for the paleographic maps displayed in figure 1-4 to 1-7.

![Paleogeographic Map Elements](image)

Figure 1-2. Explanation for figures used on paleogeographic maps.
1.5.1 Oligocene

This study will evaluate only the Latest Oligocene through the Pleistocene age section of an upper slope area that developed minibasins that localized subsidence during the Upper Oligocene.

In the GoM stratigraphic record, one of the longest and most voluminous episodes of deposition, referred to as “deposodes” (Galloway et al., 2011), is the Oligocene Frio/Vicksburg. During this deposode, the Rio Grande was the largest fluvial input to the GoM, followed by the Mississippi, Houston-Brazos, and Rio Bravo (Galloway et al., 2011) (Figure 1-4). The Rio Grande derived sandstones are characterized by abundant volcanic rock fragments and feldspars (Galloway, 1977; Loucks et al., 1986), which reflect the major influence of the Trans-Pecos, southern New Mexico, and northern Mexico volcanic centers (Galloway et al., 2011). Sands of Houston-Brazos River also contained high proportions of volcanic lithic fragments during this time. But high levels of quartz and feldspar in these sands suggests a mixed source (Galloway et al., 2011) (Figure 1-4). Sands from the Mississippi, on the eastern GoM, are quartz-rich and lack evidence of direct volcanic input (Galloway et al., 2011). Oligocene fluvial sediment output to the western GoM basin was controlled by thermally-driven regional uplift, associated with development of volcanic centers to west and northwest, the distribution of easily eroded volcanic ash, as well as aridity across the Western Interior, which reduced runoff (Galloway et al., 2011).
Figure 1-3. Cenozoic fluvial sources of detrital sediments to the GoM. A. Geographic location of the Cenozoic fluvial axes. B, temporal history of the eight extra-basinal fluvial axes of the northern GoM margin. Modified from Galloway et al. (2011).
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1.5.2 Early Miocene

The Rio Bravo and Rio Grande remained prominent source rivers to the GoM during the early Miocene (Galloway et al., 2011) (Figure 1-5). The ancestral Red River appears due to a shift in the Houston-Brazos axis and change in drainage basin configuration (Galloway et al., 2011). In central Louisiana, the Mississippi river increased in discharge becoming the dominant source area for the Early Miocene era (Galloway et al., 2000).

The Mississippi tributaries drain an enormous area from both the northern Rockies and the upper Midcontinent and eastern plateaus (Galloway et al., 2011). Volcanic relics of northern New Mexico and the Rocky Mountains were drained by the tributaries of the new Red River (Galloway et al., 2011). Basin-margin uplands continued to be drained by the Rio Bravo, and may have extended into the adjacent Basin and Range province (Galloway et al., 2011). During this time outlying, easily eroded volcanic uplands decreased in elevation, ash dispersal was largely diminished, drainage basin area decreased, local subsidence along the Rio Grande Rift began to segregate sediment drainage systems, and arid conditions limited erosion and transport of sediments from the Western Interior resulting in decreased sediment supply into GoM (Galloway et al., 2011). This changed, later in the Early Miocene, when the eastern interior became the dominant source area (Galloway et al., 2011).
Figure 1-5. Early Miocene paleogeography (modified from Galloway et al., 2011). RB, Rio Bravo. RG, Rio Grande. R, Red River. M, Mississippi River. Compiled from Bart, 1975; Scott, 1982; Cather et al., 1994; Chapin and Cather, 1994; Pazzaglia and Kelley, 1998; Connell et al., 1999; Holm, 2001; Buffler, 2003; McMillan et al., 2006; Cather et al., 2008; Flowers et al., 2008. See explanation of map patterns on Figure 1-2.
1.5.3 Middle Miocene

During this time, the Mississippi river became the main fluvial sediment source to the GoM (Galloway et al., 2011). Immediately to the east of the Mississippi a new fluvial system appeared (Tennessee River), which had a drainage basin coming close to that of the modern Tennessee (Figure 1-6) (Galloway et al., 2011). In the northwestern GoM, the Guadalupe River peaked as a supplier of sediment.

High rates of erosion in the Appalachian uplands, transportation, and subsequent sedimentation into the GoM are suggested as the consequence of climate changes, such as increasing storm intensity or frequency (Boettcher and Milliken, 1994). Thus, the drainage basin evolution, and consequent pattern of sediment yield to the GoM were thoroughly influenced by the interplay of tectonic and climatic events (Galloway et al., 2011):

1. The Rio Grande Rift accelerated extension and consequent subsidence (Galloway et al., 2011), breaking up the Rio Grande drainages.
2. The Central and northern Rocky Mountains were the main sources during this arid climate (Galloway et al., 2011).
3. The Cumberland Plateau and Blue Ridge were reactivated, eventually resulting in ~1 km of erosional unroofing (Boettcher and Milliken, 1994).
Figure 1-6. Middle Miocene paleogeography (Galloway et al., 2011). G, Guadalupe River, M, Mississippi River, Tennessee River. Compiled from Chapin et al., 2004a; Seni, 1980; Scott, 1982; Boreman, 1983; Boreman et al., 1984; Gustavson and Winkler, 1988; Brister and Gries, 1994; Cather et al., 1994; Chapin and Cather, 1994; Steven et al., 1997; Connell et al., 1999; Holm, 2001; Buffler, 2003; McMillan et al., 2006; Chapin, 2008; Flowers et al., 2008. See explanation of map patterns on Figure 1-2.
1.5.4 Late Miocene

In the northern GoM, the Mississippi and Tennessee fluvial axes were the main depocenters. While still present, the Guadalupe axis declined in influence; and along the southwestern margin, the Rio Grande and Rio Bravo axes existed as discreet, but important elements (Galloway et al., 2011). A trench and barrier were created by the Rio Grande Rift, which trapped the sediments in the central and south Western Interior and barred them from reaching the GoM, (Figure 1-7) (Galloway et al., 2011).

The sediment supply to the GoM decreased markedly during Late Miocene due to several factors (Galloway et al., 2011):

1. Decrease in unroofing and uplift of Appalachians (Galloway et al., 2011).
2. Sediment trapping by the Rio Grande Rift and relic Rocky Mountains on the Western Interior, and eventual progressive decrease deposition in the GoM (Galloway et al., 2011).
3. Low precipitation over the eastern Rocky Mountains generated energy-deficient rivers that carried little sediments to the GoM (Galloway et al., 2011).
Figure 1-7. Late Miocene paleogeography (Galloway et al., 2011). RB, Rio Bravo. RG, Rio Grande. G, Guadalupe River. M, Mississippi River. Tennessee River. Compiled from Chapin et al., 2004a; Scott, 1975; Seni, 1980; Boreman, 1983; Boreman et al., 1984; Gustavson and Winkler, 1988; Brister and Gries, 1994; Cather et al., 1994; Chapin and Cather, 1994; Steven et al., 1997; Connell et al., 1999; Holm, 2001; Perkins and Nash, 2002; Buffler, 2003; Mack, 2004; McMillan et al., 2006; Chapin, 2008. See explanation of map patterns on Figure 1-2.
1.6 Description of study area

The project is located on the western upper continental slope of the GoM, between 500 and 1,000 meters of bathymetric depth. Geologically, the study area is defined by an Oligocene – Miocene upper-slope minibasin oriented NE-SW. The minibasin partially overlies allochthonous salt and diapiric shale, and was filled by over 4,000 m of fine-grained mostly turbidite sediments derived from the distal Rio Bravo and Rio Grande.

Salt withdrawal also resulted in an extensional system, characterized by normal faults oriented mostly perpendicular and radially to the salt bodies or domes. The main part of this faulting is only effective from Neogene to Recent sequences, offsetting the seafloor surface. However, particularly in the deeper sequences of the minibasin, it is possible to interpret some thrust faults that could have influenced Paleogene strata.
Chapter 2: Integrated interpretation.

2.1 Technology.

This project was completed using Petrel E&P software platform due to its ability to integrate seismic, well logs, and other types of data in a convenient, and efficient structure. Additionally, structural restorations were performed in Move™ to fully integrate 2D and 3D model building and analysis.

2.2 Well A data interpretation.

Four types of well logs were analyzed (gamma ray, resistivity, density and sonic) from a single well, located near the southern edge of the study area (well A) in order to identify generalized fining and coarsening upward trends in turbidite strata. The amount of shale compaction, and stratal stacking patterns were interpreted to describe depositional sequences in the Late Oligocene to Late Upper Pliocene section.

2.2.1 Gamma ray log observations.

The gamma ray log measures the natural radioactivity of the formations, which reflects the shale content due to concentration of the radioactive elements in clay minerals (Serra, 1984). Radioactive elements, mostly potassium, thorium and uranium, tend to concentrate in clays contained in shales, while clay free strata commonly exhibit low levels of radioactivity (Schlumberger, 1991).

Stratigraphic variation in facies such as the grain size increase (coarsening upward) or decreases (fining upward) was to analyze major trends in depositional facies (Serra and Sulpice, 1975). Even though this methodology is mainly used for small-scale interpretations (stratal bodies of 10’s meter thick) it can be applied to major stratal stacking patterns (stratal bodies of 100’s meter thick) within a sequence stratigraphic connotation. Coarsening upward successions are
inferred to represent a shoreline migration into the basin (Posamentier et al., 1992) and the upward coarsening is inferred to result from a progradational stacking pattern (Wagoner et al., 1990). Conversely, fining upwards succession represent landward migration of the shoreline (Loutit, 1988), forming a retrogradational stacking pattern (Wagoner et al., 1990) (Figure 2-1).

Figure 2-1. Parasequence stacking patterns (modified from Wagoner et al., 1990).

The observed gamma ray log values from the well A are all from very fine-grained lithologies. The minimum Gamma-Ray value observed in the log is 72 API (Figure 2-3a), which commonly indicates siltstones (Figure 2-3B). The maximum, gamma ray is about 120 API (Figure 2-3a), which corresponds to typical values of non-marine to marine or bituminous shale (Figure
However, in the study area the most likely interpretation is marine shale or shales with interspersed fine-grained silts or sands. Other possible interpretations include coarser-grained sediments containing concentration of glauconite or heavy mineral bands (Figure 2-3B). To test this idea further information, such as natural gamma ray spectrometry and litho-density log, are needed to detect, identify, and evaluate the presence of large amounts of radioactive minerals besides clays (Schlumberger, 1991) (Figure 2-2).

