Cambridge University Press
978-1-108-41902-4 — The Gulf of Mexico Sedimentary Basin
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Part I Introduction

Part I is the foundational introduction to the Gulf of Mexico basin. It provides a detailed description of the unique tectonic setting that is necessary to understand how depositional trends emerge and evolve within the basin. This includes analysis of 10 basin-scale cross-sections across the USA, Mexico, and Cuba, onshore to offshore. What follows is a robust discussion of the Gulf of Mexico tectonostratigraphic framework, including stratigraphic terminology for the Mesozoic and Cenozoic strata and explanation of depositional systems classifications for the ancient carbonate and siliciclastic domains.

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Chapter 1

Introduction
Tectonic and Stratigraphic Framework

1.1 General Setting

In this book, we describe the greater Gulf of Mexico (GoM) basin as extending from the coastal plain in the southern USA to the coastal plain of southern Mexico, the Chiapas and Tabasco region, and east across the Yucatán Platform to Cuba, the Florida Straits, and the Florida onshore area (Figure 1.1). The Gulf basin has a central abyssal plain that generally lies at 13 km depth (Bryant *et al.* 1991). The eastern Gulf floor is dominated by the morphology of the Late Quaternary Mississippi Fan.

The continental slope of the northern Gulf margin displays a bathymetrically complex morphology that terminates abruptly in the Sigsbee Escarpment to the west and merges into the Mississippi Fan to the east (Steffens *et al.* 2003). The hallmark of the central Gulf continental slope is the presence of numerous closed to partially closed, equi-dimensional, **slope**

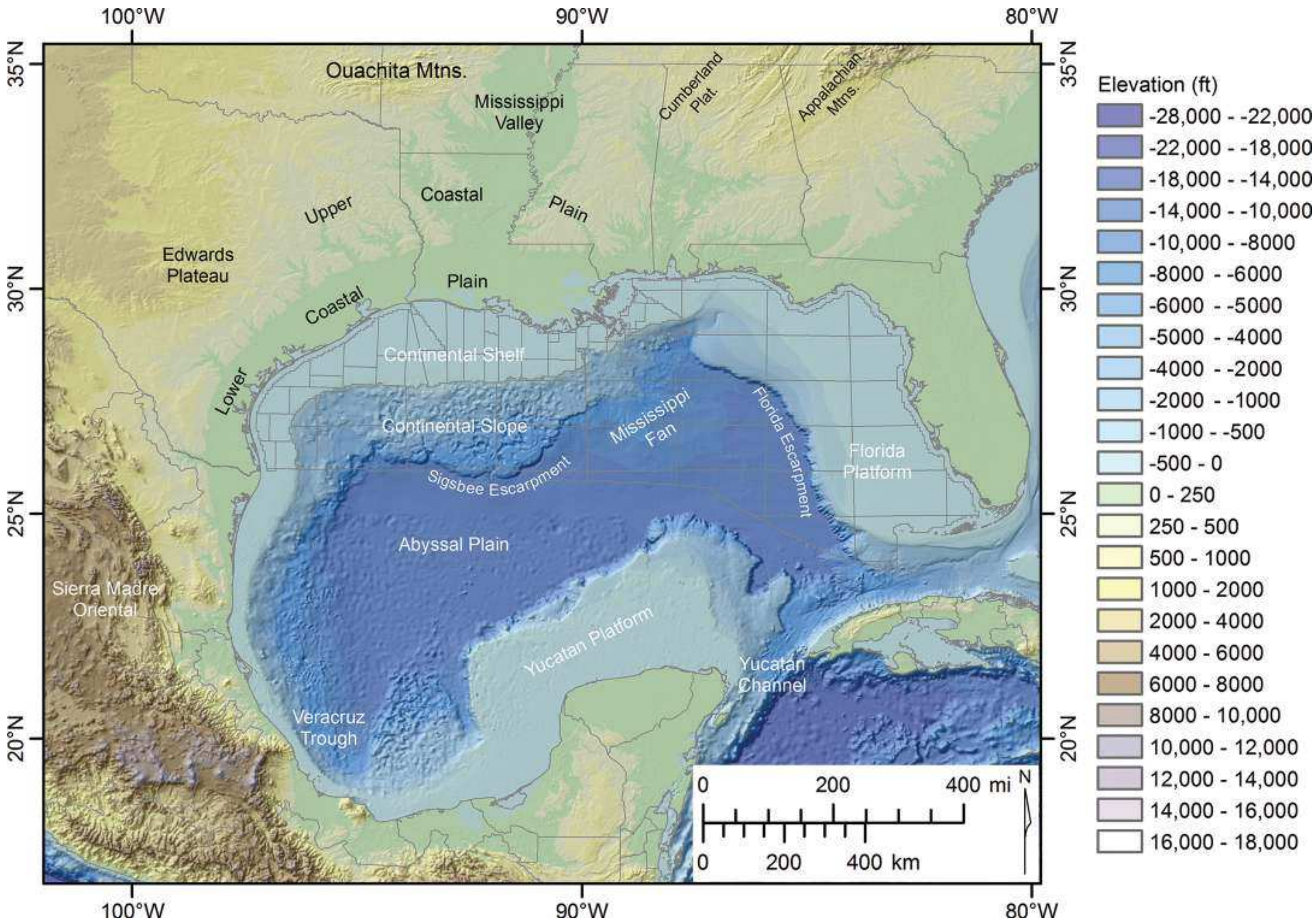

Figure 1.1 Location map for greater GoM basin, including important geographic and bathymetric features.

minibasins. In contrast, the Florida Platform forms a broad ramp and terrace that terminates at depth into the nearly vertical Florida Escarpment. The western Gulf margin displays intermediate width, and it too is quite bathymetrically complex. Here, numerous contour-parallel ridges and swales dominate the mid- to lower-slope morphology. The modern **shelf margin**, as reflected by a well-defined increase in basinward gradient, generally lies at a depth of 100–120 m. Landward, the northwestern, northern, and eastern GoM is bounded by broad, low-gradient shelves that range from 100 to 300 km in width (Figure 1.1). Today, and throughout its history, the Florida and Yucatán Platforms, which bound the basin on the east and south, persist as sites of carbonate deposition.

On shore, the northern and northwestern Gulf margins display a broad coastal plain (Figure 1.1). The lower coastal plain, a flat, low-relief surface, is underlain by Neogene and Quaternary strata. The upper coastal plain displays modest relief of less than about 100 m (328 ft) created by Quaternary incision into older Neogene, Paleogene, and Upper Cretaceous strata by numerous large and small rivers. The basin is bounded by a variety of Cenozoic, Mesozoic, and remnant Paleozoic uplands, including the Sierra Madre Oriental of Mexico, the Trans-Pecos mountains of west Texas, the Lower Cretaceous limestone-capped Edwards Plateau, Ouachita Mountains of southern Arkansas, and the Cumberland Plateau and southern Appalachian Mountains of northern Mississippi and Alabama. The northeast Gulf basin merges into the southern Atlantic coastal plain across northern Florida; however, the structural basin boundary is generally placed near the current west coast of the Florida peninsula.

