Jurassic evolution of the Gulf of Mexico salt basin

Michael R. Hudec, Ian O. Norton, Martin P. A. Jackson, and Frank J. Peel

ABSTRACT

We present a new hypothesis for the Jurassic plate-tectonic evolution of the Gulf of Mexico basin and discuss how this evolution influenced Jurassic salt tectonics. Four interpretations, some based on new data, constrain the hypothesis. First, the limit of normal oceanic crust coincides with a landward-dipping basement ramp near the seaward end of the salt basin, which has been mapped on seismic data. Second, the deep salt in the deep-water Gulf of Mexico can be separated into provinces on the basis of position with respect to this ramp. Third, paleodepths in the postsalt sequence indicate that salt filled the Gulf of Mexico salt basin to near sea level. Fourth, seismic data show that postsalt sediments in the central Louann and the Yucatan salt basins exhibit large magnitudes of Late Jurassic salt-detached extension not balanced by equivalent salt-detached shortening.

In our hypothesis, Callovian salt was deposited in preexisting crustal depressions on hyperextended continental and transitional crust. After salt deposition ended, rifting continued for another 7 to 12 m.y. before sea-floor spreading began. During this phase of postsalt crustal stretching, the salt and its overburden were extended by 100 to 250 km (62–155 mi), depending on location. Sea-floor spreading divided the northern Gulf of Mexico into two segments, separated by the northwest-trending Brazos transform. The eastern segment opened from east to west, leaving the Walker Ridge salient in the center of the basin as the final area to break apart. In some areas, salt flowed seaward onto new oceanic crust, first concordantly over the basement as a parautochthonous province, then climbing up over stratigraphically younger strata as an allochthonous province.

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INTRODUCTION

Overview

The Gulf of Mexico Basin began forming during the Triassic–Early Cretaceous rifting between the Yucatan microplate and the North American plate (Pindell and Dewey, 1982; Salvador, 1991a, b). However, despite more than 60 yr of study, thousands of boreholes, and one of the most intensive seismic acquisition and processing efforts in the world, many fundamental aspects of the geologic evolution of the basin remain speculative. This uncertainty is especially true of the Jurassic, during which Louann Salt was deposited and most crustal extension occurred (Figure 1).

The uncertainty in the Jurassic evolution of the Gulf of Mexico basin exists mostly because, until recently, little reliable seismic imaging of the Mesozoic section over much of the basin has existed. In the northern Gulf of Mexico, much of the Jurassic structure is obscured beneath sediments 9 to 18 km (6–11 mi) thick (e.g., Sawyer et al., 1991; Peel et al., 1995), cut by an intricate network of allochthonous salt and related welds (defined by Hudec and Jackson, 2011, as surfaces or zones joining strata originally separated by autochthonous or allochthonous salt). Most of the southern Gulf of Mexico has been surveyed only by widely spaced two-dimensional seismic data, most of which have not been available outside Petróleos Mexicanos (PEMEX) or the Instituto Mexicano del Petróleo.

In the last few years, however, availability of high-quality data has increased. In the deep-water northern Gulf of Mexico, wide-azimuth seismic acquisition and prestack depth migration have permitted widespread imaging of the Jurassic section. In the southern Gulf of Mexico, PEMEX and Instituto Mexicano del Petróleo have acquired large swaths of three-dimensional data and are publishing more interpretations and participating in more collaborative studies (e.g., Hernández-Mendoza et al., 2008; Bartolini and Román Ramos, 2009). As a result, a seismically constrained interpretation of Jurassic basin history is now possible.

Our goals in this article are to synthesize information on Jurassic plate movements, salt deposition, and salt tectonics in the Gulf of Mexico Basin. Understanding how plate tectonics controlled salt deposition and flow is critical to an understanding of basin evolution. In turn, the Jurassic history of salt flow constrains interpretations of plate kinematics and tectonic history. The salt isopach and structures created during Jurassic rifting strongly influenced all subsequent salt tectonics in the Gulf of Mexico Basin. These influences are discussed in

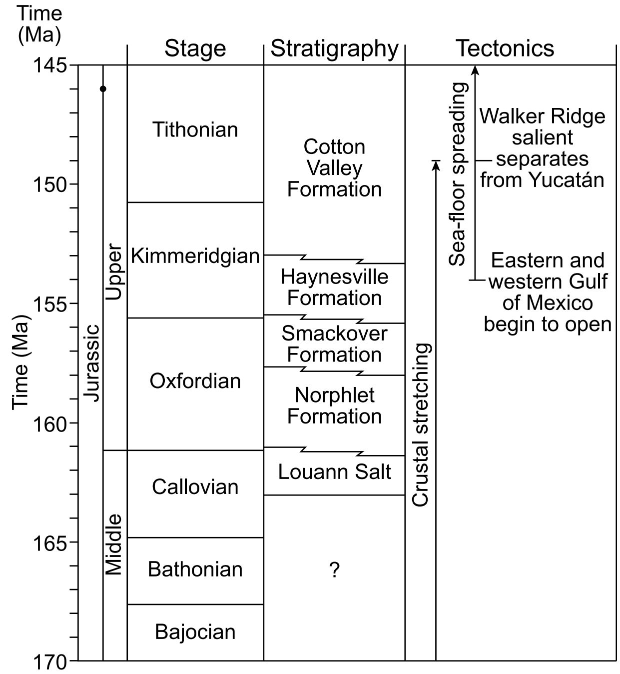


Figure 1. Timing chart and stratigraphic column for Middle–Late Jurassic evolution of the Gulf of Mexico Basin.

more detail in our companion articles (Dooley et al., 2013, this issue; Hudec et al., 2013, this issue).

Questions addressed in our study include the following:

- Where is the limit of normal oceanic crust (LOC) in the Gulf of Mexico? What is its age?
- What was the temporal relationship between sea-floor spreading and salt deposition? Was the Louann Salt deposited entirely on thinned continental and magmatic transitional crust or partly on the normal oceanic crust?
- How far did salt flow seaward after it was deposited? How did this distance vary within the basin?

Salt Basins in and around the Gulf of Mexico

A major problem in the Gulf of Mexico geology is the abundance of published names for geologic features (Salvador, 1991c). Many structures have several names, for example, the Apalachicola embayment boasts at least 14 different names (Arthur, 1988). We have followed the most common names and have tried to minimize introducing new names.

We use “Gulf of Mexico salt basin” to refer collectively to all Callovian salt basins and subbasins in the Gulf of Mexico (Figure 2). As in many passive-margin salt basins, salt in the Gulf of Mexico currently lies in two pieces on the opposite sides of an ocean basin. We refer to the northern Gulf of Mexico salt basin as the “Louann salt basin.” The southern Gulf of Mexico salt basin has many different names. We prefer “Isthmian salt basin” (Cuenca Salina del Istmo) because it has a long history in the literature (Viniegra, 1971; Garrison and Martin, 1973) and is consistent with the official nomenclature for Mexican geologic provinces (Mexican National Hydrocarbons Commission, 2010).

The Louann salt basin consists of the central Louann salt basin and two sets of interior basins around its periphery (Figure 2). On the north, the Toledo Bend flexure (Anderson, 1979) separates the central Louann salt basin from the Mississippi, North Louisiana, and East Texas interior salt basins. The paucity of salt domes and growth faults above the Toledo Bend flexure suggests that salt is thin or absent there, which, in turn, indicates a structural high during salt deposition. Anderson (1979) inferred that the flexure localized the Louisiana shelf edge throughout most of the Cretaceous. To the west, the Sabinas and La Popa interior salt basins and the Monterrey salt basin (part of the Monterrey trough) are separated from the central Louann salt basin by the Burro-Salado and Tamaulipas arches (e.g., Goldhammer and Johnson, 2001; Lawton et al., 2001). The northwestern end of the Sabinas salt basin connects to the border rift system of Dickinson and Lawton (2001), but most of this rift does not contain significant Callovian salt.

On the basis of unpublished data, we suggest dividing the Isthmian salt basin into two subbasins, with thinner salt separating them across a basement ridge extending to the northwest from the northwestern corner of the Yucatan platform (Figure 2). The Campeche salt basin is southwest of this ridge, in the Bay of Campeche and adjacent onshore areas. The Yucatan salt basin is northeast of the ridge, in the offshore Yucatan state.

Most early workers in the Gulf of Mexico interpreted salt across the entire basin (Lyons, 1957;