Figure 2-2. Mineral identification from Litho-Density log and natural gamma ray spectrometry log. (A) Photoelectric factor vs thorium/potassium ratio, (B) Photoelectric factor vs potassium concentration (modified from Schlumberger, 2009).

2.2.2 Resistivity log observations.
Resistivity logs are acquired by either transferring current into the formation and a receiver quantifying the ease of the electrical flow through it or by inducing an electric signal into the formation and measuring the response (Serra, 1984; Schlumberger, 1991). The majority of the rocks in formations are fundamentally insulator due to the water confined in the interstitial porous is conductor, but highly resistive when hydrocarbon is present instead (Rider, 1996). Even though the principal use of the resistivity log is to find hydrocarbons, it can contribute information on weathered layers, aquifers, lithology, texture, facies, over-pressure, and source rock aspects (Palacky, 1987; Rider, 1996) (Table 2-1 and Figure 2-4).
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The range of responses measured by the resistivity log also indicate very fine-grained sediments within the minibasin. Interpretation is challenging when plotting resistivity values using conventional or typical log responses (0.1 to 1,000 ohm.m) (Figure 2-4B), because the range between minimum and maximum resistivity values is so small (0.33 ohm.m to 0.8 ohm.m) (Figure 2-4A left hand log). Once the plot is constrained to its lowest and highest range, general tendencies are observed (Figure 2-4A right hand log).
2.2.3 **Density log observations.**
Density logs are acquired by applying a radioactive source to the borehole wall, which releases gamma rays into the formation. The radiation is scattered, and collides with atomic nuclei in the formation. A count of dispersed gamma rays reaching the tool detector is an indication of formation density (Serra, 1984; Schlumberger, 1991).

Similarly to the gamma ray and resistivity logs, the density well log data from well A suggests mostly fine to very fine-grained sediments in the study area. Taking this into account, the subtle density fluctuations are interpreted to be a function of compaction. Poorly compacted shales are assumed for the lower range of values (1.87 to 2.0 g/cm$^3$), and moderately compacted shales for the higher range (2.0 to 2.36 g/cm$^3$) (Figure 2-5). Lithologies with similar density responses are salt, sandstones with 20% of porosity containing gas, oil or water (Figure 2-5B). However, these interpretations are not probable since there is no indication of the presence of oil, gas or salt in the corresponding gamma ray, resistivity and sonic logs (Figure 2-4 and Figure 2-3).

2.2.4 **Sonic log observations.**
The sonic log measures the transit time that it takes for a sound pulse emitted from a transmitter to travel into the formation and get back to the receiver (Schlumberger, 1991). This transit time, also known a slowness, is the reciprocal of the velocity of the sound, which depends upon the formation porosity and lithology (Serra, 1984).

The sonic log values in well A also indicate dominantly silt to shale lithologies, matching the other well logs. The velocity of sound decreases as the porosity of the rock increases and, consequently, increases the interval transit time. The sonic values of well A range from 115 to 155 μs/ft, suggesting poorly compacted shales in general, where the lower values suggest slightly more compacted shales.
Figure 2-3. A, Blocky gamma ray log signature interpretation for well A and its corresponding symbology. B, Typical gamma ray log responses (adapted from Rider, 1996).
Figure 2-4. A, Blocky resistivity log signature interpretation for well A and its corresponding symbology. B and C, Typical resistivity log responses (adapted from Palacky, 1987; Rider, 1996). Note, apparent differences are enhanced due to expanded scale.
Figure 2-5. A, Blocky density log signature interpretation for well A and its corresponding symbology. B, Typical density log responses (adapted from Rider, 1996).
Figure 2-6. A, Blocky sonic log signature interpretation from well A and its corresponding symbology. B, Typical sonic log responses (adapted from Carmichael, 1982; Rider, 1996).
2.2.5 Interpretation of well logs.  
The stacking patterns identified from the gamma ray and resistivity log signatures were used to associate the interplay between deposition and the rate at which the space in which sediments can fill was increased or decreased. Thus, when the accommodation space is increased, the deposition of fining strata is favored due to trapping of coarser sediment farther from the slope margin, resulting in fining upward successions. The opposite case, when the accommodation space is reduced, results in a progradation of the source closer to the basin, depositing coarsening upward strata. Density and sonic logs (Figure 2-5 and Figure 2-6), were analyzed in compaction rates context. Therefore, density increasing and sonic decreasing values occur where correlated to an increase in the compaction rate, relative to deposition. Meanwhile, density decreasing and sonic increasing values trends where associated to a decrease in the compaction rate relative to deposition.

2.2.5.1 Upper Oligocene  
The Upper Oligocene is characterized by a general fining upward stratal successions expressed in the gamma ray log of well A. This indicates that the accommodation space creation rate increased during this period (Figure 2-7). The Oligocene interval of the resistivity log exhibits an upward value decrease, which, based on the very small range of resistivity values, may indicate fluctuations in the salinity of formation water in addition to a coarsening-upward trend (Figure 2-4 and Figure 2-7). This interpretation is reasonable since the formation resistivity could be influenced by the lithologic matrix, formation water, zone of transition, and flushed zone (Schlumberger, 1991). General trends on the density and sonic logs of the Oligocene section display slight increasing compaction, which partially coincides with the accommodation space increase observed in the gamma ray log (Figure 2-7).
2.2.5.2 **Lower Miocene**

The Oligocene major fining upward stratal trend observed in the gamma ray log of the well A lasted to almost the top of the Lower Miocene strata, when it began to coarsen upward (Figure 2-7). This shift is possibly an indication of a change in the accommodation space creation rate, which allowed progradation of the shoreline and deposition of slightly coarser sediments. The resistivity log exhibits a low frequency fining upward stacking pattern that continues up to the central part of the Middle Miocene (Figure 2-7). The increasing compaction trend that started in the Upper Oligocene lasted to the top of the Lower Miocene in the density log, and to the basal section of the Middle Miocene in the sonic log (Figure 2-7). Further analysis presented in the following chapters will discuss the tectonic and stratigraphic elements that could influenced the deposition of a thin section of Lower Miocene where the well A is located.

2.2.5.3 **Middle Miocene**

High frequency of fining and coarsening upward strata are developed from the center to the uppermost Middle Miocene section of the gamma ray log (Figure 2-7). This particular pattern can be related to local lateral shifts of turbidite systems that brought slightly different size-grained sediments during certain short episodes. Another possible reason is due to rapid changes in the subsidence rate that generated short periods of increasing and decreasing accommodation space rates. The resistivity signature presents a shift from fining to coarsening upward at the central part of the Middle Miocene that last to the top of the Upper Miocene (Figure 2-7). This long frequency tendencies are related with a rapid and slow uplifting rate that resulted in an increasing and reducing accommodation creation rate (Chapter 4: Tectonostratigraphic controls on evolution). The lower half part of the Middle Miocene presents a high correspondence of the density and sonic signature with the gamma ray well log, which suggests that the compaction tendency might be influenced by the sediment grain size deposited during the change of reducing to increasing
accommodation space creation rate (Figure 2-7). The upper half part of the Middle Miocene section of the density log shows a long term lessening compaction trend, which is similar to the low frequency accommodation decrease tendency of the resistivity log, suggesting the density log signature was probably influenced by the slow uplifting rate of the area (Chapter 4: Tectonostratigraphic controls on evolution). However, the same section recorded by the sonic log shows short term increasing and decreasing compaction trends, similar to those fluctuating accommodation space creation rates of the gamma ray log, suggesting both the density and gamma ray log signatures were influenced by either local lateral migration of turbidite systems or rapid changes in the uplifting rates (Chapter 4: Tectonostratigraphic controls on evolution).

2.2.5.4 Upper Miocene

The gamma ray log signature of the Upper Miocene section of well A is a thick succession (~650 m) that reflects two major gradation patterns that differ from each other because of a slightly grain size variation. The first half is compound of coarsening upward strata resulting from a decrease of accommodation space, and the second half correspond to fining upward successions derived from an accommodation space decrease. The resistivity, density and sonic log signatures of the Upper Miocene are defined by some long term tendencies. The resistivity log presents a general slightly coarsening upward strata that starts from the central part of the Middle Miocene and that lasts to the basal part of the Lower Pliocene. The density and sonic logs are characterized by a continuous lessening compaction tendency that starts from the latest section of the Middle Miocene and lasted to probably recent strata. This major tendencies are believed to be related with an isostatic subsidence evolutionary stage that will be discussed in Chapter 2: Palinspastic restoration and Chapter 4: Tectonostratigraphic controls on evolution.
Figure 2-7. Integrated electrofacies interpretation of well A. Gamma Ray (GR) and Resistivity (Res) logs showing accommodation increase and decrease tendencies. Density (RHOB1) and Sonic (DTC) logs showing more or less compacted periods.
2.3 Seismic interpretation.
A detailed seismic three-dimensional interpretation of both significant stratal horizons and structures (faults) was completed in order to delineate the depositional sequences as a basis to interpret the history of the study area. Sequences were mapped throughout the study area in terms of stratal termination types, thickness and seismic facies in order to determine controls on the depositional systems through time.

2.3.1 Correlation of significant seismic surfaces and delineation of stratal packages.
Three sequences were bounded by identifying the stratal horizons that potentially represent sequence boundaries (SB), which are regional stratal surfaces defined by stratal truncation below the surface and stratal onlap above (Vail et al., 1977; Galloway, 1989; Copus, 1997). Sequence 1 comprises strata deposited during the Late Oligocene to the lower part of the Middle Miocene. Sequence 2 is developed from the lower to the middle section of the Middle Miocene. Sequence 3 corresponds to the central Middle Miocene to top of the Upper Miocene interval.

2.3.2 Interpretation of stratal terminations.
The methodology used to represent each horizon consisted of mapping stratal termination types in order to visualize how the strata terminate in map view. The corresponding lower sequence boundary surface is displayed in the background using a blue-white shaded surface for relative depth below sea level. Meanwhile, subsequent stacking of parasequences are presented as transparent contour surface overlapping the sequence boundary surface. Hence, it is possible to identify through time and space when and where successive stacked parasequences were present, pinched out, and or erosionally or structurally truncated.

2.3.3 Analysis of depocenters.
Thickness maps were calculated for each sequence by subtracting the sequence base elevation from the sequence top elevation in order to determine where the principal minibasin
depocenters were located through time. Sequence thickness maps are important because local depocenters in a minibasin may shift during the interplayed growth of the restricting topography and accommodation space creation (Sinclair and Tomasso, 2002; Montoya, 2006; Madof et al., 2009).