Mexico's onshore topography strongly reflects the Sierra Madre Oriental in the north and the Chiapas deformational belts in the south of the country. The eastern onshore portion of Mexico is marked by short but steep gradient rivers that carry modern sediments toward a wave-dominated shoreline, a narrow shelf, and steep slope that terminates abruptly at the abyssal plain. Offshore, bathymetric maps show the sea floor complexity resulting from recent tectonic events: (1) the elongate, generally north–south oriented structures called the Mexican Ridges; and (2) the recent salt inflation and compression evidenced in the rugose hydrography of the Campeche and Yucatán salt provinces.

Across the Bay of Campeche lies the Yucatán carbonate platform, with equally steep margins that circumscribe the platform and its border with the adjacent Caribbean basin. The Yucatán channel separates Yucatán from Cuba, a tectonically complex mélange of various microplates that merged over 100 million years. Cuba lies across the Florida Straits from the South Florida basin, a short distance, but a world away in terms of its geological evolution.

1.2 Structural Framework

In order to understand the depositional evolution of the GoM, it is necessary to consider the structural framework that underpins and influences the sedimentary loading history of this immense natural repository. This extends to the deep crystalline crust and even mantle that can, in some cases, be detected by modern seismic reflection and refraction data. The accumulated sediment mass, including both siliciclastics and carbonates, also drove **gravity tectonics**, particularly where **evaporites** like salt respond in a ductile fashion at burial depths attainable by modern wells.

1.2.1 Deep Crustal Types

For many years, the form and lithology of the deep structure in the GoM was a matter of conjecture and inferences based upon rare penetrations of **basement** rock or sometimes-equivocal gravity and magnetic data. Recently, seismic refraction studies have greatly illuminated the form of the mantle and overlying crystalline and sedimentary crust (Van Avendonk *et al.* 2013, 2015; Christeson *et al.* 2014; Eddy *et al.* 2014). In addition, new plate tectonic models have altered previous suppositions on timing of basin opening and emplacement of **oceanic crust** (Norton *et al.* 2016). Alternative models, particularly for the pre-spreading rift phase, show convergence toward a consensus solution.

In general, these studies agree that the Gulf basin is largely surrounded by normal continental crust of the North American plate. Most of the structural basin is underlain by **transitional crust** that consists of **continental crust** that was stretched and attenuated by Middle to Late Jurassic rifting (Hudec *et al.* 2013a). Two types of transitional crust are differentiated (Figure 1.2). The basin margin is underlain by a broad zone of thick transitional crust, which displays modest thinning and typically lies at depths between 2 and 12 km subsea (Sawyer *et al.* 1991). The area of thick transitional crust

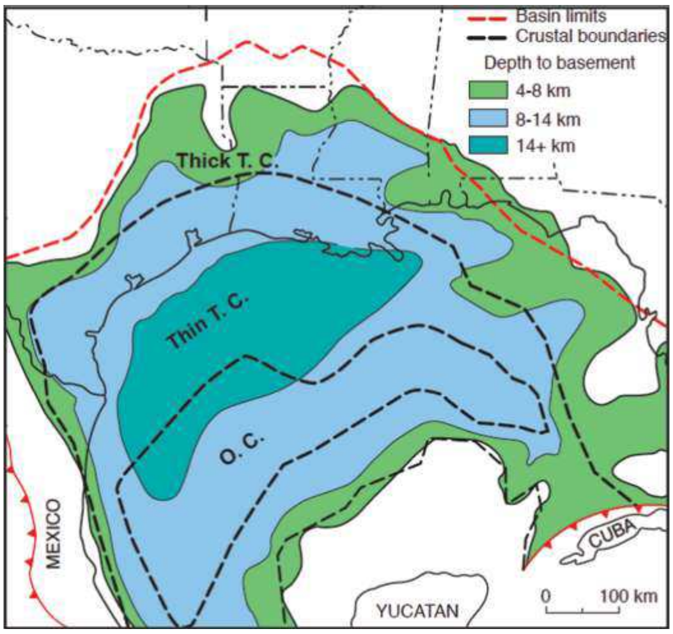

Figure 1.2 GoM crustal types. Modified from Galloway (2008).