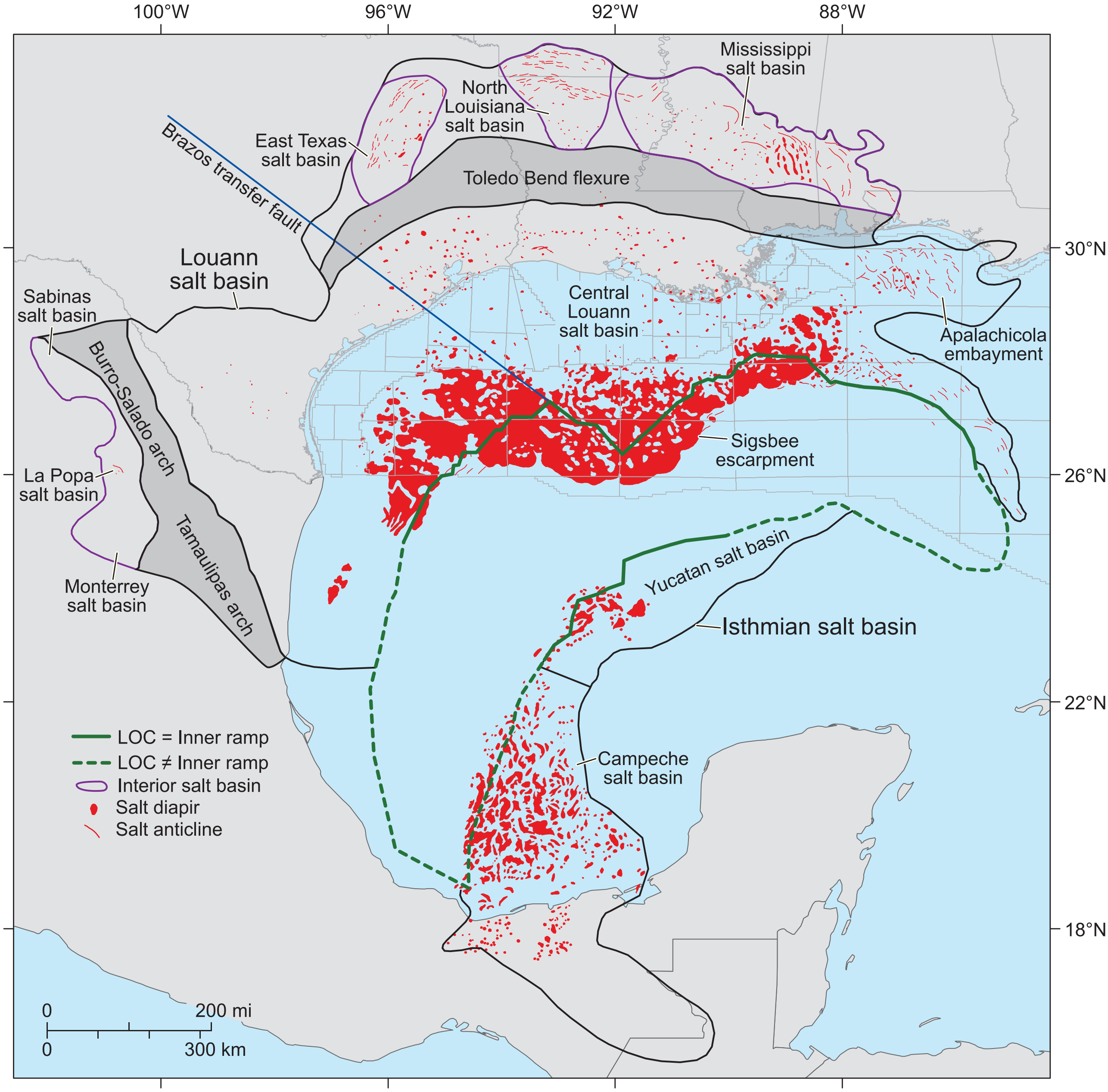


Figure 2. Salt basins in the Gulf of Mexico region, showing locations of salt structures and our interpreted limit of normal oceanic crust in the basin. Structural features from Martin (1980); Simmons (1992); Diegel et al. (1995); Lopez (1995); Goldhammer and Johnson (2001); Lawton et al. (2001); Padilla y Sánchez (2013). See Hudec et al. (2013, this issue), for evidence for the Brazos transfer fault.

Andrews, 1960; Ewing et al., 1962; Wilhelm and Ewing, 1972; Watkins et al., 1978). However, mapping of offshore diapir provinces on shallow-penetration seismic data revealed that no salt diapirs were in the center of the basin, leading to the hypothesis that salt was deposited only in geosynclines around the basin margin (Meyerhoff, 1967; Antoine and Bryant, 1969).

The advent of plate-tectonic theory soon led to the speculation that the diapir provinces on the northern and southern sides of the Gulf of Mexico were once a continuous salt basin that split apart during rifting. The idea of a bisected Gulf of Mexico salt basin goes back at least as far as articles by Kupfer (1974) and Burke (1975) and has been a feature of most regional syntheses and plate

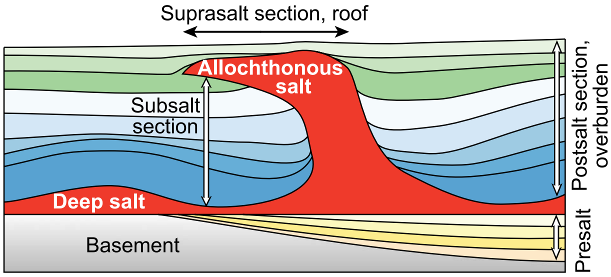


Figure 3. Terminology used to describe location with respect to salt in the Gulf of Mexico. Key terms include overburden (n), sediments lying above the deep-salt layer; postsalt (adj), lying above the deep-salt layer; roof (n), sediments lying above allochthonous salt; presalt (adj), lying beneath the deep-salt layer; suprasalt (adj), lying above allochthonous salt; subsalt (adj), lying below allochthonous salt but above the deep-salt layer.

restorations since then (Humphris, 1978; Buffler et al., 1980, 1981; Hall et al., 1982; Pindell and Dewey, 1982; Buffler and Sawyer, 1985; Pindell, 1985; Salvador, 1987, 1991a; Winker and Buffler, 1988; Buffler, 1989; Sawyer et al., 1991; Buffler and Thomas, 1994; Marton and Buffler, 1994; Watkins et al., 1995; Pindell and Kennan, 2001, 2007, 2009; Bird et al., 2005; Imbert and Philippe, 2005).

We explore how the Gulf of Mexico Basin opened by focusing on four key interpretations, some of which are based on new data: (1) the LOC coincides with a landward-dipping ramp near the seaward end of the salt basin; (2) the deep salt in the deep-water Gulf of Mexico salt basin can be separated into provinces, depending on position with respect to this ramp; (3) salt filled the Gulf of Mexico salt basin to near sea level; and (4) postsalt sediments in the central Louann and Yucatan salt basins exhibit large magnitudes of Late Jurassic salt-detached extension that is not balanced by equivalent salt-detached shortening (see Figure 3 for an explanation of terminology used to describe location with respect to the salt in the Gulf of Mexico). We discuss these interpretations then propose a chronology for the Jurassic history of the basin.

KEY OBSERVATIONS AND INTERPRETATIONS

Limit of Normal Oceanic Crust in the Deep-Water Gulf of Mexico

More than 25 yr ago, Lin (1984) was the first to recognize a major landward-dipping basement ramp in the deep-water Gulf of Mexico at the seaward end of the Yucatan salt basin. As seismic data quality improved, this ramp ("inner ramp" in our terminology; Figure 4) has been recognized over large areas near the seaward margin of salt in both the northern and southern Gulf of Mexico (Rowan et al., 2000; Pindell, 2002; Pindell et al., 2003; Imbert, 2005; Imbert and Philippe, 2005; Pindell and Kennan, 2007; Winker, 2007; Barker and Mukherjee, 2011).

In most of the deep-water Gulf of Mexico, because the Louann Salt rests directly on the acoustic basement, the inner ramp is imaged as a 1- to 4-km (0.62–2-mi)-high step in the base of the salt (Figures 5, 6). We contend that the inner ramp marks the LOC in the Gulf of Mexico, given three lines of evidence.

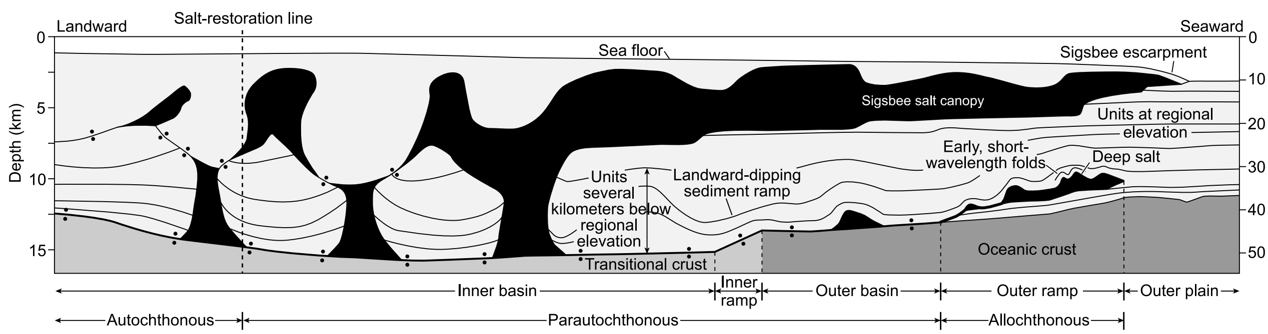
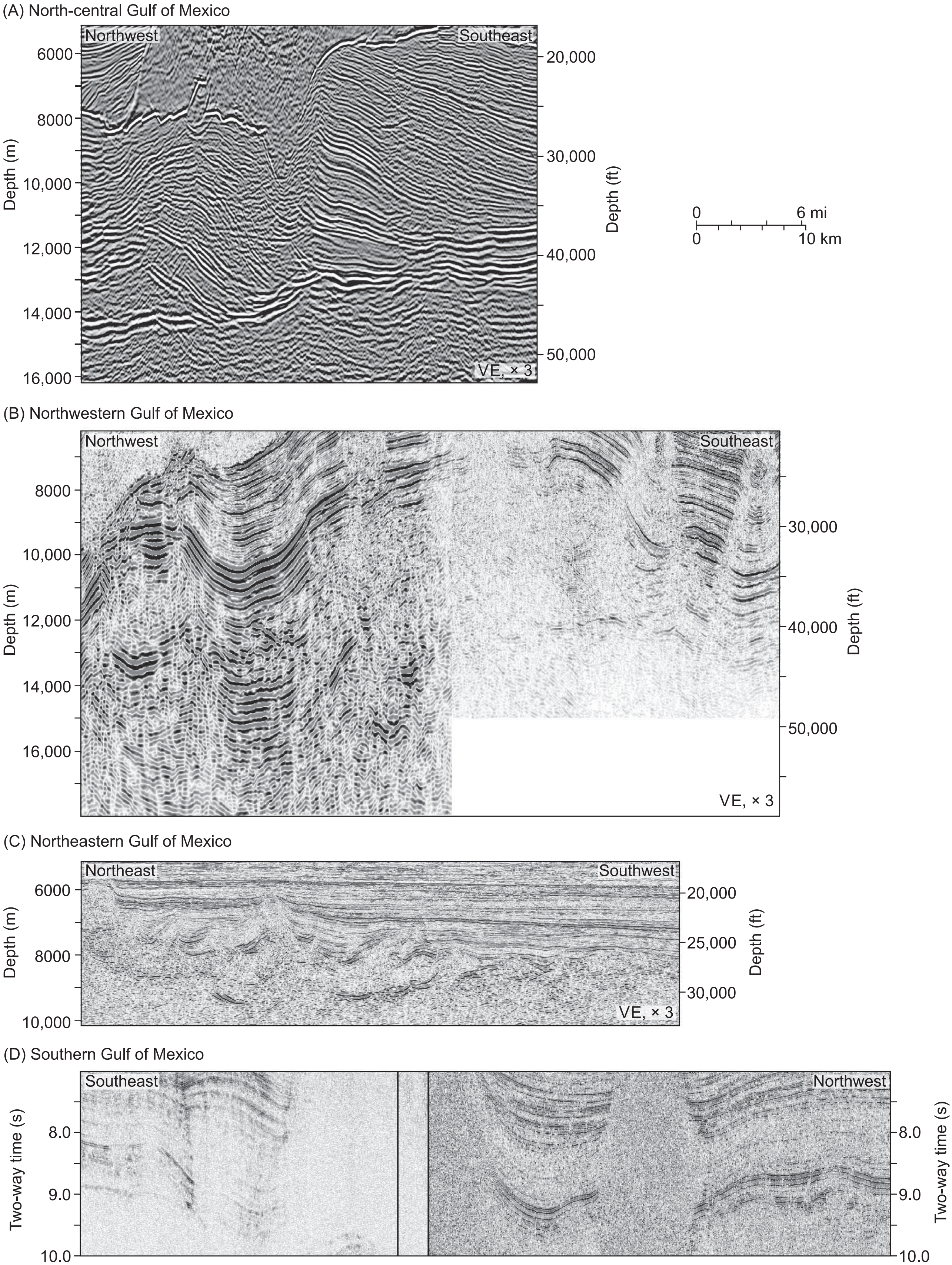