In the interest of obtaining the correct thickness of each sequence, true stratigraphic thickness maps were calculated. This process consisted of calculating the thickness of a rock layer measured perpendicular to the layer, with corrections for bed dip (Schlumberger, 2016) (Figure 2-8).

Figure 2-8. True stratigraphic thickness scheme (modified from Schlumberger, 2016).

For successively younger sequences, the same scheme utilized on the previous section is followed except now the lower sequence boundary is displayed in the background using a blue-white scale shaded surface for elevations. This surface is overlayed with the thickness of each depositional event presented as a gray scale color surface overlapping the lower sequence boundary. In addition, stratal terminations from the underlying interval are also displayed in order
to appreciate their relationship to overlying depositional system. Therefore, it is possible to recognize through time and space, when and where successive set of parasequences were stacked or blanketed the study area.

2.3.4 **Interpretation of seismic facies.**

The procedure applied to characterize the seismic facies consisted in calculating the corresponding seismic attribute of each interpreted surface due to in many cases, the amplitude of reflection tied directly to the porosity or to the saturation of the underlying layer (Taner et al., 1979; Robertson and Nogami, 1984; Kalkomey, 1997). The most elementary attribute, and the one most commonly utilized, is seismic amplitude, which quantifies the reflectivity within a time or depth window, giving back the maximum positive or negative number in the defined window (Schlumberger, 2014).

All, either positive or negative amplitude maps were interpreted in terms of the likelihood to represent reservoir facies. High values of amplitude were defined as the most likely reservoir facies, the intermediate values to be the most likely overbank reservoir facies, and low values as the non-likely reservoir facies (Shanmugam et al., 2009) (Figure 2-9). This interpretation was overlapped onto the underlying lower sequence boundary of each sequence, following the same arrangement as for those of stratal termination and thickness maps. The lowstand system tract (LST) was identified by displaying a SB at the bottom, transgressive surface (TS) at the top, and progradational stacking patterns. Transgressive system tracts (TST) are bounded below by the TS and above by the maximum flooding surface (MFS), and internally form retrogradational stacking patterns. Highstand system tracts (HST) are bounded below by the MFS and above by the SB, and internally form progradational to aggradational stacking patterns.
2.3.5 **Seismic sequence stratigraphy**

This section defines the geological elements that influenced the stratigraphic history of the studied minibasin utilizing the sequence stratigraphic methodology and concepts proposed by Catuneanu (2006), and from examples described by different authors (Blackwelder, 1909; Campbell, 1967; Vail et al., 1977; Posamentier et al., 1992; Coe et al., 2003; Montoya, 2006). In this section, the significance, methodology, results, and interpretations from Sequence 1, Sequence 2 and Sequence 3, will be described utilizing a Wheeler diagram generated from the Seismic Section A, and the gamma ray log signature.

The sequence stratigraphic method is based on the recognition and identification of stratal termination types, major stratal surfaces, and parasequence set stacking pattern types to place stratal packages into systems tracts and a depositional sequence stratigraphic framework.
A Wheeler diagram (Wheeler, 1958; Wheeler, 1964) from the interpreted sequence stratigraphic framework was generated in order to display both the horizontal distribution of the coexistent component sedimentary layers of the minibasin and also the substantial hiatuses in sedimentation in the area. A stratigraphic summary chart was generated on which geological time was plotted as the vertical scale, and distance across the minibasin as the horizontal scale, and on which a variety of structural and stratigraphic information was gathered (Mitchum, 1977; Treviño et al., 2005; Catuneanu, 2006) (Figure 2-10 and Figure 2-11). In order to compliment the stratigraphic summary, a sequence stratigraphy analysis was added on the left of the Wheeler diagram. Thus, the spatial distribution of each sequence can be tied to their corresponding system tract (Figure 2-11).

The sequence stratigraphic interpretation was accomplished using the stratal termination and major stratal surfaces identified in Section A and the well log data of well A. Section A is down regional dip line located in a structurally active zone (salt and shale tectonics) that crosses the margins of three active sub basins in the area (depocenter X, Y, and Z) (Figure 2-11 and Figure 2-13). The well log utilized to interpret the sequence stratigraphic framework was the Gamma Ray because of its reliance on the amount of clay content, which in turbidite systems can be used as a proxy for sedimentation rate and relative sea-level changes in accommodation (Figure 2-12)

Three main depocenters where observed through time, and named for simplicity as X, Y, and Z (Figure 2-11). In addition, three principal ridges were distinguished to be controlling the deposition in depocenter X, Y, and Z by being active or inactive through time. These structural highs were denominated as A and B (salt), and C (shale), and locally confined deposition within the surrounding sub basins (Figure 2-11).
Figure 2-10. Wheeler diagram illustrating the time/distance relationships of facies (Catuneanu, 2006).
Figure 2-11. Wheeler diagram and sequence stratigraphy interpretation of section A.
Figure 2-12. Sequence stratigraphic interpretation on schematic Gamma Ray well log.


2.3.5.1 **Sequence stratigraphy of Sequence 1**

The Sequence 1 is bounded at the base by SB-1 and at the top by SB-4. The sequence is composed primarily of LST strata with a relatively thin TST and HST (Figure 2-11 and Figure 2-12). During the LST, the accommodation space creation rate is reduced resulting in restricted and confined available areas to capture sediments during Sequence 1 deposition (Figure 2-11 and Figure 2-12). The Wheeler diagram of Section A shows a progradational stacking pattern where depocenter X is located in the minibasin, confirming the interpretations regarding a confined sediment allocation (Figure 2-11) due to the shale-controlled structural high A that was active to the west of depocenter X, and the salt-controlled structural high C active to the east. (Figure 2-11). Depocenter Y received less sediment than depocenter X during the LST, suggesting it did not represent a significant structural low feature. The LST of Sequence 1 is bounded at the top by the TS denoted by a change in stacking pattern from progradational (coarsening upward) to retrogradational (fining upward), corresponding to the beginning of the TST (Figure 2-11 and Figure 2-12).

The TST of Sequence 1 is characterized by a widespread sediment distribution over the study area, covering depocenters X, Y, and Z (Figure 2-11), and by fining and thinning upwards parasequences showing a retrogradational stacking pattern (Figure 2-12). The retrogradational stacking pattern reflects an incremental increase in accommodation space creation. The structural highs A and C are blanketed by sediment at this time to the west and east of the area, respectively, suggesting that tectonic activity was reduced during this time (Figure 2-11). The end of the TST is characterized by the finest grained succession of sediments deposited during the MFS (Figure 2-12)
The MFS that capped the TST also delineates the beginning of the HST, which is distinguished by the widespread distribution of sediments across the area, again covering depocenters X, Y, and Z (Figure 2-11), and by a series of coarsening-upward successions displaying a progradational stacking pattern (Figure 2-12). As observed in the TST, the structural highs A and C seem to be inactive during this period permitting the deposition of sediments across the area. At the top of the HST, the SB-4 is located indicating both the upper limit of the Sequence 1 and the lower limit of the Sequence 2.

2.3.5.2 Sequence stratigraphy of Sequence 2

The Sequence 2 is bounded at the base by SB-4 and at the top by SB-5. It is dominated by the LST, but has a thin TST above it (Figure 2-11 and Figure 2-12). Identification of the SB-4 that separates the Sequence 1 and Sequence 2 was based on the particular observations and interpretations when comparing the Wheeler diagram of Section A and the Gamma Ray log signature. SB-4 at the base of the Sequence 2 is a regional stratal surface demarcated by erosional truncations below it and stratal onlap above (Figure 2-11). It is clear from the well log signature that the stacking pattern above the HST of the Sequence 1 is retrogradational, suggesting a TS instead of a SB (Figure 2-12). This differences observed at the contact between the Sequence 1 and Sequence 2 is due to the different location of the Section A and the well log. The Section A lies where the differential subsidence led to a local increment in the generation of accommodation space to develop depocenter Y, which was the principal area that capture sediments during the Sequence 2 period (Figure 2-11). Nonetheless, the well log is a punctual data that recorded information from a place located to the south of depocenter Y, allowing only documentation fining-upwards strata in the TST, but not the LST deposited above the Sequence 1 in depocenter Y (Figure 2-11 and Figure 2-12).
The LST of the Sequence 2 is bounded below by a SB and TS above, and internally characterized by onlapping stratal packages (Figure 2-11). The locus of deposition shifted due to subsidence of depocenter Y, reducing the deposition in depocenter X and Z (Figure 2-11). The Wheeler diagram of Section A shows that the early part of the LST is defined by strata that onlap against the shale-controlled structural high A at the western margin of depocenter X, and erosionally truncated on the eastern margin due to the active salt-controlled structural high B (Figure 2-11). Furthermore, since the salt-controlled structural high B and C were active (to the west and east respectively) the margins of depocenter Y were characterized by erosional truncations (Figure 2-11). However, the Wheeler diagram and the well log show that the deposition apparently occurred only within depocenter Y during the LST of Sequence 2. This localize deposition probably occurred because this area had rapid subsidence in comparison with depocenter X, a probable rapid structural uplifting resulted in the erosional truncation of sediments and no LST strata in well A (Figure 2-11 and Figure 2-12). This pattern continues until a TS marks the end of the LST and beginning of the TST.

The uppermost strata in Sequence 2 corresponds to a TST, which is bound at its base by a TS and a SB above. The TST is internally contains a series of flooding surfaces stacked in a retrogradational pattern, and distributed all around the area of study (Figure 2-12). Finally, the SB-5 limits the TST at top, indicating the end of the Sequence 2, and the starting point of Sequence 3.

2.3.5.3 Sequence stratigraphy of Sequence 3

The Sequence 3 is bounded by the SB-5 at the base and the SB-6 at the top. The LST dominates the sequence, but is overlain by a thin TST, and overlying HST (Figure 2-11 and Figure 2-12).
The LST of the Sequence 3 is bounded at bottom by SB-5 and at top by a TS. Strata onlap SB-5 (Figure 2-11) and form a progradational stacking pattern (Figure 2-12). This sequence is distinct from the LST of the Sequence 1 and Sequence 2 because the deposition occurred in depocenter X and Y at the same rate, suggesting a constant subsidence over this areas (Figure 2-11). From the Wheeler diagram of the Section A, it is clear that both margins of the LST strata in depocenter X and Y onlap against the structural highs A, and B and C, respectively (Figure 2-11). The LST of Sequence 3 is bounded by the TS where the stacking pattern shift from progradational to retrogradational, denoting the beginning of the TST (Figure 2-11 and Figure 2-12).