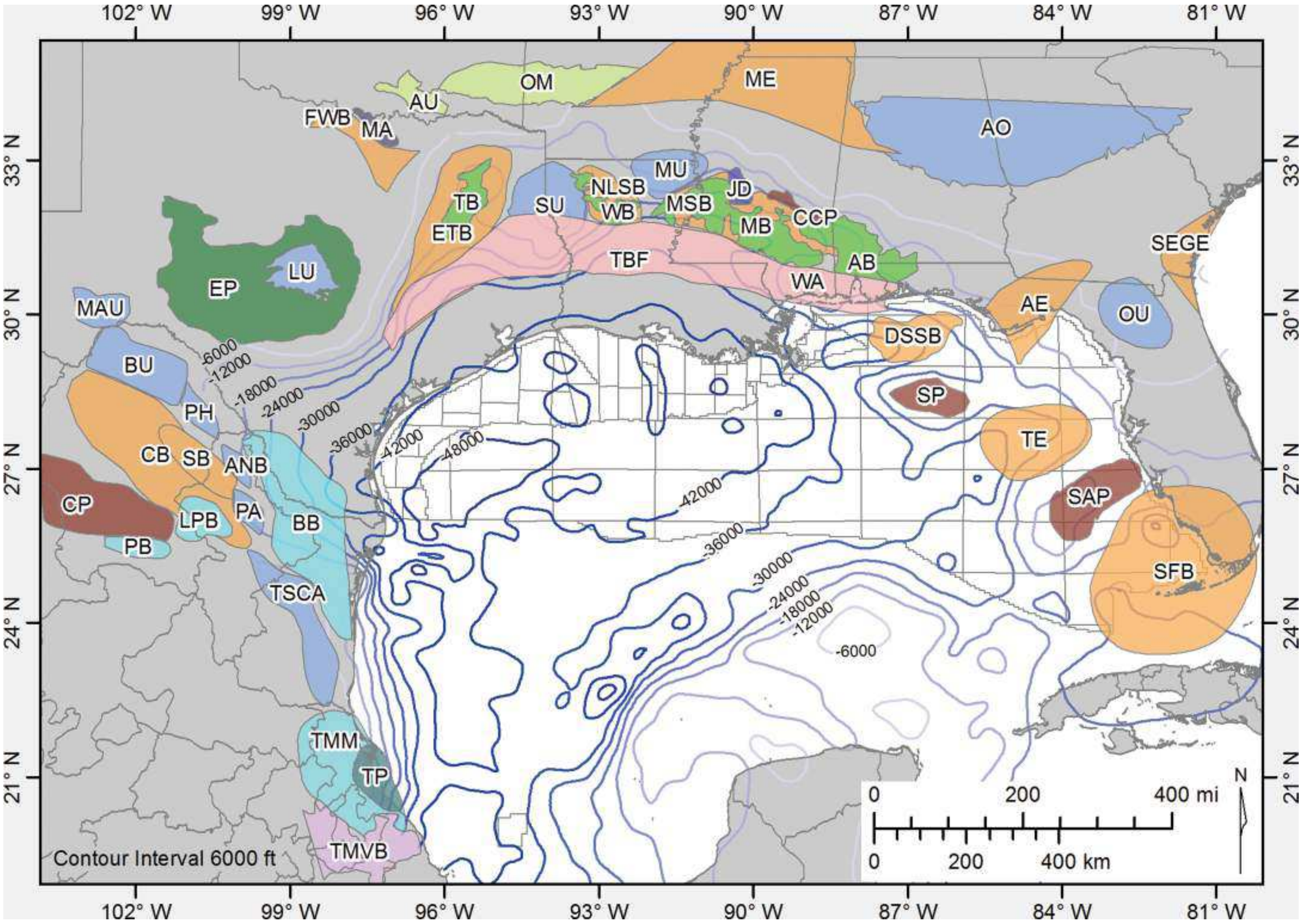

Figure 1.3 Key tectonostratigraphic features, northern GoM. Basement depths based on seismic structural mapping. Abbreviations: AB, Alabama basin; AE, Apalachicola Embayment; ANB, Anahuac Block, BB, Burgos basin; AO, Appalachian Orogen (Cretaceous limit); AU, Arbuckle Uplift; BU, Burro Uplift; CCP, Clarke County Platform; CP, Coahuila Platform; DSSB, DeSoto salt basin; EP, Edwards Platform; ETB, East Texas basin; FWB, Fort Worth basin; JD, Jackson Dome; LPB, La Popa basin; LU, Llano Uplift; MA, Muenster Arch; MAU, Marathon Uplift; MB, Mississippi Basin; ME, Mississippi Embayment; MSB, Mississippi salt basin; MU, Monroe Uplift; NLSB, North Louisiana salt basin; OM, Ouachita Mountains; OU, Ocala Uplift; PB, Parras basin; PH, Peyotes High; SAP, Sarasota Platform; SEGE, Southeast Georgia Embayment; SFB, South Florida basin; SP, Southern Platform; SU, Sabine Uplift; TB, Tyler basin; TE, Tampa Embayment; TMM, Tampico–Misantla–Magiscatzin; TP, Tuxpan Platform; TSCA, Tamaulipas/San Carlos Arch; WA, Wiggins Arch; WB, Winnfield basin. Terminology from various public sources, including Ewing and Lopez (1991).

consists of blocks of near-normal thickness continental crust separated by areas of stretched crust that has subsided more deeply. The result is a chain of named arches and intervening embayments and salt basins around the northern periphery of the Gulf basin (Figure 1.3).

Much of the present inner coastal plain, shelf, and continental slope is underlain by relatively homogeneous thin transitional crust, which is generally less than half of the 35 km thickness typical of continental crust and is buried to depths of 10–16 km below sea level. Reconstructions of deep seismic traverses (Peel *et al.* 1995; Radovich *et al.* 2007, 2011; Hudec *et al.* 2013b) indicate that basement may lie below 20 km in the central **depocenter** beneath the south Louisiana coastal plain and adjacent continental shelf. The deep, central Gulf floor is underlain by an arcuate belt of basaltic oceanic crust that was intruded during Late Jurassic through Early Cretaceous sea floor spreading (Hudec *et al.* 2013a; Norton *et al.* 2016). Surprisingly, the central Gulf crust generally lacks the magnetic signature typical of oceanic crust (Figure 1.4), which compounds interpretation difficulties, but recent gravity mapping (Sandwell *et al.* 2014) confirm earlier models of the location of the updip or landward limit of oceanic crust (LOC).

1.2.2 Seismic Refraction Studies of Deep Crust

The majority of data obtained for petroleum exploration is **seismic reflection data**, which allows both imaging through common depth point solutions and measurement of compressional seismic velocities to depths approaching 40,000 ft (12.2 km), depending on the energy source and cable. **Seismic refraction** data involves measurement of the compressional seismic velocities at much greater depths, approaching 40 km (25 miles). These velocities are a function of density in the deep earth and allow one to differentiate between mantle,