Figure 4. Schematic cross section showing tectonic provinces in the deep-water central Louann Salt basin. Horizontal scale is variable. Vertical scale is appropriate for areas just east of the Brazos transfer fault (basement in the inner basin would be deeper to the west of the fault).


(A) North-central Gulf of Mexico
Northwest
Southeast
Depth (m)
6000
8000
10,000
12,000
14,000
16,000
Depth (ft)
20,000
30,000
40,000
50,000
VE, × 3
0
6 mi
0
10 km
(B) Northwestern Gulf of Mexico
Northwest
Southeast
Depth (m)
8000
10,000
12,000
14,000
16,000
Depth (ft)
30,000
40,000
50,000
VE, × 3
(C) Northeastern Gulf of Mexico
Northeast
Southwest
Depth (m)
6000
8000
10,000
Depth (ft)
20,000
25,000
30,000
VE, × 3
(D) Southern Gulf of Mexico
Southeast
Northwest
Two-way time (s)
8.0
9.0
10.0
Two-way time (s)
8.0
9.0
10.0

First, the inner ramp coincides with LOCs defined on potential-field data in offshore Florida and Yucatan (Figure 7). The inner ramp is easily mapped here because of thinner cover and the absence of shallow allochthonous salt (Figure 6C, D). In both areas, the crest of the inner ramp coincides roughly with prominent lineaments on magnetic and isostatic gravity maps, which we interpret to mark the LOC. The relationship between the LOC and the inner ramp here should apply across the central Louann salt basin, where 9- to 13-km (6–8-mi)-thick sediments obscure the magnetic and gravity signals that are typically used to locate the LOC.

Second, kilometer-scale basement ramps have been interpreted from seismic data at the seaward ends of the Santos (Gladczenko et al., 1997), Campos (Unternehr et al., 2010), Espírito Santo (Mohriak et al., 2008), and Kwanza (Marton et al., 2000; Hudec and Jackson, 2004; Unternehr et al., 2010) salt basins. In each basin, the basement ramp lies just inboard of the LOC, as defined by potential-field data. These analogs suggest that landward-dipping basement ramps marking the LOC are common at the seaward ends of passive-margin salt basins. Why transitional crust inboard of the ramps should systematically be deeper than adjoining the normal oceanic crust farther seaward is unclear. Kneller and Johnson (2011) suggested that much of the transitional crust beneath the northern Gulf of Mexico slope may have been produced by ultraslow lithosphere spreading. Imbert and Philippe (2005), Pindell and Kennan (2007), and Mickus et al. (2009) argued that much of the basement beneath the northern Gulf of Mexico slope consists of oceanic or protooceanic crust. Unternehr et al. (2010) suggested that the crust immediately inboard of ramps in South Atlantic salt basins is exhumed subcontinental mantle. Alternatively, Huismans and Beaumont (2011) inferred that the crust inboard of ramps is a thin layer of hyperextended continental crust overlying oceanic lithosphere. In any case, the greater depth of basement inboard of the ramps implies that the crust there is thinner or denser than the normal oceanic crust, or both.

Finally, all interpretations of the inner ramp as something other than the LOC have serious shortcomings. We have considered three alternatives. (1) The inner ramp is a fracture zone or other feature within the oceanic crust. However, the inner ramp is a relatively steep, 1- to 4-km (0.62–2-mi)-high rift-parallel ramp that extends more than 800 km (497 mi) alongstrike. We are unaware of similar structures formed entirely on the normal oceanic crust anywhere in the world. Any interpretation that places the LOC inboard of the inner ramp (e.g., Imbert and Philippe, 2005; Mickus et al., 2009) must therefore explain why a 1- to 4-km (0.62–2-mi)-high ramp would form on the normal oceanic crust and why it appears unique to the Gulf of Mexico. (2) The inner ramp is a result of flexural loading by overlying sediments. Flaws of this idea are the short width of the inner ramp (5–30 km [3–19 mi]) and the sharpness of its upper and lower hinges (Figure 6). Flexural bending, by contrast, typically operates on scales of hundreds of kilometers and has long-wavelength, gently curved hinge zones (e.g., Watts, 2001; Allen and Allen, 2005). (3) The inner ramp is a late deformational structure. This interpretation fails to explain why younger sediments are not folded above the inner ramp in some areas (Figure 6A, C), but instead, onlap it. Furthermore, it would be difficult to explain the origin of an 800-km (497-mi)-long, short-wavelength kink fold in the crust.

We emphasize that our interpretation of the LOC is based primarily on seismic interpretation, instead of the more traditional use of gravity and magnetic data. To date, gravity and magnetic data have been of limited use in identifying crustal boundaries in the north-central Gulf of Mexico,

Figure 5. Uninterpreted seismic sections across the inner ramp. Sections A to C were depth migrated and are shown at the same scale. Section D is in time but has the same horizontal scale as that of the first three sections. (A) Central Louann salt basin, north-central Gulf of Mexico. Seismic data courtesy of WesternGeco. (B) Perdido fold belt, central Louann salt basin, northwestern Gulf of Mexico. Seismic data courtesy of CGGVeritas. (C) Central Louann salt basin, northeastern Gulf of Mexico. Seismic data courtesy of Fugro. (D) Yucatan salt basin. Seismic data courtesy of Gulf Basin Depositional Systems project, University of Texas Institute for Geophysics. LOC = limit of normal oceanic crust; VE = vertical exaggeration.

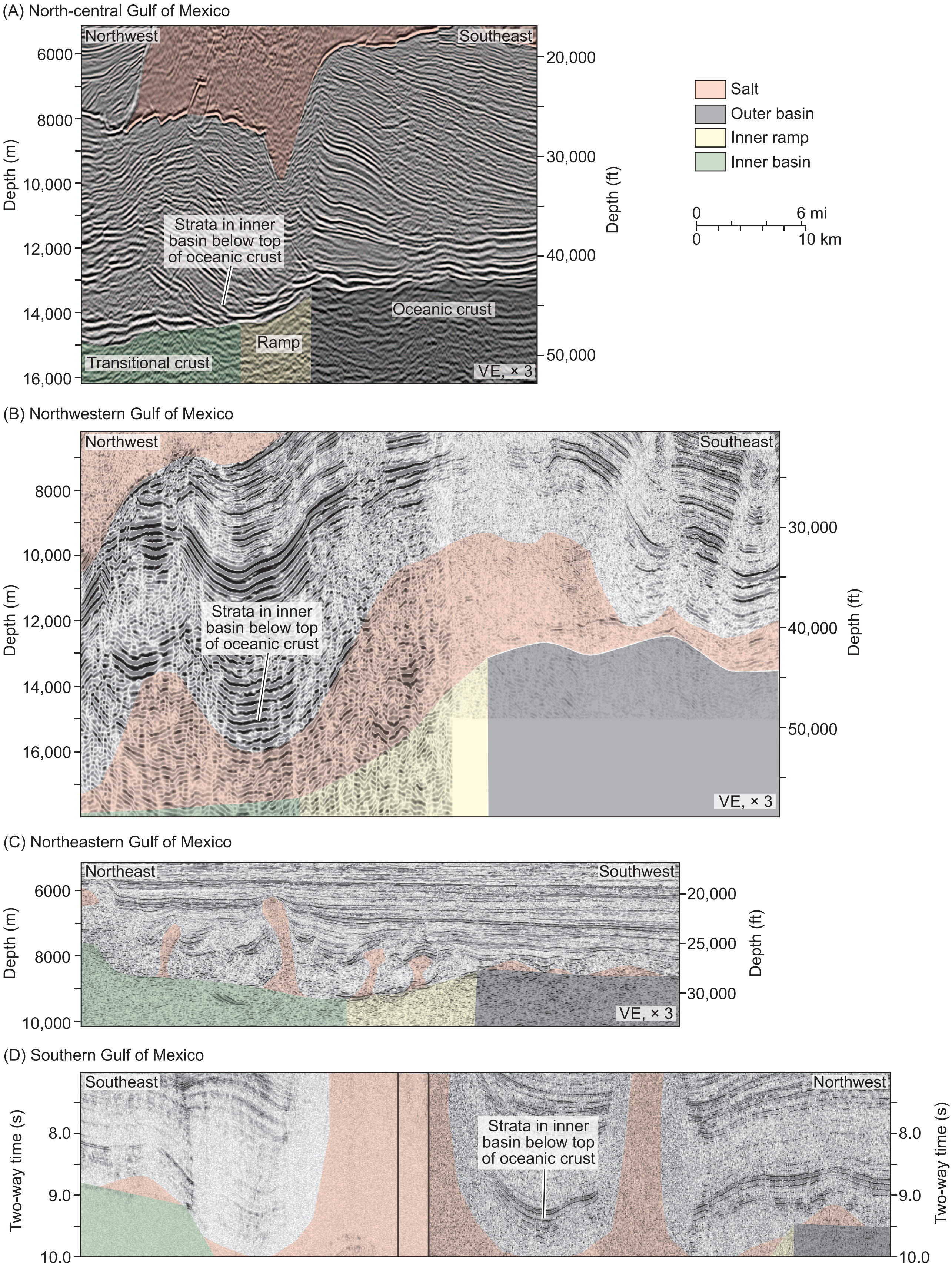


Figure 6. Interpreted versions of seismic sections in Figure 3. VE = vertical exaggeration.

presumably because of the great thickness of sediment there. We look forward to the more sophisticated use of potential-field data in the future, which may further resolve the LOC in the deeper parts of the basin.