The following genetically associated stratigraphic units correspond to a TST, which is bounded at bottom by a TS and at top by a MFS (Figure 2-11 and Figure 2-12). This stage is characterized by deposition that blanketed depocenters X, Y, and Z (Figure 2-11), and by a retrogradational stacking pattern demarcated by a finning-upward succession (Figure 2-12). As well as the TS of the Sequence 1 and 2, the ample spreading of sediments during the TST of the Sequence 3 is probably due to an associated increment in the accommodation space creation rate, and a growing cessation of the structural highs A, B, and C. The deposition of the TST lasted until it reached its shaliest portion at the MFS, which in turn marked the beginning of the HST of the Sequence 3.

The upper most system tract of the Sequence 3 is the HST, which was formed when sediment accumulation rates exceed the rate of increase in accommodation space. The HST is limited below by a MFS and above by the SB-6, and internally defined by a progradational stacking pattern (Figure 2-11 and Figure 2-12). The amount of available space led to the deposition of sediment over depocenters X, Y, and Z.
2.3.6 **Results of seismic interpretation.**
This section of this chapter is aimed to present the areal distribution of the interpreted SB and internal depositional features, such as the stratal termination, thickness and reservoir facies maps present in Sequence 1, Sequence 2, and Sequence 3.

2.3.6.1 **Sequence boundaries**
Sequence 1, 2, and 3 are stratigraphically bound at the bottom and top by their corresponding SB. The lower SB of Sequence 1 is the SB-1 and SB-3, and is displayed in Figure 2-13, while the upper limit corresponds to surface S1-4 (SB-4) (Figure 2-14 and Figure 2-15, S1-4), which in turn corresponds to the lower SB of Sequence 2 (Figure 2-13). The last depositional stage, and the upper boundary of Sequence 2 is the S2-6 (SB-5) (Figure 2-14 and Figure 2-18, S2-6). Sequence 3 is bounded at the base by the S2-6 (SB-5) and at the top by S3-5 (SB-6). All lower SB of each Sequence display their current deformed state, and define three main depocenters designated depocenter domains X, Y, and Z (Figure 2-13).
Figure 2-13. Map view of the interpreted lower sequence boundary of Sequence 1, 2, and 3. Dark blue deeper, and light blues to white are the shallower sections.
Figure 2-14. Cross-section A showing the Sequence 1, 2, and 3 most significant internal surfaces, S-#. 
2.3.6.2 Sequence 1

Sequence 1 comprises strata deposited during the Late Oligocene to the lower part of the Middle Miocene, and contains four major internal seismic reflectors (Figure 2-14 and Figure 2-15). The general depositional trend of the Sequence 1 interpreted from the most likely reservoir facies orientation is northwest – southeast. The distribution of Sequence 1 onlaps at west against a shale-controlled structural high, and erosionally and structurally truncated over the east by a salt-controlled structural high. Local lateral shifts of thick successions are observed through time.

Stratal package (SP) 1-1 represents strata bounded at the base by the SB-1 and SB-3, and above by surface S1-1. SP1-1 was primarily deposited during a LST (Chapter 2.3.5.1 Sequence stratigraphy of Sequence 1) in the center and deeper regions of depocenter X, which corresponds to the main depocenter of the study area (Figure 2-15 depocenter X). The eastern limit of S1-1 is defined by onlapping fronts, indicating deposition onto a westward dipping inclined surface, which is coincident with a down-regional dip structural high displayed by the topographic relief of basal SB surface (Figure 2-15, S1-1, locality 1). In conjunction with the onlapping fronts, there is also structural truncation and local erosional surfaces, which persist for the next two stratigraphic levels suggesting a continuous erosional process through S1-1 to S1-3 (Figure 2-15, S1-2 and S1-3, locality 2). The western region is partially covered by SP1-1, which may indicate that the western predepositional bathymetric relief was not as high as the relief to the east, but still elevated enough to confine the SP1-1 deposition within depocenter X (Figure 2-15, S1-1, locality 3). Erosional and structural truncations restrict the distribution of S1-1 in the west, which may signify that this region was structurally active contemporaneous with its deposition (Figure 2-15, S1-1, locality 4). SP1-1 shows a relatively consistent thickness over its distribution delimited by onlapping and truncations on the eastern and western limits respectively, which in turn are associated with slight thickness
reduction toward the depositional limits (Figure 2-16, SP1-1, locality 1). Despite its somewhat uniform thickness, some prominent northeast to southwest depositional trends can be interpreted for SP1-1, with a possible main depocenter located on the southwestern region, indicated by the thicker succession (Figure 2-16, SP1-1, locality 2). A general northeast-southwest depositional trend of channelized facies is observed, having some of the most likely reservoir facies confined to where the sediments were stacked resulting in a thicker succession (Figure 2-17, SP1-1, location 2), and other areas with likely reservoir facies with thin thickness (Figure 2-17, SP1-1, location 1). The depositional pattern of reservoir facies from SP1-1 is closely associated with the local thick areas shown on the thickness map. Some other possible reservoir overbank facies are irregularly deposited over the northwest region, and show no relationship to thickness of SP1-1 (Figure 2-17, SP1-1, location 3).

SP1-2 represents strata bounded at the base by the S1-1, and above by the S1-2. While SP1-3 is defined in the lower limit by the S1-2 and the upper limit by the S1-3. Both SP1-2 and SP1-3 were deposited during a LST (Chapter 2.3.5.1 Sequence stratigraphy of Sequence 1) Deposition of SP1-2 and SP1-3 are restricted to depocenter X (Figure 2-15, S1-2 and S1-3). SP1-2 has an areally smaller distribution than SP1-3, but is similar in depositional history. Both, SP1-2 and SP1-3 display a west, north and southwestern onlapping stratatal termination pattern (Figure 2-15, S1-2 and S1-3, location 4), which may suggest their depositional entrance into the minibasin was in a preferentially northwest – southeast direction. In the meantime, mostly erosional and structural truncation on the eastern margin of both SP1-2 and SP1-3 (Figure 2-15, S1-2 and S1-3, location 5) are interpreted to be associated with a continuous subsidence of depocenter X, and coeval rise of the eastern margin and associated erosion. Despite the similar distribution of SP1-2 and SP1-3, within depocenter X, their thickness maps suggest a slightly different depositional
history. The SP1-2 thickness map shows two discrete minibasin entrances, one from northwest to southeast, and other north to south, with an evident area of maximum deposition displayed by dark gray along the eastern edge of the minibasin (Figure 2-16, SP1-2, location 3). These features were identified as well in the seismic facies interpretation of SP1-3, which corroborates the presence of two possible sediment entrances, along with a longitudinal possible reservoir facies distributed northeast to southwest lying where the thickest area is (Figure 2-17, location 4). Nevertheless, the geographic location of the thickest part of SP1-3 seems to be divided into two local depocenters, one small to the northeast and a larger second one to the southeast (Figure 2-16, SP1-3, locations 4). In addition, reservoir facies distribution corresponds directly to where the principal depocenters are located (Figure 2-17 SP1-3, location 5). A possible crevasse splay deposit is interpreted near location 6 based on its morphology (Figure 2-17 SP1-3). These differences observed between SP1-2 and SP1-3 suggest that the underlying salt layer migrated laterally from northeast to southeast due to a possible southward accommodation space increase. Both, SP1-2 and SP1-3 display a diminished thickness towards the western margin where strata terminate by onlap, and lack reservoir facies (Figure 2-15, S1-2 and S1-3, location 4; Figure 2-16 and Figure 2-17, SP1-2 and SP1-3, location 4). Also, the abrupt thickness and the most likely reservoir facies culmination on the eastern side supports the interpretation of erosional and structural truncation termination on that margin (Figure 2-15, S1-2 and S1-3, location 2 and 5; Figure 2-16 and Figure 2-17, SP1-2 and SP1-3, location 2 and 5).

SP1-4 contains strata bounded at the base by the S1-3, and above by the S1-4 (SB-4), and represents the last stage of Sequence 1 minibasin fill deposited during a TST and a HST (Chapter 2.3.5.1 Sequence stratigraphy of Sequence 1). At this point, local basinal topography was mostly filled and SP1-4 blankets the study area. Onlapping fronts are localized to the south-central part of
the study area. Very local erosional truncation fronts are present mostly over the northwest and south central (Figure 2-15, S1-4, location 6). It is important to notice that the eastern erosional truncation front present from SP1-1 through SP1-3, appears to be progressively shifting southward. It may imply that the erosion processes ceased as a result of a slow salt rise, but continued salt rise in the southern region (Figure 2-15, S1-4). Thickness map distribution of stratal termination style SP1-4 at the S1-4 indicate this surface is the upper SB of Sequence 1. Thickness fluctuations are not found on the central part of the minibasin, depocenter X, where SP1-2 and SP1-3 were previously deposited, but over the northwest and southeast region of SP1-4 (Figure 2-16, SP1-4, location 5). This suggests that depocenter X was by-passed and that sediments eventually accumulated where accommodation space was available due to minibasin subsidence similar styles are reported by Sinclair and Tomasso (2002). This concept of “fill and spill” is supported by the distribution of reservoir facies, which shows how sedimentation was primarily carried out and stacked over the peripheries of depocenter X (Figure 2-17, SP1-4, location 7).
Figure 2-15. Map of Sequence 1 stratal termination interpretation. A (S1-1), B (S1-2), C (S1-3) and D (S1-4) are sorted by older to younger respectively. Localities described in text.
Figure 2-16. Map of Sequence 1 thickness interpretation. Sequence 1. A (SP1-1), B (SP1-2), C (SP1-3) and D (SP1-4) are sorted by older to younger respectively. Localities described in text.
Figure 2-17. Map of Sequence 1 reservoir facies interpretation. A (SP1-1), B (SP1-2), C (SP1-3) and D (SP1-4) are sorted by older to younger respectively. Localities described in text.
2.3.6.3 Sequence 2

Sequence 2 is developed from the lower to the middle part of the Middle Miocene, and exhibits six principal depositional events defined by SP2-1 (SB-4) to SP2-6 (SB-5) (Figure 2-14). The general depositional trend of the Sequence 2 interpreted from the most likely reservoir facies orientation is northwest – southeast. The areal distribution of the early part of the Sequence 2 onlaps on the west against a shale-controlled structural high, and is erosionally and structurally truncated on the east by a salt-controlled structural high. However, the areal distribution of the late part of the Sequence 2 is erosionally truncated over the margins by rapid uplift of the shale- and salt-controlled structural highs. Local lateral shifts of thick successions are observed through time.