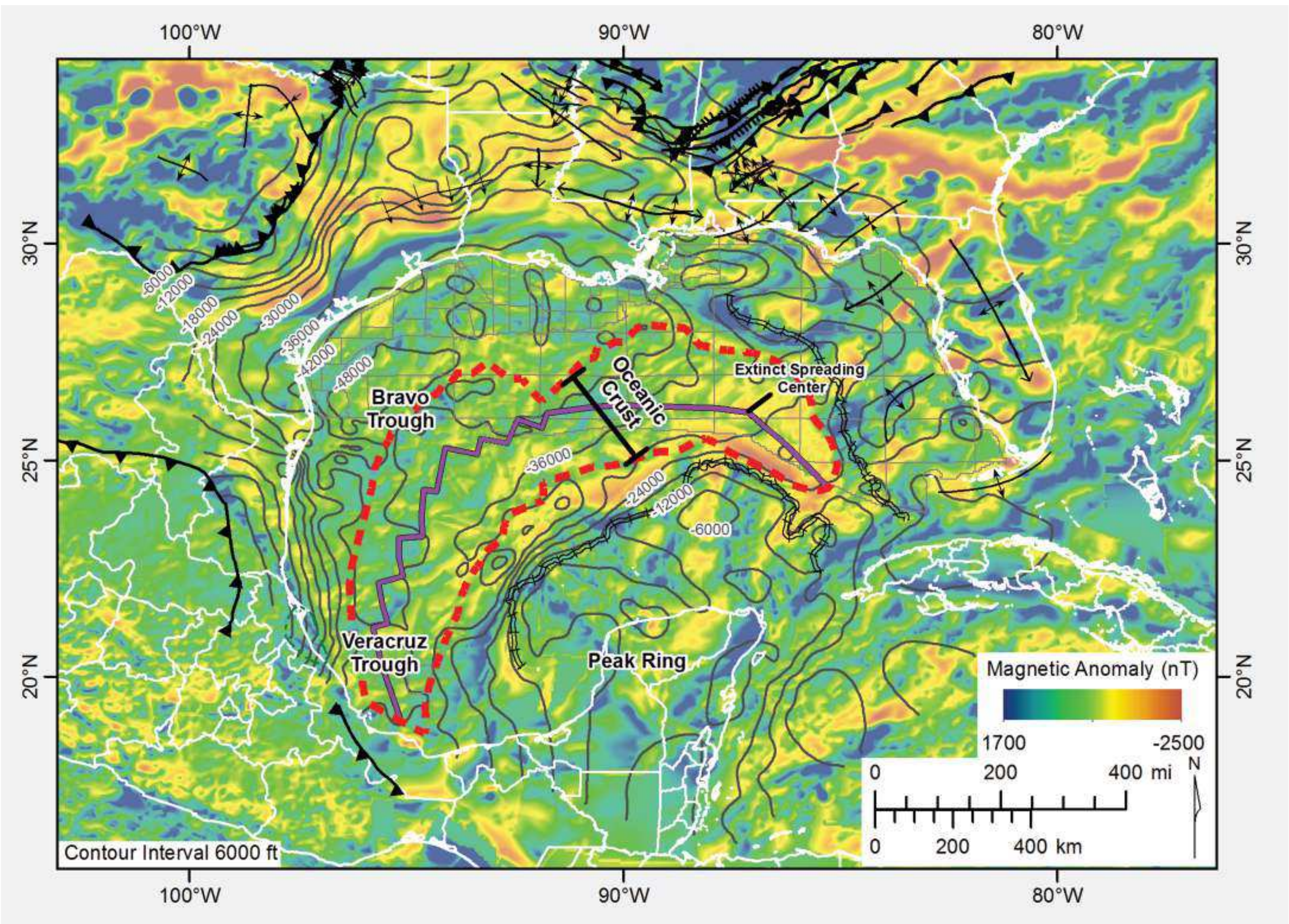

Figure 1.4 Mapped top of seismically defined basement with overlay of EMAG2 magnetic anomaly (Sandwell *et al.* 2014). Key tectonic features are discussed in the text. The limit of oceanic crust (red dashed line) is based on Hudec *et al.* (2013a, 2013b).

crystalline crust, and sedimentary crust, even where buried below thick intervals of salt and sedimentary rocks (Figure 1.5). In the northern GoM, a series of long (>500 km) seismic refraction lines were collected using bottom sensors (Figure 1.5). A line across the eastern GoM revealed the top of the mantle to shallow from about 34 km (21 miles) below the thick transitional crust below the Florida Platform to depths as shallow as 15 km (9 miles) in the area where oceanic crust is known to be present (Christeson *et al.* 2014; Figure 1.5). Above the mantle here lies a crystalline crust interval with unusually low velocities (in comparison to other areas), suggesting moderately attenuated continental crust. The sedimentary interval has compressional velocities in the range of 5.0 km/s (carbonate-dominated platform) to 3.0 km/s, where Miocene and younger strata are known to be present from well penetrations. The seismic refraction data also allow locating the boundaries of the LOC, here at a distance of 350–400 km from the start of the line just offshore of Florida. An intriguing observation is higher-than-expected seismic velocities at the LOC, suggestive of massive basalt emplacement associated with sea floor spreading (Christeson *et al.* 2014).

In the western GoM, seismic refraction data (Gumbo Line 1) revealed an unusual interval between high compressional velocity mantle and penetrated sedimentary crust (Van Avendonk *et al.* 2013). Below base of salt lies an unknown interval with considerable lateral crustal heterogeneity, thought to be rifted (attenuated) sedimentary crust with igneous intrusions. This interval ranges from 10–12 km at the top to as deep as 28 km depth above mantle rock. The lateral velocities variations that suggest igneous intrusions are documented in the shallow **pre-salt** interval of onshore areas, to be discussed in Section 2.2. The LOC is located inboard of the present-day Sigsbee Escarpment, though there is some uncertainty, given the thick salt canopy here (Van Avendonk *et al.* 2013). The presence of a pre-salt (Late Triassic[?] to Middle Jurassic[?]) interval in the deep northern GoM is consistent with observations from seismic reflection data in a pre-salt province

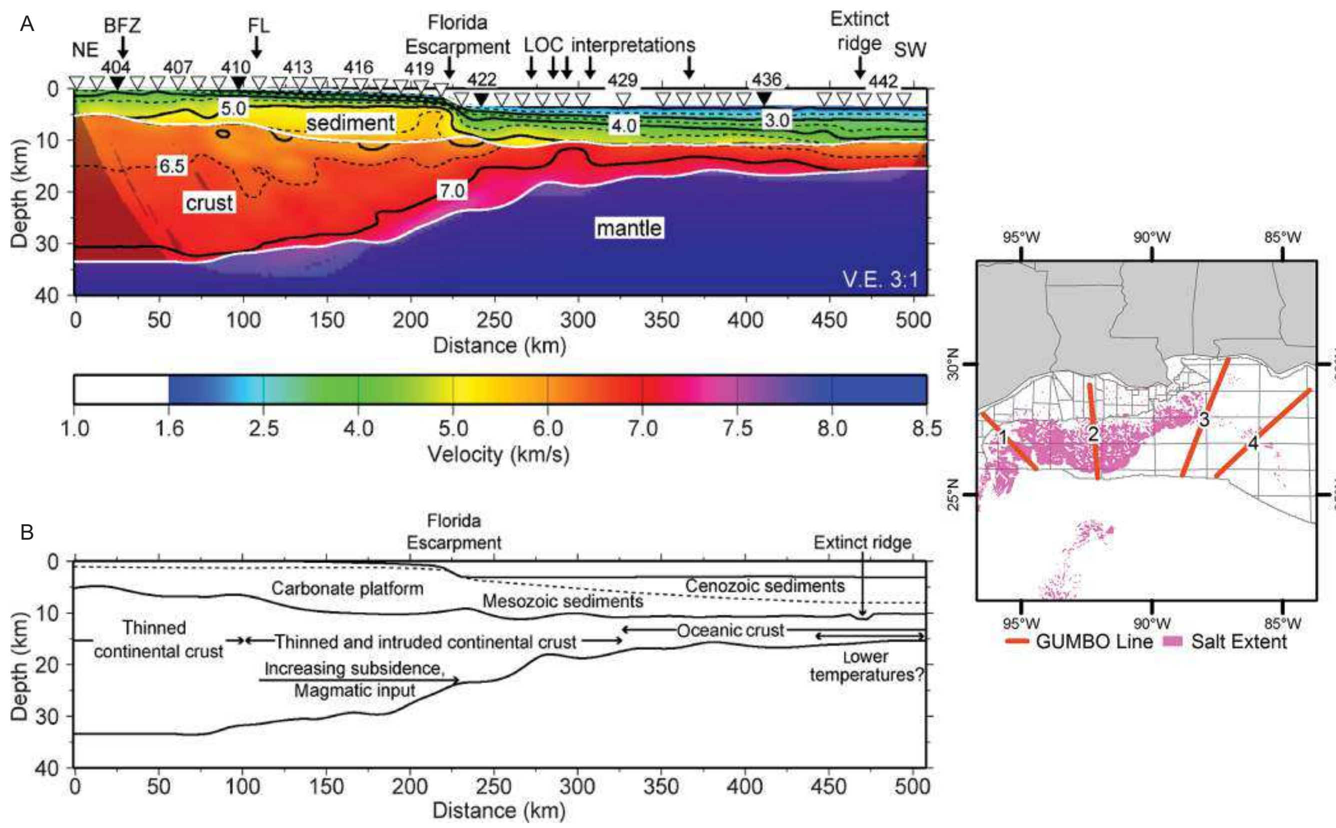

Figure 1.5 Seismic refraction data and interpretation, Gumbo Line 4, eastern GoM. Modified from Christeson *et al.* (2014).

offshore of Yucatán Province (Williams-Rojas *et al.* 2012; Miranda Peralta *et al.* 2014; Saunders *et al.* 2016).