Deep-Salt Provinces in the Deep-Water Gulf of Mexico

As described in more detail in the Jurassic Evolution section, we consider deep salt in the Gulf of Mexico to be autochthonous, parautochthonous, or allochthonous, depending on location (Figure 4). Given this variability, phrases such as "autochthonous salt" or "Louann Salt layer" lose some of their use (Pindell and Kennan, 2007). We prefer the generic term "deep salt" for the deepest salt layer because it has no genetic implications.

Seaward parts of the central Louann and Isthmian salt basins can be divided into five provinces on the basis of configuration of the deep salt. These provinces are, from seaward to landward, the outer plain, outer ramp, outer basin, inner ramp, and inner basin (Figures 4, 8). All areas include the outer plain and the inner basin, but the other salt provinces are present only locally.

Outer Plain

The outer plain consists of subhorizontal sediments resting on the acoustic basement that we interpret as a normal oceanic crust because of its position outboard of the inner ramp (Figure 8). The top of the acoustic basement in the outer plain is typically rugose (Figures 6A, B; 7; 8) but is, in places, more planar (Figure 6C, D). Near the edge of the deep salt, this province lies beneath the continental rise (e.g., Martin and Bouma, 1978; Bryant et al., 1991), but the outer plain passes seaward without structural break onto the Sigsbee abyssal plain in the basin center.

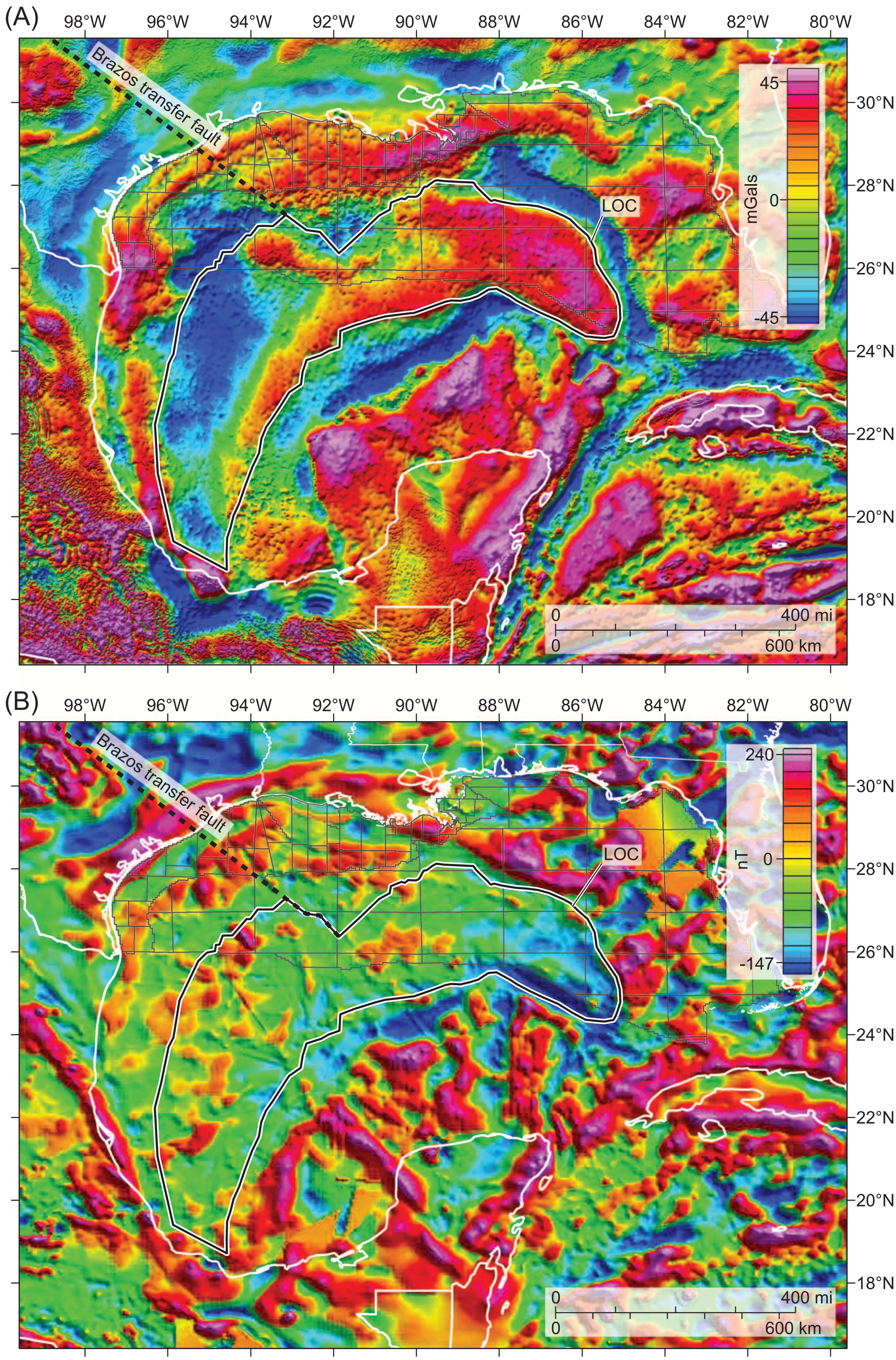
Adjoining both the central Louann and Isthmian salt basins, the landward boundary of the outer plain is the outer edge of deep salt (Figure 8). In most places, the edge of the outer plain is a salt-cored, frontal anticline (Figure 9). Where the edge of salt is poorly imaged (such as beneath some parts of the shallower Sigsbee canopy), the edge of salt has been mapped along the seaward edge of this frontal anticline.

Outer Ramp

In the outer ramp, salt overlies reflectors that can be traced onto the outer plain (Figures 4, 9). Subsalt strata appear undeformed, suggesting that no salt layer is beneath them. These reflectors can be traced to offshore Florida, where they overlie autochthonous Louann Salt (Figure 10). From this geometry, Peel (2001) inferred that salt on the outer ramp is allochthonous, an interpretation we support. Consistent with this interpretation, the seaward edge of the outer ramp is lobate in plan view (Figure 8), like the Pliocene–Pleistocene salt canopy forming the Sigsbee Escarpment. The outer ramp exists only in the north-central and northeastern parts of the deep-water Gulf of Mexico (Figure 8).

The base of the salt in the outer ramp is composed of alternating segments that are parallel to subsalt bedding (flats) and oblique to subsalt bedding (ramps). At the landward end of the province, the base of the salt rises from acoustic basement on a major ramp that typically climbs 600 to 1000 m (1969–3281 ft) upsection at a dip of 8° to 15°. Seaward of this ramp, most of the base of the salt is a flat, typically dipping landward parallel to subsalt bedding at 0° to 3°. This flat may be interrupted by smaller ramps that climb farther upsection at 3° to 8°. At the seaward end of the outer ramp, the base of the salt is typically 800 to 1500 m (2625–4921 ft) above the acoustic basement. However, unpublished seismic interpretations suggest that parts of the outer ramp in the Atwater fold belt were reactivated as thrusts during Miocene shortening, which has locally carried the base of the salt up as high as 2800 m (9186 ft) above the basement.

We lack well data to directly date the section beneath the ramp, but regional correlations suggest that the base of allochthonous salt on the outer ramp typically cuts upsection into middle or Upper Cretaceous strata (Figure 10; Fiduk et al., 2007). The only exception to this level of emplacement is in the Atwater fold belt, where thrust reactivation has locally emplaced salt in the outer


(A)
98°W
96°W
94°W
92°W
90°W
88°W
86°W
84°W
82°W
80°W
30°N
28°N
26°N
24°N
22°N
20°N
18°N
Brazos transfer fault
LOC
mGals
45
0
-45
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0
400 mi
0
600 km

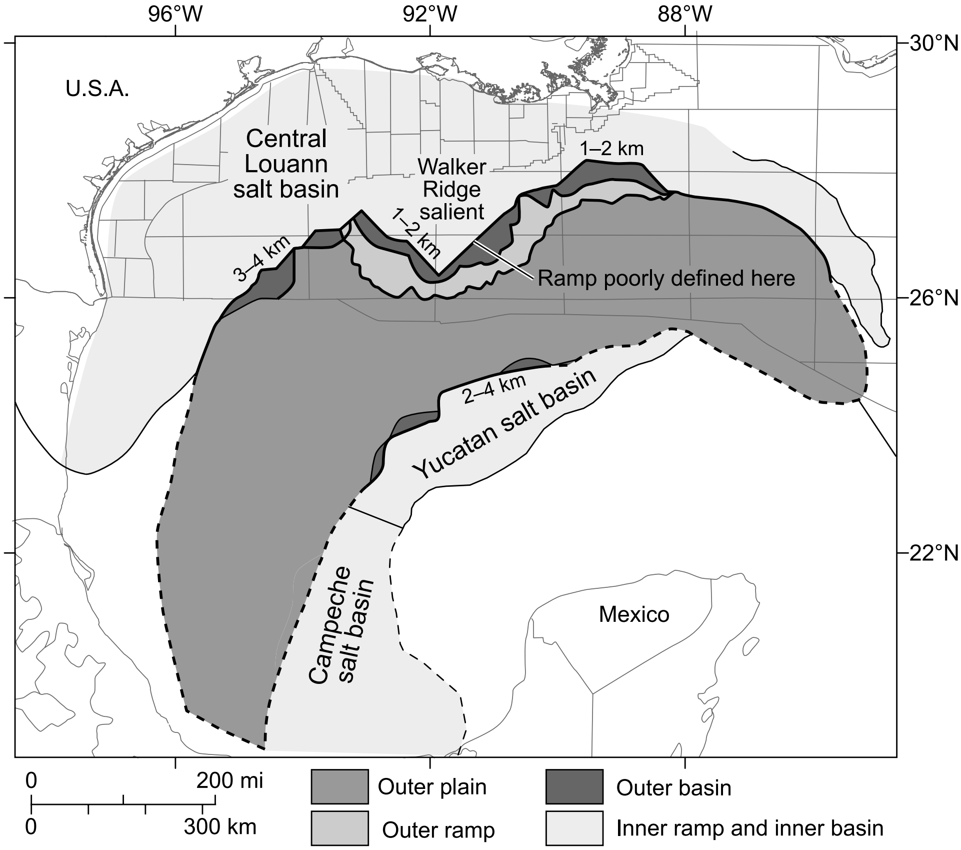


Figure 8. Map of the Gulf of Mexico showing salt provinces. The inner-ramp province is too narrow to map at this scale, so the seaward edge of the outer basin is shown as a solid line where the inner ramp exists and a dashed line where it does not. Where it exists, the height of the inner ramp is annotated.

ramp above units as young as Oligocene (unpublished seismic interpretation).