The earliest depositional event of Sequence 2 is SP2-1, which is bounded at base by S1-4 and the lower SB of Sequence 2 (SB-4), and at the top by S2-1. The next depositional event is defined by the SP2-2, limited at base by S2-1 and at top by S2-2. Both SP2-1 and SP-2-2 were deposited during a LST (Chapter 2.3.5.2 Sequence stratigraphy of Sequence 2). S2-1 and S2-2 are characteristically similar to S1-2 and S1-3 (Figure 2-15) because the western end onlaps against their corresponding lower SB (Figure 2-18, S2-1 and S2-2, location 1), and the eastern limit is mainly defined by both erosional and structural truncation (Figure 2-18, S2-1 and S2-2, location 2), with local onlap fronts. However, despite their similarity in a stratal termination, S2-1 and S2-2 are more extensive than S1-2 and S1-3 (compare Figure 2-15 and Figure 2-18). This wider coverage is probably related to an increase in accommodation space or an increase in the deposition rate in the area that resulted in S2-1 and S2-2 advancing over the edges of depocenter X and Y. S2-2 displays depositional progradation in comparison with S2-1 because of the outward displacement of the onlap front (Figure 2-18, S2-1 and S2-2, location 1). A considerable onlap front is developed in the northeast region of S2-2, where depocenter Y is located (Figure 2-18, S2-
2, location 3), which suggests that a paleo-high element may have existed over this area, which separated the main northeast-southwest depositional stream into two. In addition to the wider distribution of SP2-1 and SP2-2 observed from stratal termination interpretation, their thickness maps offer an additional understanding of depositional patterns. Thickness patterns show that some sediment entrance points into the minibasin controlled deposition of the turbidites, as also shown by the facies interpretation (Figure 2-19 and Figure 2-20, SP2-2, location 2). Sediments followed depositional paths from the northeast to the southwest, but locally, also from west to east, coinciding with where the depocenter is thicker compared to other zones (Figure 2-19, SP2-2, location 2). The northeastern region of SP2-1, depocenter Y, is consistently thicker than its southwest counterpart, depocenter X (Figure 2-19, SP2-1, location 1), implying that accommodation space to receive sediment was different through the deposition of it, generating a northeast depocenter. However, the distribution of the most likely reservoir facies seems to be mainly concentrated where SP2-1 is thinner in depocenter X, and with some sporadic overbank facies over the northeast portion, depocenter Y (Figure 2-20, SP2-1, location 1). The site of maximum deposition was shifted toward the central area of study, depocenter X, at SP2-2 time (Figure 2-20, SP2-2, location 3). Once again, thickness maps show that the minibasin could possibly tilt toward depocenter X, resulting in depocenter shifting, similar to SP1-2 and SP2-3, but on a larger scale. Both, SP2-1 and SP2-2 decrease in thickness (Figure 2-19, SP2-1 and SP2-2, location 4) and have more overbank facies component around their corresponding western edges (Figure 2-20, SP2-1 and SP2-2, location 3), suggesting that sediments were confined and pinch out against an underlying inclined surface.

SP2-3 represents strata bounded at the base by the S2-2, and above by the S2-3. SP2-4 is defined in the lower limit by the S2-3 and the upper limit by the S2-4. SP2-5 is lower and upper
limited by S2-4 and S2-5 respectively. SP2-3 to SP2-5 were deposited during a LST (Chapter 2.3.5.2 Sequence stratigraphy of Sequence 2). Contrary to the wider deposition of SP2-1 and SP2-2, the stratal termination interpretation from S2-3 to S2-5 show that SP2-3 to SP2-5 were mostly confined to the northeastern region, depocenter Y (Figure 2-18, S2-3 to S2-5). This constrained sedimentation pattern could possibly be derived from an increase in the accommodation space in depocenter Y. Sediment distribution axes coincide with the previous sequence, following a northeast-southwest trend (Figure 2-18, S2-3 to S2-5). However, the distribution of SP2-3, SP2-4 and SP2-5 is not preserved because of subsequent erosional processes stripping the edges as shown in S2-3 to S2-5 maps (Figure 2-18, S2-3 to S2-5, location 4). The syndepositional growth of a structural element in the central part of S2-3 through S2-5 may have been exposed this area to erosion (Figure 2-18, S2-3 through S2-5, location 5). A local northeast increase in the accommodation space rate during the deposition from SP2-3 to SP2-5 proposed by the stratal termination maps is reinforced by their corresponding thickness maps (Figure 2-19). Similar facies distribution (Figure 2-20, SP2-3 to SP2-3, location 4), morphology, and areas of maximum deposition of SP2-3 to SP2-5 show the thickest section is mostly located on the southeastern end of depocenter Y (Figure 2-19, SP2-3 to SP2-5, location 5). Thickness of SP2-5 displays how depocenter Y was probably segregated into two depocenters due to a local structural rise (Figure 2-20, SP2-5, location 5 and 6).

SP2-6 is bounded at the base by S2-5 and at the top by S2-6 (SB-5), and represents the stage when Sequence 2 was blanketed during a TST (Chapter 2.3.5.2 Sequence stratigraphy of Sequence 2). The interpretation indicates that the major stratal terminations are structural truncations, which suggests that sediment deposition happened throughout and beyond the study area during SP2-6 time (Figure 2-18, S2-6). Reservoir facies are relatively sparse across the area,
most of them likely are sediments captured by a northwest depocenter that may have starved depocenter X and Y of potential reservoir and overbank facies (Figure 2-20, SP2-6, location 5). The thickest SP2-6 is mainly concentrated in the peripheries of depocenter X and Y, which indicates that the entire area was sufficiently filled, that sediments bypassed the area or were captured by local lows, such as depocenter Z (Figure 2-19, SP2-6, location 7).
Figure 2-18. Map of Sequence 2 stratal termination interpretation. A (S2-1), B (S2-2), C (S2-3), D (S2-4) E (S2-5) and F (S2-6) are sorted by older to younger respectively. Localities described in text.
Figure 2-19. Map of Sequence 2 thickness interpretation. A (SP2-1), B (SP2-2), C (SP2-3), D (SP2-4) E (SP2-5) and F (SP2-6) are sorted by older to younger respectively. Localities described in text.
Figure 2-20. Map of Sequence 2 reservoir facies interpretation. A (SP2-1), B (SP2-2), C (SP2-3), D (SP2-4) E (SP2-5) and F (SP2-6) are sorted by older to younger respectively. Localities described in text.
2.3.6.4 Sequence 3

Sequence 3 spans the middle of the Middle Miocene to top Upper Miocene time interval, and is subdivided into five depositional events demarcated by SP3-1 (SB-5) to SP3-5 (SB-6) (Figure 2-11, Figure 2-12, and Figure 2-14). The general depositional trend of the Sequence 2 suggested from the most likely reservoir facies orientation is northwest – southeast, with a secondary northeast – southwest component. The distribution of Sequence 3 onlaps at west and east against shale- and salt-controlled structural highs. Local lateral shifts of thick successions are observed through time.

SP3-1 represents strata bounded at the base by the S2-6 or lower SB of Sequence 3 (SB-5), and above by the S3-1. SP3-2 is defined at the base by S3-1 and at the top by S3-2. SP3-3 is lower and upper limited by S3-2 and S3-3 respectively. SP3-1 to SP3-3 were deposited during a LST (Chapter 2.3.5.3 Sequence stratigraphy of Sequence 3). The areal limits of the three first stages of Sequence 1, SP3-1 to 3-3, become progressively wider from older to younger stratal packages by progressive onlap (Figure 2-21). The depocenter that contains SP3-1 and SP3-2 is defined by a northeast-southwest trend that remains consistent from sequences 1 and 2 (Figure 2-15, Figure 2-18, and Figure 2-21). S3-1 and S3-2 are limited on the west and east sides by onlap fronts indicating a stage of little structural activity, in which sediments terminated against an originally almost static inclined surface (Figure 2-21, S3-1 and S3-2, location 1). The eastern extent of S3-2 extends out of depocenter Y, suggesting partial sediment filling and spill of sediment into depocenter Z (Figure 2-21 S3-2, location 2). Additionally, during the deposition of S3-2, depocenter Z displays a north-south onlapping front, suggesting that sedimentation bordered a partially static paleo-high and was filled by “spilled” sediments derived from depocenter X (Figure 2-21, S3-2, location 3). The eastern and western margins of SP3-1 through SP3-3 display marked
thinning (Figure 2-22, SP3-1 to SP3-3, location 1), which in addition to the stratal termination interpretation and the lack of likely reservoir facies (Figure 2-23, SP3-1 to SP3-3), suggests that sediments onlap against the lower sequence boundary. Thickness trends from SP3-1 and SP3-3 indicates that depocenter X was the main area that captured sediments (Figure 2-22, SP3-1 and SP3-3, location 2), with some minor deposition in depocenter Y (Figure 2-22, SP3-1 and SP3-3, location 3). This corresponds with the potential reservoir facies distribution, which in general trends from northeast to southwest within depocenter X (Figure 2-23, SP3-1 to SP3-3, number 1), with a minor component northwest to southeast toward depocenter Y (Figure 2-23, SP3-1 to SP3-4, location 2). It is interesting that the thickness of SP3-2 is almost constant through the area. Its thickest section is close to the northern region of depocenter X (Figure 2-22 SP3-2, location 4) where the reservoir facies may continually discharge and stack sediment as fine-grained crevasse splays (Figure 2-23, SP3-2, location 3). This last point probably also means that the accommodation space formation was similar in depocenters X and Y during the deposition of SP3-2. SP3-3, hints at a more important eastern sediment outlet from the study area due to the diminution of the onlapping front through this region (Figure 2-21, S3-3, location 3). The continuous onlapping fronts throughout the western zone of S3-3 are interpreted to occur because of the static structural element remaining high and controlling the limits of deposition (Figure 2-21, S3-3, location 1).