1.2.3 Seismic Reflection Studies of Deep Crust

Seismic reflection surveys shot for oil and gas exploration provide some corroboration of seismic refraction interpretations, particularly for the eastern GoM where the salt canopy is absent. Here the general position of a Jurassic–Early Cretaceous **spreading center** in the eastern GoM has been suggested for many years, yet the precise location was not precisely known until Snedden *et al.* (2014) used several seismic criteria to define its location (Figure 1.6). Lin *et al.* (2019) subsequently refined its structure and evolution using newer vintage seismic reflection and gravity data. The extinct spreading center here displays morphological characteristics associated with slow-spreading mid-ocean ridges (rates of 1–4 cm/year; Perfit and Chadwick 1998): (1) large and wide axial valleys, 5–20 km wide; (2) deep axial valleys, often over 2 km deep; (3) normal faults that dip toward axial valleys; and (4) discontinuous, isolated basement highs, with elevations over 1 km above regional oceanic basement depth. Using seismic refraction data, Christeson *et al.* (2014) calculated a full spreading rate of 2.2 cm/year on a profile (Figure 1.5) in the same area. This estimate falls squarely in the slow spreading rate range globally and specifically for the comparable Mid-Atlantic Ridge system (McDonald 1982). Slow-spreading ridges express wide variety in tectonic and volcanic character, reflecting relatively unfocused magmatism (Sempere *et al.* 1993).

Structural-balanced restorations of the eastern Gulf further confirm the LOC location and timing of sea floor spreading (Curry *et al.* 2018). Upper Jurassic (Smackover and Norphlet) strata downlap onto oceanic crust, suggesting oceanic crust formation contemporaneous with deposition (Figure 1.6; see also Section 3.3.4). Latest Upper Jurassic (Haynesville-equivalent) and Cotton Valley intervals extend across all oceanic crust, constraining the end of sea floor spreading at about 155 Ma. These units are also contemporaneous with post-Smackover rafting in the eastern Gulf, suggesting a genetic relationship, as will be explored in Section 3.3.4.

1.2.4 Magnetic Data

Early attempts at mapping the extinct spreading center and LOC (Figure 1.4) were challenged by the generally indistinct character on magnetic data collected from the northern Gulf (e.g., Imbert and Phillippe 2005). This can be partly attributed to the low paleolatitude of the Gulf during the Jurassic, resulting in shallow magnetization vectors that subdued magnetic intensity at the surface but also the poor resolution of older surveys. Newer aeromagnetic data acquired for hydrocarbon exploration in Mexico have better constrained the location of oceanic crust, particularly when integrated with

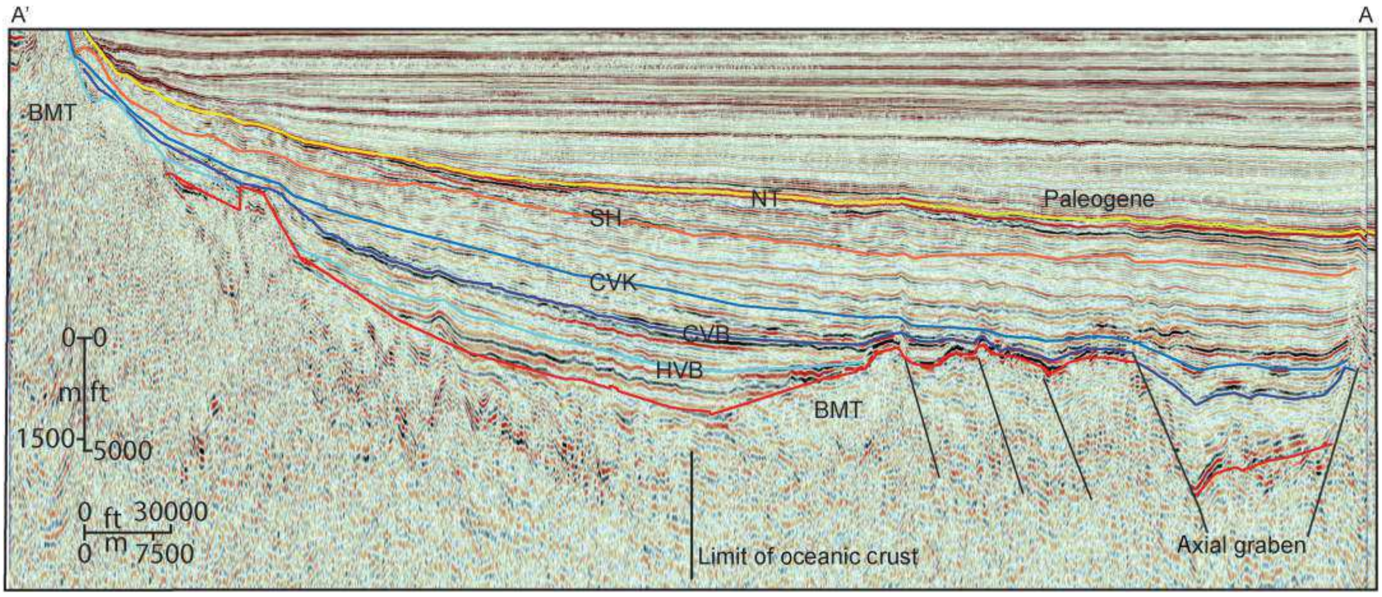

Figure 1.6 Seismic line interpretation in eastern GoM, extending from the Florida Platform across the inferred axial graben of the extinct spreading center showing lapout of HVB, CVB, and CVK supersequences onto oceanic crust. Other correlated horizons are SH, NT, and Paleogene (Wilcox) supersequences. Modified from Snedden *et al.* (2014). Seismic line courtesy of Spectrum. Abbreviations HVB, Haynesville–Buckner; CVB, Cotton Valley–Bossier; CVK, Cotton Valley–Knowles; SH, Sligo–Hosston; NT, Navarro–Taylor; BMT, basement.

comparable vintage northern Gulf data (Pindell *et al.* 2016). One prominent magnetic anomaly located in the central GoM has a distinctive pattern of orthogonally cross-cutting linear features superimposed upon an elongate margin parallel magnetic anomaly, thought to indicate the location of the youngest oceanic crust and thus the position of the extinct spreading center (Pindell *et al.* 2016). The calculated full spreading rates of 1–3.6 cm/year for the entire GoM are comparable to the slow spreading rates (2.2 cm/year) estimated for the eastern GoM (Christeson *et al.* 2014). Another trend, called the Campeche magnetic anomaly, is located downslope of the Yucatán **Platform margin** and constrains the Yucatán (Mayan) block position at the start of pre-salt deposition here, as discussed in Chapter 3.