The root of the salt allochthon defines the landward edge of the outer ramp. This ramp is typically overlain by a landward-dipping panel of sediments (Figure 4). Sediments inboard of this ramp may lie kilometers below the regional level on the outer plain. We used this ramp in suprasalt sediments as a proxy for the base-salt ramp where the base-salt ramp could not be mapped directly because of poor imaging. Even with this technique being used, however, the boundary between the outer ramp and the outer basin was challenging to map. We anticipate significant revisions to the location of this boundary once better seismic data become available.

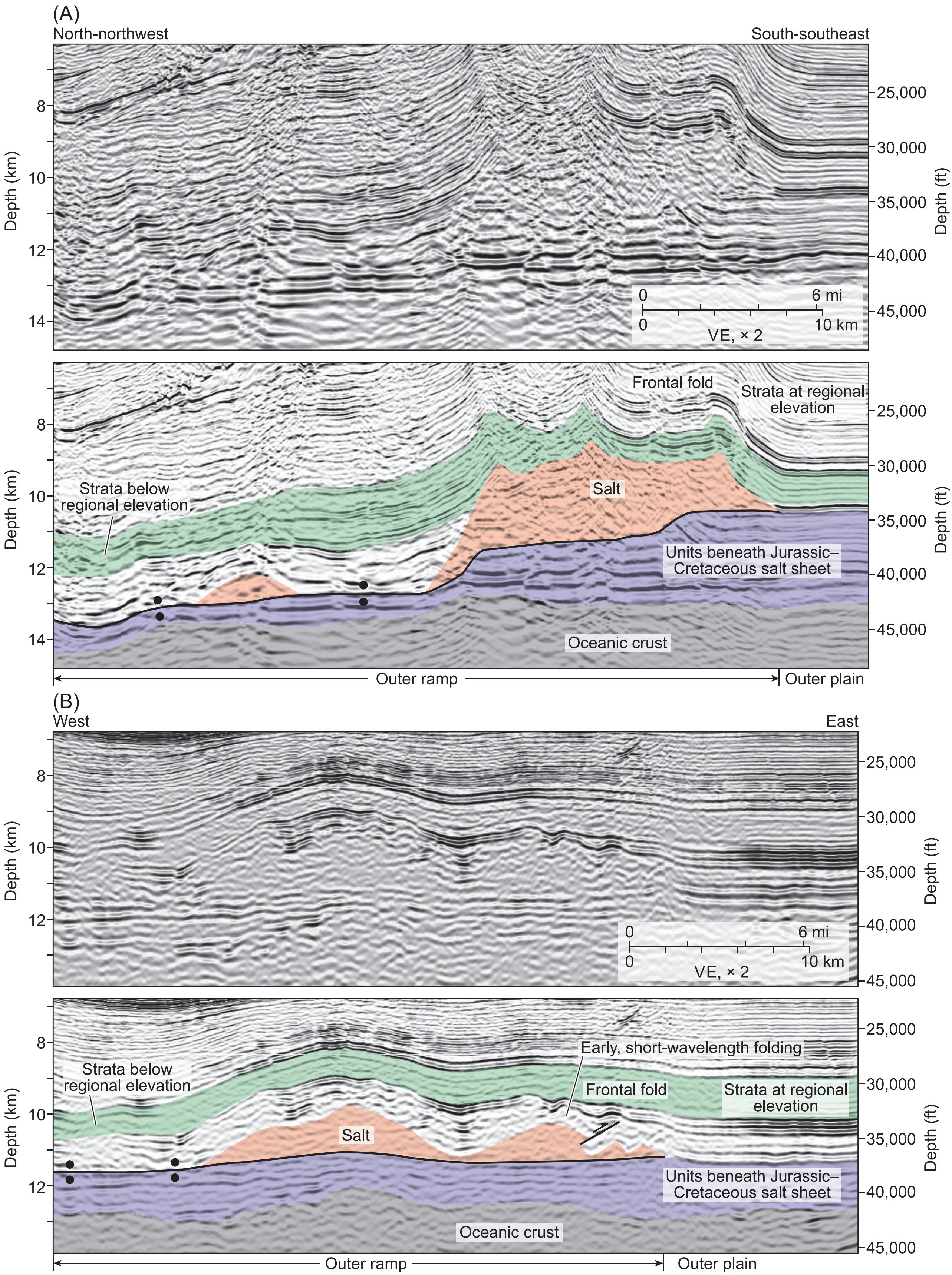
Salt in the outer ramp can form anticlines or small diapirs. These anticlines are productive in the Atwater fold belt and in the Paleogene play (Meyer et al., 2005; Hudec et al., 2013, this issue). Some anticlines have early short-wavelength folding, overprinted by longer-wavelength deformation (Rowan et al., 2000; Figures 4; 9B).

Outer Basin

The outer basin, where salt sits directly on the acoustic basement, is seaward of the inner ramp (Figures 4, 6, 8). Basement beneath the outer basin is typically 12 to 15 km (7–9 mi) below sea level (Figure 6).

The seaward edge of the outer basin is typically poorly imaged on seismic data. This seaward edge appears lobate, whether the outer basin adjoins the outer ramp or the outer plain (Figure 8). The landward edge of the outer basin is at the top of the inner ramp, which is angular in plan view.

Figure 7. Potential-field maps of the Gulf of Mexico Basin. (A) Isostatic gravity, processed using a replacement density of 2000 kg/m^3. Gravity data from Sandwell and Smith (1997); bathymetry from Smith and Sandwell (1997); and isostatic calculations, using the algorithm of Fullea et al. (2008). (B) Total-intensity magnetic data. Coastlines and political boundaries in white. Data from Maus et al. (2007). See Hudec et al. (2013, this issue), for a discussion of evidence of the Brazos transfer fault.



(A)
North-northwest
South-southeast
Depth (km)
Depth (ft)
8
10
12
14
25,000
30,000
35,000
40,000
45,000
0
6 mi
0
10 km
VE, × 2
Frontal fold
Strata at regional elevation
Strata below regional elevation
Salt
Units beneath Jurassic–Cretaceous salt sheet
Oceanic crust
Outer ramp
Outer plain
(B)
West
East
Early, short-wavelength folding
Frontal fold
Strata at regional elevation
Strata below regional elevation
Salt
Units beneath Jurassic–Cretaceous salt sheet
Oceanic crust
Outer ramp
Outer plain

The style of salt tectonics in the outer basin varies. In the northwestern deep-water Gulf of Mexico, large compressional anticlines of the Perdido fold belt have formed (Figure 6B; Camerlo and Benson, 2006; Hudec et al., 2013, this issue). Elsewhere, most structures in the outer basin are small relict salt pillows, turtle structures, and salt diapirs. Away from the diapirs and salt-cored folds, salt is welded over most of the outer basin. In the northern Gulf of Mexico, because evacuation of salt into diapirs caused overlying stratigraphic units to subside, most are well below their regional levels on the nearby outer plain (Hall, 2002; Fiduk et al., 2007).

Inner Ramp

Salt in the inner ramp rests on the acoustic basement, which dips landward at 4° to 10° (Figure 6). Ramp dips are steepest in the northwestern Gulf of Mexico, near the Perdido fold belt. Dips are gentler farther east on the east flank of the Walker Ridge salient, where the inner ramp is not much steeper than the outer basin. The inner ramp forms a step in the base of salt in the deep-water Gulf of Mexico, with an estimated total relief of 1 to 4 km (0.62–2 mi). Relief of the inner ramp cannot be measured everywhere because its base is too deep (>15 km [9 mi]) to be imaged on most seismic data. Uncertainty in the location of the base of the ramp also means that we can only estimate the width of the province at between 5 and 30 km (3 and 19 mi) in the study area. The inner ramp is grouped with the inner basin on the province map (Figure 8).

Because the seaward edge of the inner ramp is a prominent hinge in the acoustic basement, it is easily mapped from good seismic data (e.g., Figure 6A, C). However, in parts of the northern Gulf of Mexico, because this hinge lies at depths of 12 to 15 km (7–9 mi) and is obscured by the shallower Sigsbee salt canopy, it can be difficult to locate. Three techniques can help locate the hinge, however, even where poorly imaged. First, the inner ramp is typically draped by a major landward-dipping monocline that drops sediments below the level of the basement in the outer basin (Figure 6). This monocline can lower stratigraphic units by several kilometers and is more readily imaged than the basement. Second, in some areas, seaward-leaning diapirs are rooted at the top of the ramp. In these areas, projecting the diapirs (or, more commonly, the leaning weld) to depth can approximately locate the top of the inner ramp. Third, salt and salt-related structures are typically much larger in the inner ramp and the inner basin than in the outer basin because of an initially greater salt thickness. An increase in diapir size or in the amplitude of overburden structures therefore suggests transition into the inner ramp and the inner basin. The combination of these techniques allows us to map the seaward edge of the inner ramp beneath the Sigsbee canopy, although we anticipate revisions as seismic data improve.

Inner Basin

Because of its great depth, the top of the acoustic basement in the inner basin is poorly imaged over much of the deep-water Gulf of Mexico, especially beneath the Sigsbee canopy. Where imaged, the basement in the inner basin appears fairly smooth, although this smoothness could be the result of sparse signal and plentiful noise. Given regional changes in stratal elevations above deep salt, the basement in the inner basin dips gently landward near the inner ramp, as would be expected after flexural loading. Farther inboard, basement climbs landward toward the onshore pinch-out of the salt basin (e.g., Diegel et al., 1995; Peel et al., 1995; McBride, 1998; Figure 4).