Two interbasinal canyons, similar to those described by Satterfield and Behrens (1990); and Sinclair and Tomasso (2002) were conceivably developed during the deposition of SP3-3, and are identified by two indentations on the western onlap margin of S3-3 (Figure 2-21 S3-3, white arrows). These interbasinal canyons are also observed in the corresponding thickness map as a convex geometry where the outgoing of the interbasinal canyons are located (Figure 2-22, SP3-3,
white arrows). Due to the lower and high-density fraction of the sediment flow decelerates and stacks close to the outgoings of the interbasinal canyons, a thick succession was identified in the proximities of the this area in the thickness map of SP3-3 (Figure 2-22, SP3-3, location 5). The most likely reservoir facies close to the interbasinal canyon outlets of SP3-3 are oriented parallel to the interbasinal canyons, which suggests that the distribution of these reservoir facies was influenced by the northwest – southeast interbasinal canyon flow (Figure 2-23, SP3-3, location 4).

The overlying low-density flow portion is reflected and deflected, as proposed by Sinclair and Tomasso (2002), and corroborated by the longitudinal stacking of sediment through the eastern margin of SP3-3, and the thickness increment over depocenter Z (Figure 2-22 and Figure 2-23, SP3-3, location 6).

SP3-4 lower boundary is S3-3 and upper boundary is S3-4, and was deposited during a TST (Chapter 2.3.5.3 Sequence stratigraphy of Sequence 3). SP3-4 shows significant increase in accommodation space since most of the area seems to be filled (Figure 2-21, S3-4). The western margin blanketed by SP3-4 and onlaps remnants of the shale-controlled structural highs that had confined sediments from SP3-1 through SP3-3 (Figure 2-21, S3-4, number 4). On the opposite end, a mainly southeastern erosional truncation front (Figure 2-21, S3-4, location 5), with some onlapping and structural truncation, indicates that a significant tectonic activity occurred, which partially eroded SP3-4 over this region. The thickness of SP3-4 on depocenter X exhibits a narrow lanceolate shape tapering to a point at the northeast end, (Figure 2-22, SP3-4, location 8). This geometry suggests that sedimentation occurred from northeast to southwest, culminating in the central area of depocenter X, where the accommodation space rate in higher. This observation is reinforced by the distribution of the possible reservoir facies that are distributed from northeast to southwest (Figure 2-23, SP3-4, location 5).
The SP3-5 lower boundary is S3-4 and upper boundary S3-5 or upper SB of Sequence 3 (SB-6), and was deposited during the HST (Chapter 2.3.5.3 Sequence stratigraphy of Sequence 3). Sediment deposition probably continued beyond the study area during this period due to the lack of onlapping or erosional truncation terminations within the study area (Figure 2-21, S3-5). Also, the thickness map indicates that depocenter X and Y still were generating enough accommodation space to capture sediments, but probably were insufficient to have capacity for more sediments, and got bypassed (Figure 2-22, S3-5, location 8). Additionally, there are almost no reservoir facies in the area, except in the north (Figure 2-23, S3-5, location 5), supporting the interpretation that SP3-5 is the upper SB of Sequence 3.
Figure 2-21. Map of Sequence 3 stratal termination interpretation. A (S2-1), B (S2-2), C (S2-3), D (S2-4) and E (S2-5) are sorted by older to younger respectively. Localities described in text.
Figure 2-22. Map of Sequence 3 thickness interpretation. A (SP2-1), B (SP2-2), C (SP2-3), D (SP2-4) and E (SP2-5) are sorted by older to younger respectively. Localities described in text.
Figure 2-23. Map of Sequence 3 reservoir facies interpretation. A (SP2-1), B (SP2-2), C (SP2-3), D (SP2-4) and E (SP2-5) are sorted by older to younger respectively. Localities described in text.
2.3.7 Conclusions

Geophysical and geological methodologies were used to interpret well logs, which helped to create a framework for understanding the nature and trends of the lithologies deposited in the area through time. Even though the main lithologies observed from the well log signatures are shales to silts, slight variations, in grains size, content of radioactive clays, capability to conduct the electricity current, density and hardness, permitted identification of trends. The gamma ray and resistivity well log are interpreted to show that long and short term increases and decreases in accommodation space influenced the fluctuations in the lithological contents of the minibasin. Additionally, long and short term compaction trends were identified using the density and sonic well logs.

The stratigraphic observations suggested that the Sequence 1 was developed through a LST, TST, and HST, and limited at the bottom by SB-1 and SB-3, and the top by SB-4. The LST deposited sediments mainly in depocenter X in a progradational stacking pattern. The TST and HST covered the entire area with retrogradational and progradational stacking pattern strata respectively. The Sequence 2 was dominated by the LST, which focused the deposition over depocenter Y, a TST that blanket the minibasin, and limited at the bottom by SB-4, and the top by SB-5. Finally, Sequence 3 is bounded by the SB-5 at the bottom and by the SB-6 at the top, contained a LST that carried sediment to depocenter X and Y in the same magnitude, a TST that as well as the HST covered the entire area of study in a retrogradational and progradational staking pattern respectively.

One of the most significant methodologies applied in this chapter was the detailed mapping of the stratal terminations, which was an important tool to outline the closet to reality lateral distribution of each stratal package, and to accurately delineate the controls on the distribution of
Sequence 1, Sequence 2, and Sequence 3. Stratal terminations, in conjunction with the thickness and the most likely reservoir facies mapping led to the understanding of the areal distribution, morphology, architecture, and structural relationship of depocenters in the area through time. The principal depositional trend is northeast – southwest, and was inferred from their areal coverage and the most likely reservoir facies orientation through all three sequences. However, during the deposition of the Sequence 3, the presence of two intrabasinal canyons added a northwest – southeast entrance point component to the main NE-SW depositional trend.

Substantial tilting of the minibasin through time focused deposition over certain areas. The oldest interpreted stage, Sequence 1, was mainly deposited in the deeper and central part of depocenter X, onlapping on the west against the shale-controlled structural high, Sequence 1 was erosionally truncated on the east by the salt-controlled to the structural high. Sequence 2 was concentrated in depocenter Y, and limited by erosional truncations at its margins due to a rapid uplift of the shale- and salt-controlled structural highs. Finally, the distribution of the Sequence 3 through depocenter X and Y, suggests that the structural highs were slowly uplifting. In addition to the major depocenter shifts through the Sequence 1, Sequence 2, and Sequence 3, local depocenter shifts were observed from their composing stratal package thickness maps, suggesting small lateral variations of subsidence.

The upper limit of all sequences was marked by regional stratal packages thatblanketed depocenter X and Y, and deposited sediments over their peripheries (depocenter Z), which were interpreted as the stages of maximum accommodation space available, and eventual upper sequence boundary. It was observed that the northeast – southwest distribution trend of the earlier SP of each sequence shifted to northwest – southeast during the last SP of each sequence. This depositional orientation shift was due to the active depocenter that captured sediments during the
LST was blanketed up to its structurally higher margins during the last SP, and the sediment distribution was more influenced by the regional northwest – southeast depositional tendency of the ancestral Rio Bravo, Rio Grande, and Guadalupe rivers.

The accommodation space creation and the regional sediment source are two factors that controlled depositional into the study area. The decrease of the space available to capture sediment during the LST confined the sediments into the sub basins that were active through time (depocenter X and/or Y) between the shale- and salt-controlled structural highs. However, once the space available was increased due to TST or HST, the regional sediment source was the main controlling factor that distributed the sediments because of the diminished influence of the local structural highs.
Chapter 3: Palinspastic restoration

3.1 Palinspastic restoration
This section of this chapter is focused on the palinspastic restoration of Section A, using the methodology and assumptions made from regional studies (Salomón-Mora, 2013), and the correspondent evolutionary stages model proposed by Rowan (2002) and Hudec et al. (2003).

3.1.1 Introduction and shortening model
This section introduces the methodology and evolutionary model on which the palinspastic restoration was based. This analysis was focused on distinguishing what structural elements influenced the deposition of the Sequence 1, 2, and 3, as well as quantifying the amount of deformation due to shortening or extension through time.

The methodology followed the iterative and sequential steps described by Salomón-Mora (2013):

1) Import the cross section on the 2D kinematic modeling of Move™ 2016.
2) Create a database of horizons and establish a nomenclature (Table 3-1).
3) Apply decompaction and isostatic corrections according to defined parameters.
4) Generate a regional level as a reference for every stage of restoration.
5) Restore extensional stages applying the vertical shear algorithm.
6) Restore contractional stages applying the flexural slip algorithm.
7) Validate structural interpretation and reinterpret if required.
8) Establish a final model and continue restoration of units from point 4.

Steps 3 to 8 are sequentially applied after the final model has been generated for each progressively deeper deformed units (Salomón-Mora, 2013).
The structural assumptions utilized to understand the processes that led to the initiation and development of the minibasin are those proposed by Rowan (2002) and Hudec et al. (2003). Their hypothesis is that various minibasins located in deep water settings were initiated by compressional stress, allowing the sediments that fill the minibasin to subside, even though they are not denser than salt, but are stronger and more difficult to deform. Hence, any ductile material (salt and/or shale) beneath the minibasin is less affected by deformation than the adjoining strata, resulting in uplifted massifs, and subsiding minibasins.

Hudec et al. (2003) proposed a series of guidelines to discern between compressional topography and topography generated by gravitational subsidence:

1) Thrust faults and folding may form during shortening at the bottom of minibasins.

2) Differential compressional rise of adjacent massifs may be suggested by depocenter shifts within a minibasin.

3) Even though the minibasin is not thicker and denser than salt, the center of a minibasin is defined by a topographic low during shortening.
Constant thickness strata and local depocenter thick may exist in preshortening stage (Figure 3-1, a). During the following period, the thrusting stage, compressional structures are generated at the bottom of the minibasins deforming salt roof layers (Figure 3-1, b), until: (1) the sediment filling the minibasin are thicker and stronger, and harder to shorten, (2) top-salt relief is large enough to drive shortening into the adjoining salt bodies, or (3) shortening stops (Hudec et al., 2005). At this time, deposition of the most likely reservoir rock facies occurred preferentially in the synclines between thrust anticlines (Montoya, 2006). Eventually, during compressional folding, shortening is pulled apart into the adjacent ductile bodies, which then inflate to fold the minibasin peripheries (Hudec et al., 2005), allowing erosional truncation of pre-existing strata on the uplift highs (Figure 3-1, c). The reservoir facies are typically deposited during this stage because the relief of the minibasin is more regular (Montoya, 2006). Ultimately, the minibasin fill becomes thick and dense enough to gravitationally sink and stop the shortening, which is replaced by isostatic subsidence into the salt (Figure 3-1, d).
Figure 3-1. Stages of subsidence in compressional minibasins (modified from Hudec et al., 2005)
3.1.2 Evolutionary model

This presents the structural history of the study area based on the sequential restoration and structural analysis of the sequence boundaries in Section A. This analysis utilized the compressional minibasin model proposed by Hudec et al. (2005).