1.2.5 Gravity Data

Sandwell gravity maps (Sandwell *et al.* 2014) also provide further documentation of the present-day crustal types and their position. Continental crust is generally indicated by gravity highs (e.g., Yucatán block) and oceanic crust by gravity lows, but local variations can occur as a function of igneous intrusions, salt, and depth variations along prominent escarpments.

1.3 Gravity Tectonics

Above the crystalline basement in the greater GoM basin, a thick sedimentary interval exists, deposited largely in the Mesozoic and Cenozoic. Beginning in the Jurassic, robust depositional systems delivered sediment into the basin, the siliciclastic systems fed by rivers draining a variety of source terranes in the northern Rockies, southern Rockies, Appalachians, Quachita Mountains (USA), and Sierra Madres and other areas of Mexico. Siliciclastic systems are particularly prominent in the Cenozoic, but Mesozoic systems of the Jurassic and Cretaceous were, at times, equally impressive in terms of accumulated thickness and caliber of sediment grade. Cenozoic deposition, which extended past the rigid Mesozoic carbonate margins, induced significant basinward translation due to gravitational loading. Shelf margin sediment loading and faulting created accommodation space and, where the Louann Salt was encountered, major salt evacuation. The resulting sedimentary accumulations were unusually thick (often >25,000 ft) but barely kept pace in the northern GoM with sediment influx from numerous continental-scale rivers. Loading onto salt also created complex salt mobilization and salt–sediment interaction that set up a wide diversity of trap types, heat flow variations, pathways for hydrocarbon migration, **depositional architectures**, and seal rock distributions.

As will be discussed in Section 9.4, improvements in imaging and illumination of the **subsalt** structure has vastly enhanced our understanding of the early basin history in the slope and **abyssal plain**. Regional to basinal scale seismic analysis has led to recognition of both extensional and contractional tectonics (and even raft tectonics) throughout the Mesozoic and Cenozoic. The extensive seismic and well control means that the structures here are well-imaged and thus studied (Worrall and Snelson 1989; Nelson 1991; Jackson *et al.* 1994; Diegel *et al.* 1995; Peel *et al.* 1995; Watkins *et al.* 1996a; Rowan *et al.* 2000, 2016; Radovich *et al.* 2007).

It is therefore worthwhile to describe some of the important structural styles that have been identified to date. It is also useful to view these tectonic features in the context of structural domains (Section 1.4) and 10 basin-scale cross-sections (Section 1.5).

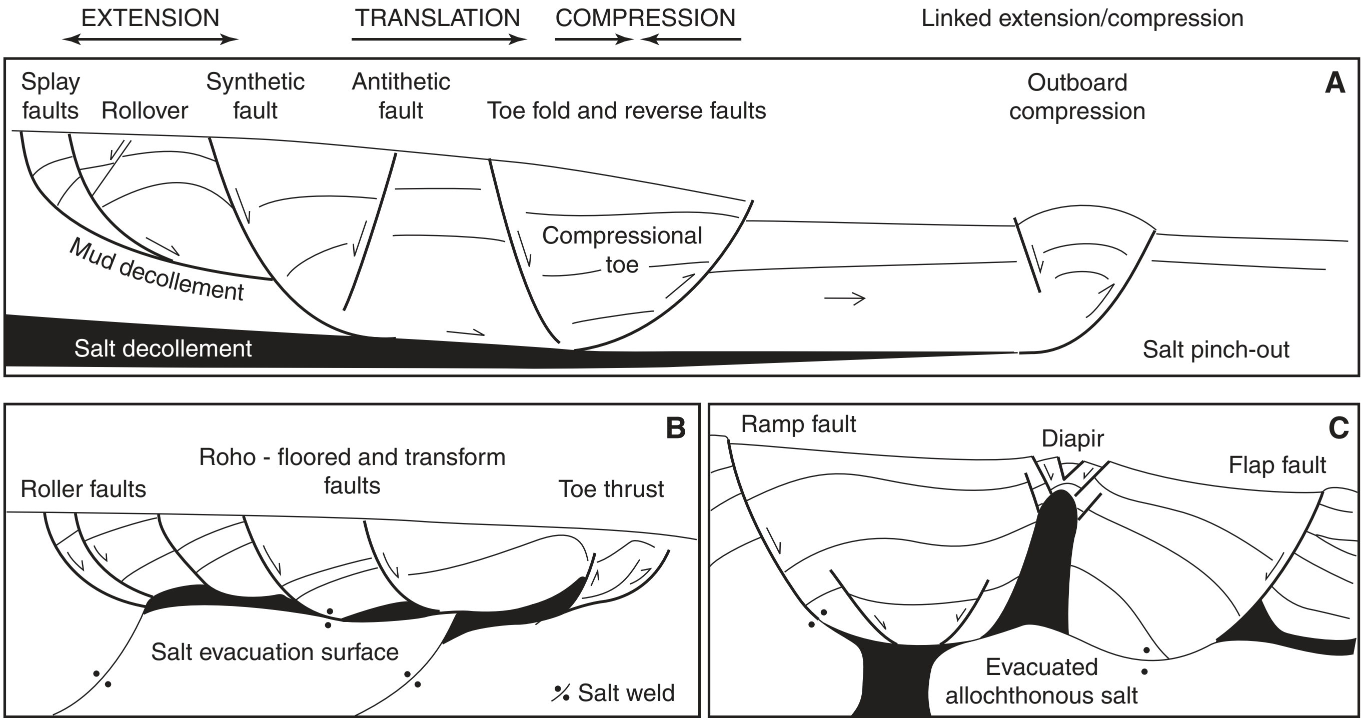

Figure 1.7 GoM gravity tectonics. **(A)** Linked extension and compression. **(B)** Roho salt detachment. **(C)** Salt withdrawal minibasin. From Galloway (2008).

Several pre-conditions set up the complex and diverse assemblages of GoM basin gravity tectonic structures. The combination of a thick, basin-floor Louann Salt substrate, rapid sediment loading, and offlap of a high-relief, continental margin sediment prism has resulted in mass transfer of salt and overpressured mud upward and basinward throughout Gulf history.

1.3.1 Growth Fault Families and Related Structures

Growth faults tend to nucleate and grow during active deposition at the **continental margin** (Winker 1982; Watkins *et al.* 1996b; Jackson and Hudec 2017). Here, extension results from basinward gravitational gliding or translation of the sediment wedge along a **detachment** zone, typically found within salt or overpressured deep marine mud (Rowan *et al.* 2004). Extension creates a family of features, including primary synthetic growth faults, splay faults, antithetic faults, and rollover anticlines (Figure 1.7A). In many parts of the GoM, updip extension is more or less balanced by a similar degree of contraction in downdip areas, as discussed in the following sections.