Top-Salt Elevation at the End of Salt Deposition

Warren (2006) argued that basinwide megahalite deposits such as the Gulf of Mexico salt basin accumulate in preformed tectonic depressions

Figure 9. Seismic sections showing the outer ramp in the deep-water Gulf of Mexico. Allochthonous salt is emplaced above units deposited on the oceanic crust. Both images are from the central Louann Salt basin, north-central Gulf of Mexico. Seismic data courtesy of CGGVeritas. VE = vertical exaggeration.

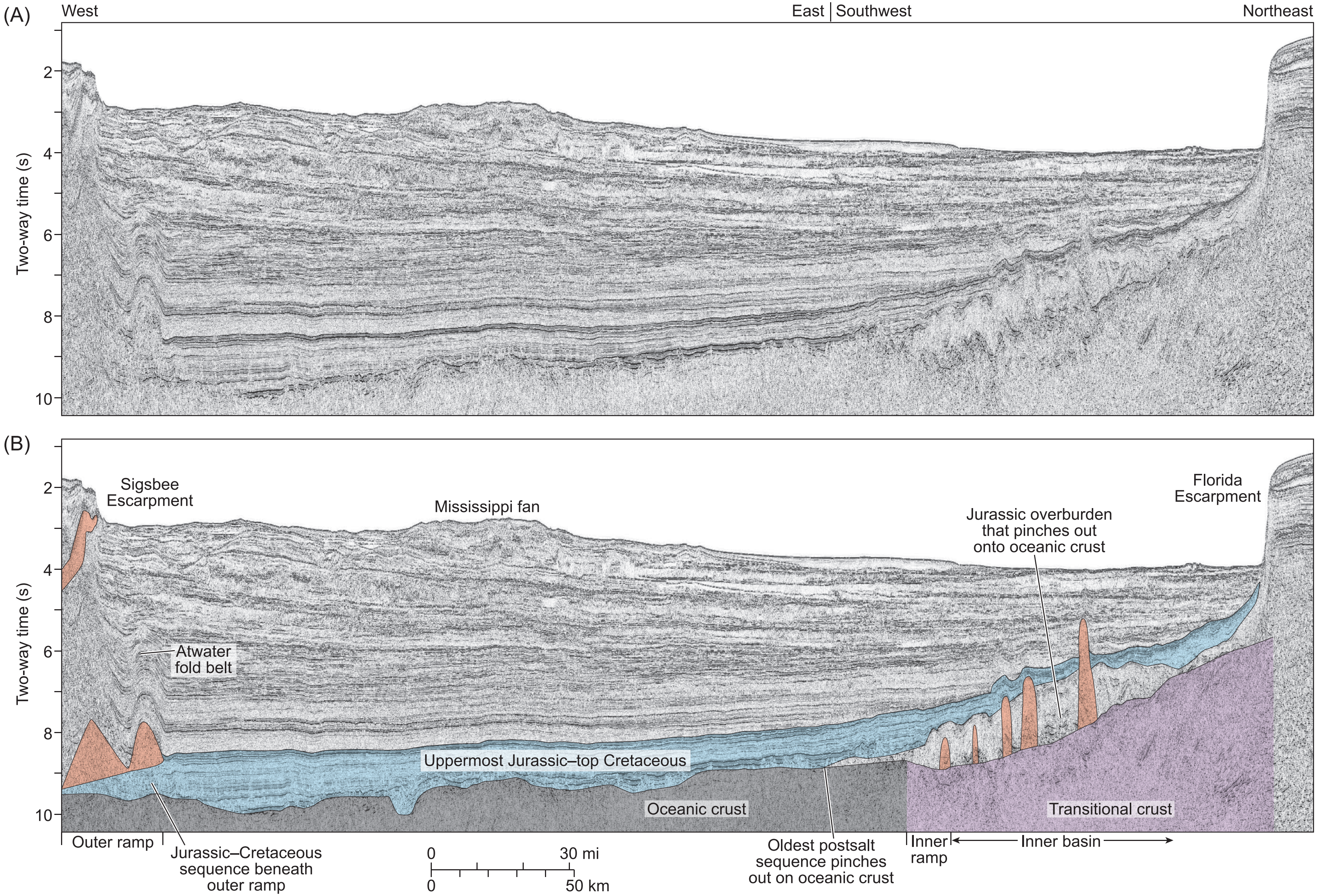


Figure 10. Uninterpreted (A) and interpreted (B) seismic section showing the correlation of units from the Florida Escarpment to the Atwater fold belt. The uppermost Jurassic–top Cretaceous (blue) megasequence is underlain by an older postsalt sequence (uncolored) that pinches out on the oceanic crust. In the Atwater fold belt, postsalt units clearly project beneath the outer ramp, proving that the outer ramp is allochthonous. Seismic data courtesy of Fugro.

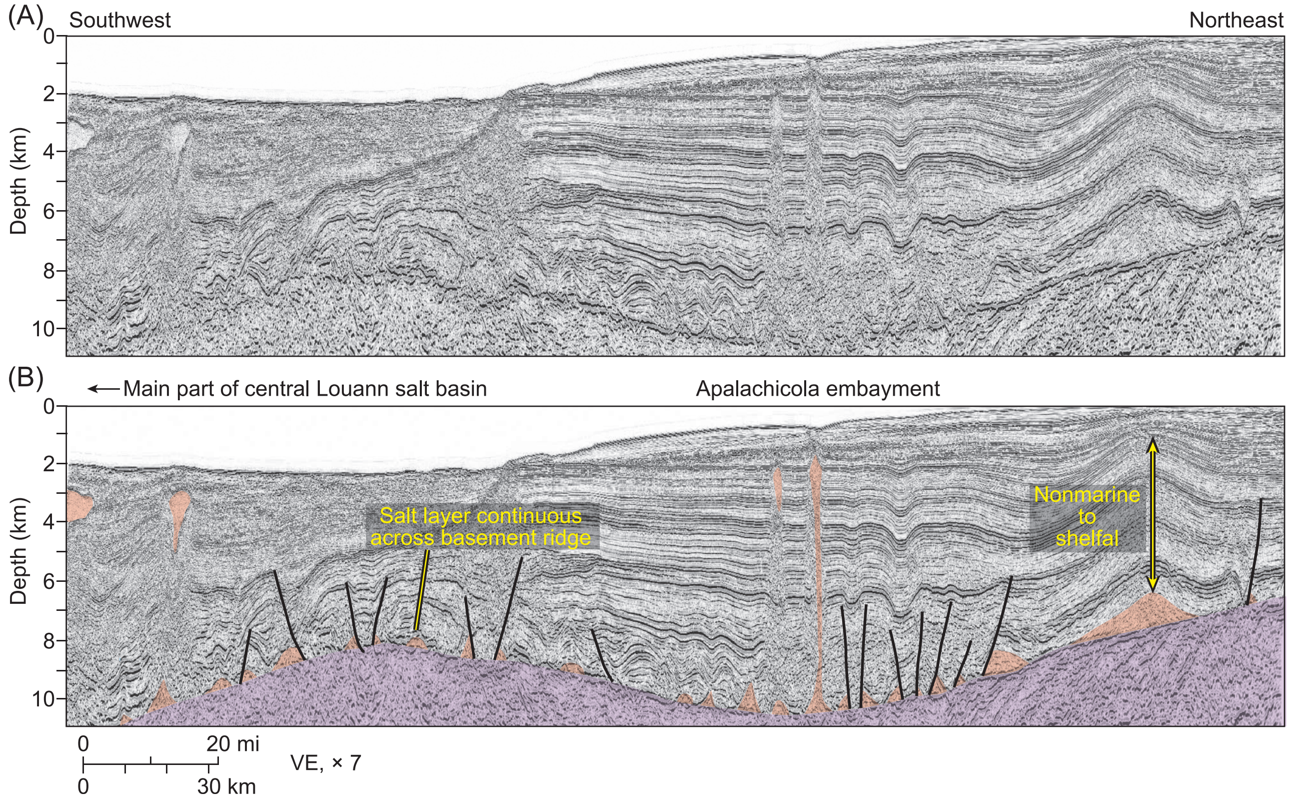


Figure 11. Uninterpreted (A) and interpreted (B) seismic section showing a continuous salt layer or salt weld connecting the Apalachicola embayment to the rest of the central Louann salt basin. Nonmarine to shelfal deposits in the Mesozoic sequence above the salt in the Apalachicola embayment implies that salt filled the basin to near sea level. Seismic data courtesy of Fugro. VE = vertical exaggeration.

below sea level, which act as evaporite sumps. These sumps are fed by marine water (seepage and ephemeral surface overflow) and by continental runoff. An evaporite sump must be isolated from the world ocean to become saline enough to deposit halite, aided by a favorable latitude and climate and by adiabatic heating and drying because of its low elevation. The preformed depression accommodates thick salt and avoids the need for any sudden unexplained basin subsidence as salt accumulates.

A basinwide evaporite deposit continues to fill with salt as long as it (1) has accommodation space, (2) has sources of saline water, and (3) is isolated from the world ocean. Deposition can continue until the basin fills with salt to world sea level or above or until the ocean rushes in when the basin is only partly filled with salt, such as for the massive halite in the Messinian Mediterranean Basin (Hsu et al., 1973; Warren, 2006). If normal seawater floods an evaporite basin only partly filled with salt, deep-marine facies should be deposited directly above the salt. In contrast, if the basin were to fill with salt to near sea level when the world ocean reconnects, then the first facies above the salt should be shelfal to nonmarine.

In the Gulf of Mexico salt basin, most evidence of the paleodepth of sediments deposited on the Callovian salt is from the interior salt basins of the United States. Oxfordian–Kimmeridgian successions in the East Texas, North Louisiana, and Mississippi salt basins are dominated by continental deposits, shallow-marine sequences, and evaporites, suggesting that salt filled these basins to near sea level (e.g., Ahr, 1973; Presley, 1984; Salvador, 1987; Prather, 1992; Mancini et al., 2004). However, these observations do not necessarily mean that the salt in the central Louann salt basin filled to sea level because the interior basins may have been perched above the salt in the central basin. Perched marginal basins formed as Mediterranean sea level started to fall just before Messinian salt deposition in the central basins (Butler et al., 1995; Clauzon et al., 1996; Westbrook and Reston, 2002; Warren, 2006).