When restoring the salt and shale related structures, some inferences were taken into account. One of the most significant assumptions was that latest Paleogene successions represent the prethrusting strata. There is no substantial data to calibrate their stratigraphy, and therefore thrusting may have begun prior to this. Another important assumption is that ductile material formed a continuous layer beneath the prethrusting strata. This is assumed to be a paleo-canopy of salt emplaced into Paleogene sequences that eventually was withdrawn and emplaced into younger Neogene strata, similar to those proposed by Cruz et al. (2010) and Salomón-Mora (2013).

3.1.2.1 Prethrusting stage

The restoration of the prethrusting stage was performed using flexural slip algorithm in Move™ 2016 due to its compressional structural context. The rocks that are part of the prethrusting stage are below sequence boundary 1 (SB1). This strata thickness is assumed to be nearly constant and continuous throughout the area. However it is clear that an early local depocenter developed above the ductile material of the western shale and eastern salt areas (Figure 3-2). During this period there is no quantifiable deformation due to lack of structures and thickness changes.

3.1.2.2 Thrusting stage

The thrusting stage was also restored using flexural slip algorithm Move™ 2016 due to its compressional structural context. The rocks of the thrusting stage are defined by those limited by the sequence boundary 2, 3, and 4 (SB2, SB3, and SB4 respectively). This last sequence boundary (SB4) corresponds to the upper limit of the Sequence 1.
Shortening of 0.31 km of SB2 and 0.15 km of SB3, and generation of a depocenter above the shale massif occurred over the western margin during the deposition of SB2, and SB3 (Figure 3-4). Eventually, at SB 4 or Sequence 1 deposition, the main depocenter was developed on the synclines that formed between the middle-late Oligocene anticlines (depocenter X and partially Y) with a shortening of 0.6 km (Figure 3-4). The total of shortening of the thrusting stage is about 1.06 km (Figure 3-4).

3.1.2.3 Compresional folding stage

The compresional folding stage occurs within a compressional structural framework, it was restored using the flexural slip algorithm. The rocks of the compressional folding stage are defined by those limited by the sequence boundary 5, 6, and 7 (SB5, SB6, and SB7 respectively). Sediments below SB5 correspond to Sequence 2 strata, and underneath SB6 to Sequence 3.

Shortening continues through the compresional folding stage, but is now partitioned into the adjacent ductile bodies. The edges of the minibasin are folded due to inflation of the adjacent salt and shale diapirs (Figure 3-5). The depocenter of Sequence 2 (SB5, depocenter Y) shifts from the earlier depocenter in X during Sequence 1 (SB4, X) probably due to an increment in the uplifting rate over the relatively thin margins of the sub basins. Middle-late Oligocene to early Miocene strata were shortened by 0.17 km resulting in folding of the minibasin periphera, and tilting of the minibasin northeastward (Figure 3-5). The restoration of SB6, at the top of Sequence 3, shows that depocenters X and Y were active, separated one from each other by the folded margins of depocenter Y. This folding resulted in a shortening of 0.34 km (Figure 3-5). The restoration of SB7 indicates that depocenter Y was filled, causing deposition to focus on depocenter X (Figure 3-5). The folding of depocenter Y during SB7 time produced by an additional
0.03 km of shortening. The total of shortening of the compressional folding stage is about 0.54 km, with a cumulative shortening of 1.6 km since the prethrusting stage (Figure 3-4).

3.1.2.4 Isostatic subsidence stage

The final episode, isostatic subsidence, occurred after the shortening resulting from the thrusting and compressional folding ceased, and deformation became extensional. For this reason, the isostatic subsidence was restored using the vertical shear algorithm in Move™ 2016. The isostatic subsidence stage is defined by those sediments between sequence boundaries 8, 9, 10, and the current sea floor (SB8, SB9, SB10, and SF respectively).

Thinner, less dense, semi-compacted sediments rise near the edge of the minibasin, forming a bathymetric rim around it (Hudec et al., 2003) (Figure 3-3). Denser sediments in the basin center subside into underlying salt. This is probably the case of the deposition that occurs from SB8 through the sea floor. The depocenter during this event was in X rather than in Y and Z (Figure 3-6). Due to depocenter Z was continuously capturing sediment, in addition to the salt withdrawal, there was progressive extension of 0.25 km during this stage.

A summary of all the restored stages is showed by Figure 3-7.
Figure 3-2. Structural restoration of strata concerning to the sequence boundary 1, and its corresponding prethrusting evolutionary stage model.

Figure 3-3. Explanation for uplifted minibasin flanks (Hudec et al., 2003)
Figure 3-4. Structural restoration of strata concerning to the sequence boundary 2, 3, and 4, and their corresponding thrusting evolutionary stage model.
Figure 3-5. Structural restoration of strata concerning to the sequence boundary 5, 6, and 7, and their corresponding compressional folding stage model
Figure 3-6. Structural restoration of strata concerning to the sequence boundary 8, 9, and 10, current seafloor (SF), and their corresponding isostatic subsidence stage model
Figure 3-7. Restoration summary of section A.
3.1.3 Deformation

From the structural analysis of each evolutional stage of the palinspastic restoration, several observations and conclusions were made regarding the deformation. Cumulative shortening during the thrusting and compressional folding stages from SB2 to SB7, is incremental through time, and resulted in a total of 1.6 km of shortening, which represents a 6% contractional deformation. Other provinces bordering the Gulf of Mexico show similar amounts of shortening. The La Popa basin to the south of the Sabinas Basin was shortened more than 2 km (Rowan et al., 2003), which represents a contraction of 11% (Rowan and Vendeville, 2006). The Perdido fold belt was shortened by 5 km to 10 km (Trudgill et al., 1999), which represents a maximum of 15% of shortening (Gradmann et al., 2009). The shortening of the Miocene strata of the Mexican Ridges ranges from 2.4 km to 10.9 km, which in turn means a 1.6% to 6% of contraction (Salomón-Mora, 2013).

The total of cumulative extension during the isostatic subsidence in the study area, from SB8 to SF, is 0.25 km (Figure 3-8), which represents a 1% extensional deformation. However neighboring geological provinces present a larger extension, such as the Mexican Ridges, where it is as much as 34 km (63.9%) for Miocene sequences (Salomón-Mora, 2013).

![Cumulative shortening and extension of section A.](image-url)
It is interesting to notice how early depocenter X generated accommodation space during the Sequence 1 (SB4), at the end of the thrusting stage, during the period of most rapid shortening (Figure 3-9). Another peak of contraction occurred during the deposition of Sequence 3 (SB6), probably due to the increase in accommodation space creation on the platform that led to deposition of more sediments, which became denser and withdrawn the underneath salt (Figure 3-7 and Figure 3-9).

The resulting deformation generated by extension in progressively increased up to 0.35% of the total section length (Figure 3-9).

![Deformation](image.png)

**Figure 3-9.** Percentage of deformation of each sequence boundary and total deformation.

### 3.2 Conclusions

The palinspastic restoration of Section A along with the compressional minibasin model of Hudec et al. (2005) developed an understanding of how the structural configuration of the minibasin developed through time. A prethrusting stage was observed during the SB1 period, in which the thickness of the strata might be constant. Later, a thrusting stage was identified from SB2 to SB4 or Sequence 1, which formed the bottom of the minibasin, permitting differential
deposition into the synclines from between thrust anticlines. Thereafter, the compressional folding stage was developed from SB5 or Sequence 2, through SB6 or Sequence 3, until SB7, in which shortening was partitioned into the adjacent shale- and salt-controlled structural highs, which inflated to fold the edges of depocenter X and Y. Finally, the isostatic subsidence stage from SB8 to the recent formed when the minibasin became thick and dense enough to sink gravitationally and consequently shortening ceased. The contractional phase of the minibasin history recorded a cumulative 1.60 km of shortening, with two peaks of deformation during the Sequence 1 and Sequence 3 of 2.28% and 1.34% respectively. The approximate cumulative extension measured in the minibasin was 0.25 km, with a major deformation of 0.35% at present time.
Chapter 4: Tectonostratigraphic controls on evolution

In Chapter 1 an overview of the area was developed in conjunction with a structural description of the main regional evolution stages that developed the GoM, as well as the geographic location and importance of the Cenozoic fluvial input axes that provided sediment to the GoM during this period of time. Chapter 2 described the principal lithologic and stratigraphic characteristics of basin fill in the study area utilizing electrofacies and seismic analysis in order to interpret the areal distribution, thickness and reservoir rock facies characterization of three depositional sequences (Sequence 1, Sequence 2, and Sequence 3). These data were then used to develop a structural analysis and summary palinspastic restoration in Chapter 3. This chapter provides an integral analysis and discussion of the local and regional tectonostratigraphic controls that influenced the evolution of the area of study. The controlling factors are summarized and illustrated utilizing a series of schematic models from significant phases of Sequence 1, Sequence 2, and Sequence 3 that were built utilizing the geological information and details obtained on previous chapters.

4.1 Sequence 1, Late Oligocene – lower Middle Miocene

According to the electrofacies signature of the well A logs, Sequence 1 is characterized by the coarsest-grained sediment deposition in the area. Sequence 1 is associated with a regional LST that occurred starting roughly 30My during Chattian stage of the Late Oligocene (Haq et al., 1987) (Figure 4-2). Associated with this LST the Houston-Brazos and Mississippi prograded onto the northern GoM shelf, and the Rio Bravo and Rio Grande onto the western GoM shelf (Galloway et al., 2011). This forced regression commenced the Late Oligocene – Recent linked system of the evolutionary stage D3-N (Cruz et al., 2010) (see chapter 1.4), in which the abrupt, elevated sediment load deposition onto the shelf initiated gravitational downslope movement. The sediment
load generated extensional faults on the platform and influenced strata from the transition zone in the western upper slope, and concomitant downslope shortening in areas such as the Perdido Fold Belt on the lower slope of the GoM, (Fiduk et al., 1999; Trudgill et al., 1999; Cruz et al., 2010) (Figure 4-1).