1.3.2 Basin-Floor Contractional Fold Belts

Basinward gravity spreading or gliding along a detachment zone, and resultant updip extension, requires compensatory compression at the toe of the displaced sediment body (Weimer and Buffler 1992; Hall *et al.* 1993; Fiduk *et al.* 1999; Trudgill *et al.* 1999). Contractional features include anticlinal toe folds and reverse faults (Figure 1.7A). They commonly form at the base of the slope, but can also extend onto the basin plain where a stepped discontinuity or termination of the decollement layer occurs. The deepwater fold belts (Atwater, Mississippi, etc.) are thought to represent adjustments to significant updip extension (Radovich *et al.* 2007). In other areas, extension may be balanced by squeezing salt bodies or salt weld development (Jackson and Hudec 2017; see Section 1.3.6).

1.3.3 Allochthonous Salt Bodies, Including Salt Canopies and Salt Sheets

Loading of the Louann Salt has resulted in regional extrusion of salt basinward and upward (Diegel *et al.* 1995; Fletcher *et al.* 1995; Peel *et al.* 1995). **Allochthonous salt** canopies typically develop beneath the continental slope, where salt rises as a series of coalescing diapirs or as injected tongues. Salt may also be extruded to the surface, forming salt sheets, or nappes, which move basinward, much like salt glaciers (Jackson and Hudec 2017).

1.3.4 Roho Fault Families

Lateral salt extension by gravity spreading creates a linked assemblage of extensional faults and compensating, downslope compressional toe faults, anticlines, and salt injections in the overlying sedimentary cover (Rowan 1995; Schuster 1995). In some cases, the top of allochthonous salt can acts as a decollement surface for faults (Figure 1.7B), as does autochthonous salt previously described. These are called **roho systems** and often occur in stratigraphically distinct fault groups or **fault families**.

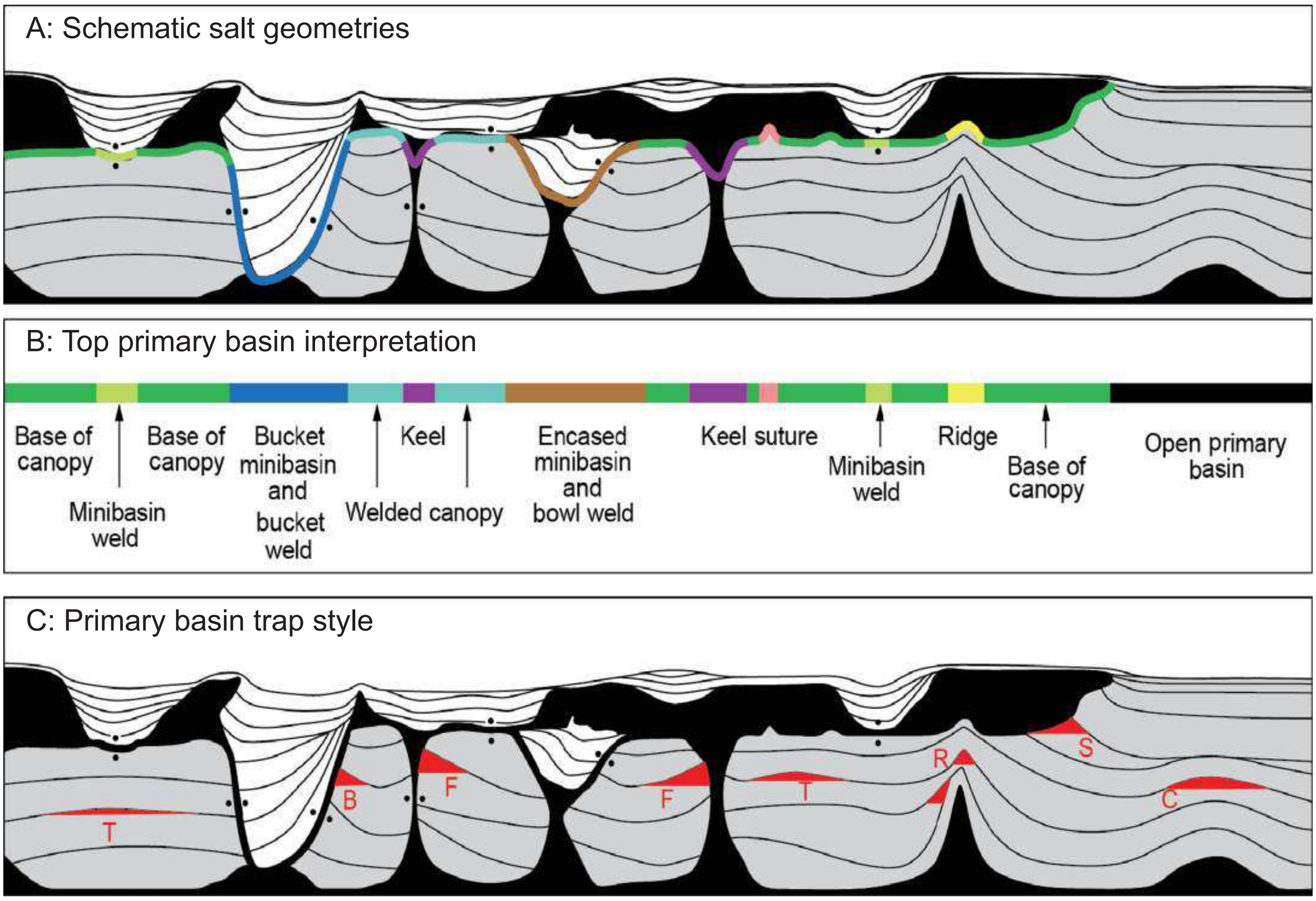

Figure 1.8 Schematic cross-sections of the bucket weld province. (A) Schematic salt geometries based on seismic interpretation. (B) Primary top basin interpretation. (C) Primary basin trap style. Letters indicate different trap styles in subsalt domain. Modified from Pilcher *et al.* (2014).

1.3.5 Salt Diapirs and Their Related Withdrawal Synclines and Minibasins

In the Gulf margin basins and embayments, salt **diapirs** rise directly from the autochthonous Louann "mother" salt (Seni and Jackson 1983; Fletcher *et al.* 1995; Rowan 1995; Rowan and Weimer 1998). Basinward, depositional loading of salt canopies and sheets beneath shelf and slope areas also causes renewed salt stock evacuation, creating high-relief salt diapirs and intervening depressions (Figure 1.7C). Progressive **salt evacuation** creates shifting, localized sites of extreme subsidence and sediment accumulation. Resulting features include withdrawal synclines created by local evacuation of salt from diapir flanks, bathymetric depressions, called minibasins, that form local depocenters, turtle structures, and local fault families, including down-to-basin ramp faults, counter-regional flap faults, and crestal faults above salt bodies.