Most information on the deposition of Oxfordian–Kimmeridgian strata in the central Louann salt basin is from the Apalachicola embayment in the northeastern part of the basin (Figure 11). Salt here was overlain first by the Oxfordian continental to shallow-marine Norphlet Formation (Marzano et al., 1988). This succession provides strong evidence that the Louann Salt deposition ended near sea level and was not drowned beneath hundreds of meters of water. Norphlet deposition ended in a marine transgression (Mancini et al., 1985), but a carbonate shelf margin prograded back across the Apalachicola embayment during the late Oxfordian and the Kimmeridgian (Dobson and Buffler, 1997). The rapid renewal of shelfal conditions in the area implies that water was never deep, supporting the interpretation that salt filled the central Louann salt basin to near sea level.

Oxfordian–Kimmeridgian Suprasalt Extension Not Balanced by Downdip Shortening

In both offshore Florida and offshore Yucatan, seismic data show tens of kilometers of salt-detached extension in Jurassic strata (Figure 12). We interpret most of the growth on these faults to be Oxfordian–Kimmeridgian. The seaward ends of both margins contain active diapirs that may have been squeezed by salt-detached shortening, but the magnitude of shortening appears to be only a small fraction of the extension recorded farther updip. Also, because much of the shortening postdates the extension, it is not kinematically related.

The gross imbalance between detached shortening and detached extension is evident everywhere alongstrike in both areas and does not appear to be an artifact of imaging problems. This imbalance is a kinematic problem and must be explained by any hypothesis on the Jurassic evolution of the Gulf of Mexico.

JURASSIC HISTORY AND PLATE TECTONICS OF THE LOUANN SALT BASIN

This section presents a hypothesis for the rifting and opening of the Gulf of Mexico Basin (Figures 1, 13, 14). The Jurassic history of the basin has four phases: (1) presalt rifting, (2) salt deposition, (3) postsalt crustal stretching, and (4) sea-floor spreading.

Phase 1: Presalt Rifting (210–163 Ma)

The Gulf of Mexico Basin began to open during the Late Triassic as grabens rimmed the basin (Salvador, 1991b). These grabens filled with Upper Triassic–Lower Jurassic redbeds and volcanic rocks (Eagle Mills Formation and equivalents).

Rifting in the Gulf of Mexico Basin was part of the breakup of the supercontinent Pangea as the Yucatan microplate moved southeastward away from North America (Figure 13A, B). Magnitudes of extension vary with position, but our plate reconstruction (Appendix) suggests that presalt extension of 200 to 250 km (124–155 mi) was widespread.

The nature of the hyperextended lithosphere produced by such large magnitudes of extension is contentious. Candidates include exhumed mantle (e.g., Unternehr et al., 2010), extremely thinned continental crust (Huismans and Beaumont, 2011), and transitional crust produced by ultraslow spreading (Kneller and Johnson, 2011). For our purposes, this distinction is unimportant; most of the salt in the central Louann salt basin was deposited on transitional hyperextended lithosphere.

Phase 2: Salt Deposition (163–161 Ma)

Compared with that of some other salt basins, salt deposition in the Gulf of Mexico is poorly dated. Evidence bearing on the age of the Louann Salt was discussed by Salvador (1991b), whose work we summarize here. Because the Louann Salt has no diagnostic fossils, the salt can be bracketed only between the ages of underlying and overlying units. In the northern Gulf of Mexico, because the presalt Eagle Mills Formation contains basalt dikes as young as 180 Ma, the salt must be post–Early Jurassic in age. Above the salt, the Norphlet Formation is nonfossiliferous, but the Smackover Formation above the Norphlet contains upper Oxfordian ammonites. Salt must therefore have been deposited between 180 and 156 Ma. On the basis of

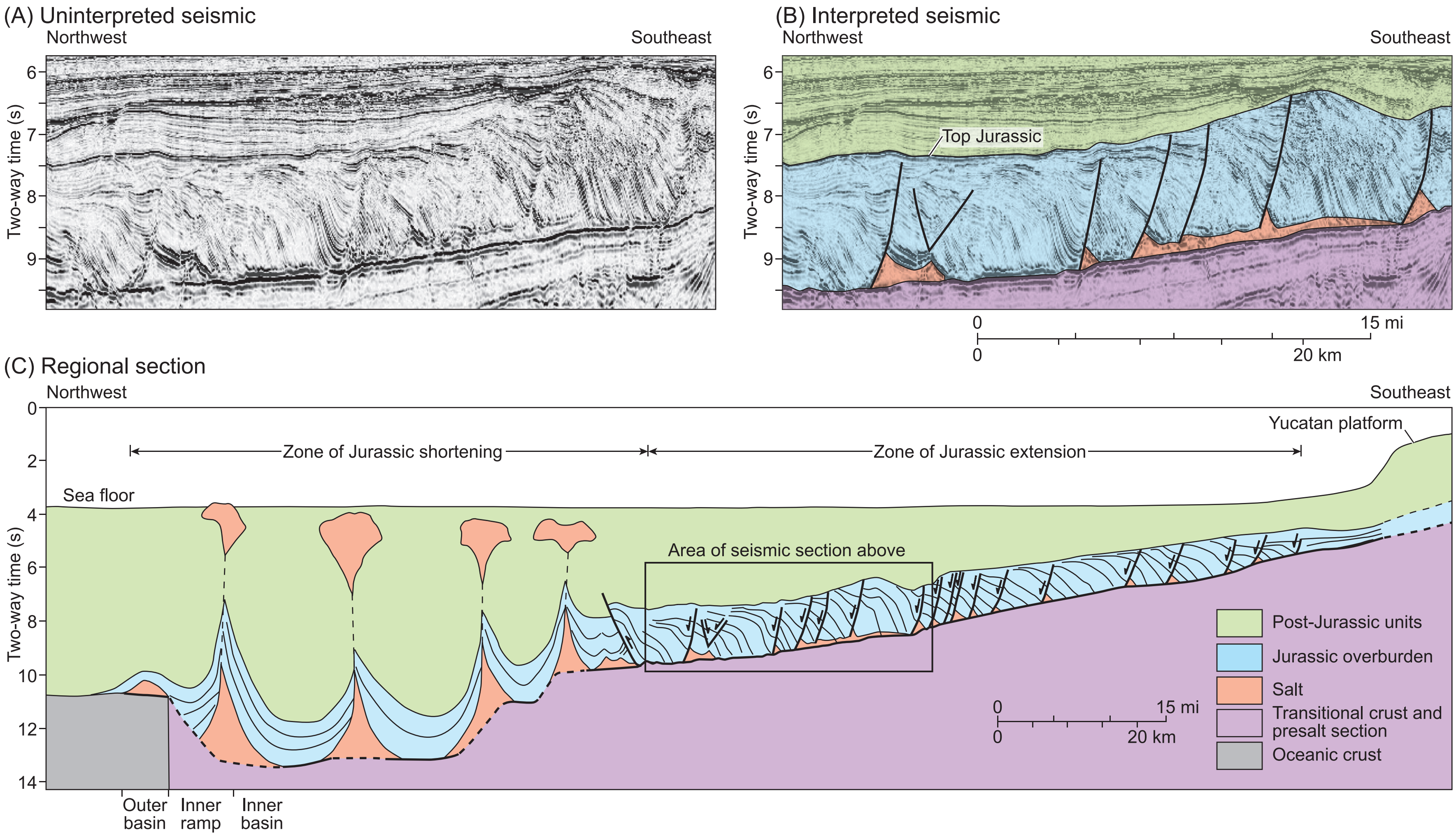
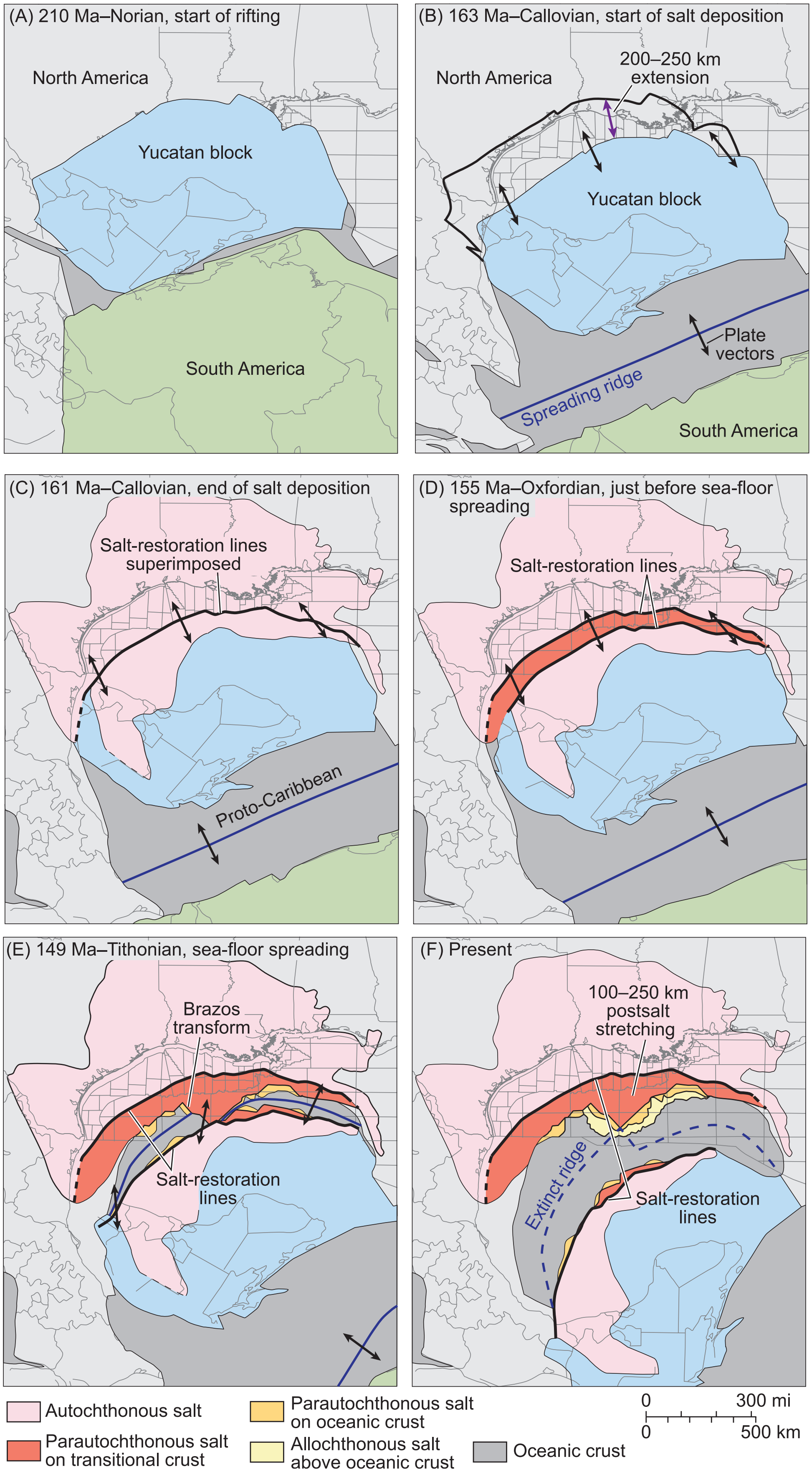