Even though the regional depositional trend of the western GoM is from northwest to southeast (Galloway et al., 2011), the general orientation of basin fill within the study area subbasins follow the orientation of the depocenter axis, which is northeast – southwest. The sequence 1 local depositional axis was northeast – southwest (Figure 4-3) within depocenter X. This perpendicular trend of local deposition coincides with initiation of the Late Oligocene – Recent linked shortening and subsidence. Shortening in the area corresponded to the uplift of shale-controlled structural highs oriented perpendicular to the regional depositional trend. These highs served as barriers to sediment direct input from the shelf whereby sediments were diverted around the highs (Figure 4-3). The deposition of Sequence 1 in the study area preferentially took place in depocenter X, in the intra basin synclines between anticlines. Anticlines / synclines resulted from the approximated 1.06 km of northwest to southeast shortening in the area. Sequence 1 onlaps against the margins of the shale-controlled structural high at west, and is erosionally truncated on the eastern margin due to the rapid uplift of the salt-controlled structural high (Figure 4-3).

The subsequent stage in Sequence 1 was characterize by a retrogradational stacking pattern of seismic reflectors produced by a regional sea level rise during the TST, which progressively flooded the area with fine-grained sediments until reaching the maximum condensation stage in the MFS (Figure 4-2). The following HST brought slightly coarser-grained sediment associated with coarsening upward successions (Figure 4-2).
Figure 4-1. Regional 30 My unconformity over the Perdido Fold Belt, GoM (adapted from Fiduk et al., 1999)
Figure 4-2. Correlation of the local sequence chronostratigraphy of the minibasin to the Late Cenozoic portion of the sea-level record (adapted from Haq et al., 1987).
Figure 4-3. Schematic model of the Sequence 1 tectonic and stratigraphic depositional controls.
4.2 Sequence 2, lower – central Middle Miocene

The Early Miocene – Recent linked system of the D3-N evolutionary stage (Cruz et al., 2010) (see chapter 1.4) commenced with a regional short term sea level fall during the Middle Miocene. This is why Sequence 2 was deposited in a LST (Figure 4-2). However, the long term eustatic curve of Haq et al. (1987) shows this time period to be a relative high sea level globally. High sea level kept the Rio Bravo and Rio Grande fluvial outputs far from the study area resulting in an overall very fine-grained system compared to Sequence 1 and Sequence 3 (Figure 4-2). The overall orientation obtained from the most likely reservoir facies indicates that the depositional trend of the Sequence 2 still continues northeast – southwest due to the uninterrupted northwest-southeast shortening, which accumulated 1.23 km of deformation (Figure 4-5).

One of the most relevant features of Sequence 2 deposition is that the main area that captured sediments was focused over depocenter Y rather than in depocenter X like Sequence 1 (Figure 4-5). The reason why the sedimentation shifts toward the north is due to shortening, which raised the shale- and salt-controlled structural highs in the south faster, reducing to zero the accommodation space and tilting the minibasin to the north (Montoya, 2006) (Figure 4-4 and Figure 4-5). This rapid uplifting resulted on the erosional truncations of the Sequence 2 depositional margins (Figure 4-5).

The well log in Well A only documented the TST part of the Sequence 2, so interpretations are less well controlled, however it is believed that the study area generated enough accommodation space by a combination of a long term sea level rise (Figure 4-2) and a slow isostatic uplift of the shale- and salt-controlled structural highs.
Figure 4-4. Technique for understanding depocenter shifts. (a) Time 1: Minibasin is under shortening and depocenter shift to the south indicates more rapid uplift of the northern massif. (b) Time 2: Inverse settings causes depocenter shift to the north indicating more rapid uplift of the southern massif. (c) Time 3: Depocenter stays in the same previous place and rims form in minibasin flanks, indicating slow static subsidence (modified from Montoya, 2006).
Figure 4-5. Schematic model of the Sequence 2 tectonic and stratigraphic depositional controls.
4.3 Sequence 3, middle Middle Miocene – Upper Miocene

The deposition of the Sequence 3 was related to the regional Late Miocene – Recent linked system of the D3-N event proposed by Cruz et al. (2010), which in turn was influenced by the major short and long term falling of the mean sea level during a LST through the Serravalian and Early Tortonian stages of the Middle and Upper Miocene respectively (Haq, 1986) (Figure 4-2). This event produced a decrease in the accommodation space on the platform allowing the sediment sources to prograde into the Gulf of Mexico, the Rio Bravo, Rio Grande, and partially the Guadalupe River were the only significant contributors (Galloway et al., 2011) (see chapter 1.5.4). However, capture of the coarser grained facies closer to shore, resulted in transport of only fine-grained sediments (Figure 4-2) into the study area due to its remoteness. As well as for Sequence 1, and Sequence 2, the distribution of sediments corresponding to the Sequence 3 has a northeast – southwest orientation due to the remnant shale- and salt-controlled structural highs, which still confined the sedimentation (Figure 4-7). However, two interbasinal canyons, similar to those described by Satterfield and Behrens (1990) and Sinclair and Tomasso (2002) (Figure 4-6) incised the outer eastern face of the shale-controlled structural high, and provided a secondary sediment source into the minibasin. These, canyons influenced the depositional trends, orienting them slightly more northwest to southeast (Figure 4-7). The interbasinal canyons were related to the reducing rate of accommodation space creation and the slow uplift of the shale- and salt-controlled structural highs (Figure 4-4 and Figure 4-7). Once the minibasin fill was thick and dense enough, it gravitationally sank, which resulted in that the lateral and vertical distribution of sediments occurred over depocenter X and Y in the same proportions (Figure 4-7).
Eventually, the minibasin was filled with fining upward stacked strata during the TST that took place into the area due to a short term rise of the sea level (Haq et al., 1987) (Figure 4-2) covering the slow uplifting remnants of the shale- and salt-controlled structural highs.

The final period of Sequence 3 is characterized by a HST derived from a short term rising of the mean sea level at the end of the Messinian stage of the Upper Miocene (Haq et al., 1987), characterized by a coarsening upward and progradational stacking pattern, and a wide distribution over the area (Figure 4-2).

Figure 4-6. Flow bypass stage of the depositional model for the progressive infill of a confined turbidite basin and associated deposits at the base of the slope of a lower basin (modified from Sinclair and Tomasso, 2002).
Figure 4-7. Schematic model of the Sequence 3 tectonic and stratigraphic depositional controls.
4.4 Conclusions

Sequence 1 was developed due to a regional short term mean sea level fall and rise, that initiated the corresponding Late Oligocene – Recent linked system of the D3-N evolutionary stage. This event resulted in the deposition of medium- to fine-grained sediments mostly carried by the Rio Bravo and Rio Grande. Deposition occurred mainly into depocenter X during a LST preferentially northeast - southwest over the synclines in between the thrusting Middle to Late Oligocene anticlines. Local differential uplift rates of the shale- and salt-controlled structural highs were also the reason of the deposition through depocenter X.

Sequence 2 was influenced by a regional short term mean sea level fall and rise, that initiated the corresponding Early Miocene – Recent linked system of the D3-N evolutionary stage. This event resulted in the deposition of fine-grained sediments mostly carried by the Rio Bravo and Rio Grande. Deposition occurred mainly during a LST, preferentially northeast - southwest through depocenter Y due to local differential uplifting rates of the shale- and salt-controlled structural.

Sequence 3 was deposited due to a regional short term mean sea-level fall and rise, that initiated the corresponding Late Miocene – Recent linked system of the D3-N evolutionary stage. This event resulted in the deposition of fine grained sediments mostly carried by the Rio Bravo, Rio Grande, and Rio Guadalupe. Deposition occurred during a LST preferentially northeast - southwest over depocenter X and Y due to local slow uplifting rates of the shale- and salt-controlled structural.

Although most LST are modeled as being coarse grained, those developed during the deposition of the Sequence 2, and Sequence 3 were mainly formed of fine-grained sediments. Sequence 1 contained the relative coarser-grained portion of the section recorded in well A. This
high content of shales is due to the lack of input from the closest sources to the area, the Rio Bravo, Rio Grande and Guadalupe, which formed small delta systems from the Oligocene to Recent. Additionally, the remoteness of the sources to the study area influenced the grain size of the sequences filling the studied minibasin due to the high-density portion of the regional sediment stream being captured by topographic lows between the source and the study area. For these reasons it is advantageous to explore closer to the output of the paleoriver-sources, such as the Rio Bravo, Rio Grande and Guadalupe rivers, and vertically look for the basal section of the LST periods in order to reach better conditions in terms of reservoir rock. Thus, the uncertainty could be reduced, and the geological exploration risk of the reservoir rocks better assessed.
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Vita

Felix Alan Ramirez Cuellar was born in Poza Rica, Veracruz, Mexico on May 2, 1986. After completing his high school in Poza Rica, Veracruz, Mexico, in 2004, he attended the Instituto Tecnológico de Ciudad Madero, where he actively participated in local and regional creativity contests, and received the degree of Bachelor of Sciences in Geosciences Engineering in October 2008. In November 2008, he began his career in petroleum exploration at Grupo de Apoyo Profesional de México S.A. de C. V working in the Northern Deepwater Exploration Managerial, Perdido Fold Belt projects of Petroleos Mexicanos (Pemex). After a year of prolific technical development, he joined Pemex to continue working in the Perdido Fold Belt projects. Over almost 8 years of experience in the hydrocarbon exploration industry as a seismic interpreter, he has dynamically participated from regional integrated petroleum system analysis to local prospect evaluation and risk assessment to detailed initial characterization and reservoir delineation. As part of the activities developed in Pemex, he has constantly mentored recently graduated professionals. In fall 2014, he entered to the Department of the Geological Sciences of The University of Texas, at El Paso (UTEP) sponsored by Pemex in order to obtain his MS degree in Geological Sciences in spring 2016. During this period, he was volunteer mentor of the 2015 UTEP Department of Geological Sciences IBA team, which took the 1st Place in the Southwest Section AAPG Imperial Barrel Competition, and official mentor of the 2016 UTEP Department of Geological Sciences IBA team, which took the 1st Place in the Southwest Section AAPG Imperial Barrel Competition. He married Karen Estefany Garcia Lezama in 2013.

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