1.3.6 Salt Welds

Salt welds are surfaces or zones that join strata originally separated by either autochthonous or allochthonous salt (Hudec and Jackson 2011). These are present where nearly complete expulsion of salt from stock feeders, dikes, salt tongues, or salt canopies has occurred (Jackson and Cramez 1989; Jackson *et al.* 1994; Figure 1.7B,C). Because the welds form some time after the deposition of adjacent strata, these juxtapose discordant stratigraphic intervals, sometimes with significant angularity of converging reflections (Hudec and Jackson 2011). Primary, secondary, and tertiary welds can be identified on the basis of the type of salt body that was welded (Jackson and Hudec 2017). Welds can also serve as detachment surfaces for younger listric faults.

Younger (secondary) sedimentary minibasins may be welded against older (primary) minibasins, resulting in drastically different ages, lithologies, and subsurface pressures (Pilcher *et al.* 2011; Figure 1.8). These are particularly prominent in a portion of the central GoM, the so-called "**bucket weld**" province. These bucket welds can act as lateral boundaries to hydrocarbon traps.

Salt welds can also act as regional decollement surfaces even when obvious linkage to downdip contraction is lacking. Regional decollements at welds are also known to be significant horizontal pressure barriers, with a significant increase in pressure in sub-weld intervals and attendant increase in risk, uncertainty, and well costs (see Section 9.20.1 on the Wilcox deep shelf play). Reverse faults can occur along welds (thrust welds), as observed in Campeche.

1.3.7 Rollovers and Expulsion Rollovers

Thickening and bending of strata toward a listric normal fault is commonly observed in the GoM Cenozoic and Mesozoic intervals. If expulsion of salt occurs to cause stratal thickening and rotation, with or without a fault, this structure is referred to as an **expulsion rollover** (Ge *et al.* 1997; Jackson and Hudec 2017). Large expulsion rollover structures have been identified in the Mississippi Canyon protraction block and represent some of the largest undrilled prospects in the basin (Harding *et al.* 2016). The orientation of these expulsion rollovers may indicate the general direction of sediment transport and loading (McDonnell *et al.* 2008), though these features are several orders of magnitude larger than depositional clinoforms and should not be used to indicate the location of paleo-shelf margins.

1.3.8 Carapaces and Rafts

When moving salt carries roof material that is not firmly attached to surrounding strata, stratigraphic discontinuities can occur. Transported roof material can be tens of kilometers in lateral extent and sometimes as thick as the salt body (Jackson and Hudec 2017). The term carapace is used here in a restrictive sense to describe detached blocks above salt that have moved vertically relative to the surrounding strata, either actively by diapir rise or passively as younger sediments are deposited around the salt-supported blocks (Figure 1.9). Early drilling at or around the allochthonous salt canopy encountered blocks which tended to be older, thinner, and/or more stratigraphically condensed than the adjacent non-carapace interval (Hart *et al.* 2004). Carapaces are often structurally much higher than the regional level of coeval strata. For example, the Norton well (GB 754 #1) penetrated a carapace block where Top Cretaceous was encountered at 7180 ft (2189 m), much shallower than the regional depths of Cretaceous, closer to 30,000 ft (9.1 km; Cunningham *et al.* 2016). Initially, stratigraphic discontinuities within carapaces caused considerable confusion, including the misinterpreted Middle Cretaceous unconformity (MCU), which later analyses proved was actually the Cretaceous–Paleogene boundary (K–Pg; Dohmen 2002).

Carapaces do accumulate sediment above a diapir documenting that diapir's history, but also record information on older strata that is relevant to regional or basin reconstructions. Carapaces containing organically enriched intervals within both the Tithonian and Ceno-Turonian intervals provide critical evidence in characterization of these source rocks (Cunningham *et al.* 2016).

Rafts are more complicated salt tectonic features that are defined in two different ways. First, we recognize rafts as stratigraphic blocks formed as part of raft tectonic processes. Raft tectonics is a form of thin-skinned extension, with unusually large degrees of extension such that the footwall and hanging wall are often not in contact, unlike growth faults (Jackson and Hudec 2017). Raft gaps are filled in by synkinematic (syn-extensional) strata. Raft tectonics is well-documented in the Albian interval of Angola and the Oxfordian interval of the DeSoto Canyon protraction block (Pilcher *et al.* 2014).

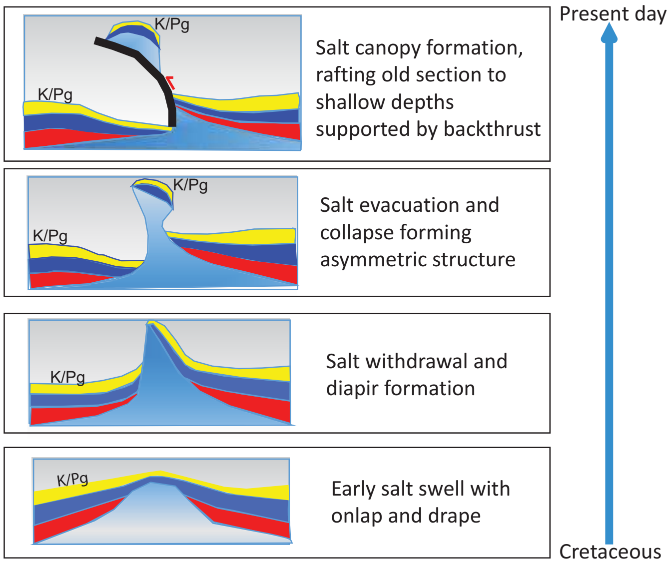

Figure 1.9 Development of a salt-related carapace structure. Modified from M. Rowan (pers. comm.).

A second use of the term raft applies to stratigraphic blocks that have been moved considerable distances downslope by allochthonous salt. For example, it is established from 3D seismic analysis that the salt canopy in the deepwater northern GoM has transported over 20 raft blocks across the Alaminos Canyon, Keathley Canyon, Walker Ridge, and Green Canyon protraction blocks, with distances ranging from less than 3 km to more than 80 km from their original positions (Fiduk *et al.* 2014). Over 3100 km^2 of rafted strata was identified, largely accumulating near the terminus of the salt canopy.

Primary or secondary minibasins (terminology of Pilcher *et al.* 2011) can become encased in salt as allochthonous salt flows over the minibasin subsiding onto a deeper salt level (Hudec and Jackson 2011). In some cases, salt evacuation continues, and the minibasin is instead surrounded by welds (Rowan and Inman 2011).

Thus, it is very important to consider the tectonic history of vertical and lateral salt transport when analyzing stratigraphic information from carapaces, rafts, and encased minibasins. Stratigraphic discontinuities are common, and in areas of poor seismic imaging are only revealed by drilling and biostratigraphic analysis. Some wells have penetrated salt-overturned intervals, where biostratigraphic datums are encountered in reverse order, resulting in major drilling "surprises" (Box 1.1).