Figure 12. Salt-detached Jurassic extension unbalanced by salt-detached shortening in the Yucatan salt basin. (A and B) Uninterpreted and interpreted seismic sections showing detached extension in the Yucatan salt basin. (C) Simplified regional section based on seismic data. The lack of well data in the basin prohibits correlation across faults, but we estimate 40 to 70 km (25–43 mi) of salt-detached extension on the basis of rollover geometries. Jurassic shortening is likewise difficult to calculate, but according to the estimates of the widths of the diapirs before they were shortened, shortening is less than 10 km (6 mi). Seismic data courtesy of Petróleos Mexicanos (PEMEX).

Figure 13. Sequential plate restoration using PaleoGIS software. See Appendix for methodology.



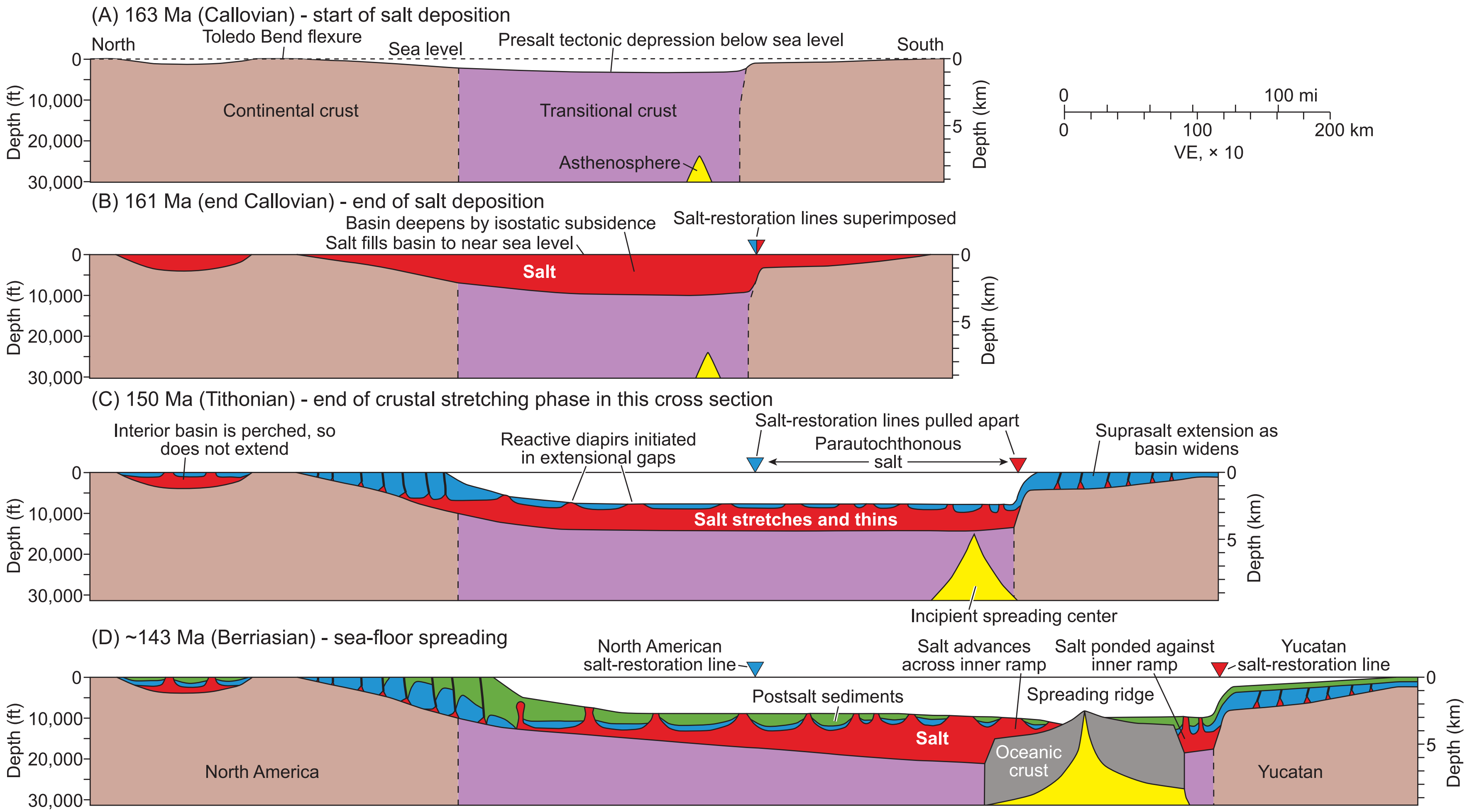


Figure 14. Schematic section restoration of basin evolution, emphasizing salt and its overburden. Internal structures in the crust are not shown. Line of section passes through the Walker Ridge salient, where sea-floor spreading began relatively late. Restoration was constructed using LithoTect software. VE = vertical exaggeration.

evaporites in the upper Bathonian or lower Callovian Huehuetepec Formation and calcarenites in the Callovian Tepexic Formation of central Mexico, Salvador (1991b) further narrowed the age window for salt deposition to the Callovian (165–161 Ma). A Callovian age is also consistent with the volcanic xenoliths in Louisiana salt diapirs, dated at 158 to 160 Ma (Stern et al., 2011). These xenoliths probably come from magmas intruded into the salt during or after deposition. Age constraints on the salt are therefore loose.

The duration of salt deposition is likewise poorly known, although analogies with better-dated basins suggest that megahalite can accumulate extremely rapidly. For example, the 1.5-km (0.9-mi)–thick Messinian salt in the Mediterranean Sea was deposited in less than 0.3 m.y., yielding an average deposition rate of 0.5 cm/yr (Krijgsman et al., 1999). In the Danakil depression (Ethiopia), alternating halite and clay interbeds accumulated to a thickness of more than 1 km (0.62 mi) in 10 k.y. (10 cm/yr; Warren, 2006). On the basis of the amount of salt currently present in the Sigsbee canopy, we estimate at least 3 to 4 km (1.9–2.5 mi) of depositional salt in the thickest parts of the central Louann salt basin. Given the salt depositional rates evident in other basins, this thickness could have been deposited in less than 1 m.y. In our model, we assume 2 m.y. for the salt to be deposited, a conservative estimate (Figure 1).

By analogy with inferences from other basin-wide evaporites (Warren, 2006), salt probably began to accumulate in a preexisting tectonic depression below sea level. We lack evidence whether this basin was marine immediately prior to salt deposition or whether it had remained isolated from the world ocean since rifting began. If it had been marine, salt deposition would have been triggered by isolation from the world ocean and evaporative drawdown of the remaining water in the sump.

Because adiabatic heating increases temperatures roughly 1°C for every 100-m (328-ft) drop in elevation, temperatures at the bottom of a 1-km (0.62-mi)-deep depression would have been 10°C higher than at sea level. Humidity also falls with adiabatic heating—ideal conditions for seawater to evaporate. Salt precipitated rapidly in this environment, fed by saline springs, ephemeral marine washover, and continental runoff.

Accumulating salt loaded the crust so that salt was partly accommodated by isostatic subsidence of the underlying crust. Assuming isostatic equilibrium and reasonable densities for asthenosphere and salt, roughly two-thirds of the final thickness of the salt would have been accommodated by isostatic subsidence. By analogy with rates of isostatic adjustment during postglacial rebound of northern Europe (McConnell, 1968), this subsidence could have occurred over only tens of thousands of years, an order of magnitude faster than our estimated duration for salt deposition. Subsequent salt flow has obscured the original thickness of the Louann Salt, but depositional thicknesses of 3 to 4 km (1.8–2.5 mi) in the deeper parts of the basin would have required an initial depression that was 1 to 1.13 km deep. Salt eventually filled this depression to near sea level.

Phase 3: Postsalt Crustal Stretching (161 to 154–149 Ma)

Most previous workers in the Gulf of Mexico have assumed either (1) that sea-floor spreading began prior to Louann Salt deposition, so that some salt was deposited on the oceanic crust (Hall et al., 1982; Hall, 1995; Imbert and Philippe, 2005; Pindell and Kennan, 2007), or (2) that spreading began near the end of the Louann deposition (e.g., Buffler et al., 1980, 1981; Buffler and Sawyer, 1985; Pindell, 1985; Buffler, 1989; Salvador, 1991a; Kneller and Johnson, 2011). However, firm evidence is absent concerning the age of oceanic crust in the Gulf of Mexico. All of these estimates are based on plate reconstructions or on the assumption that the onset of sea-floor spreading terminated salt deposition.

In the absence of isotopic or magnetic-reversal data, the age of the oceanic crust may be estimated by dating the oldest sediments resting on that crust. The oldest sediments lie atop the oldest oceanic crust and onlap progressively against younger crust at the spreading ridge. Any sediments deposited before sea-floor spreading began are restricted to areas adjoining the oceanic crust (Figure 15). Dating

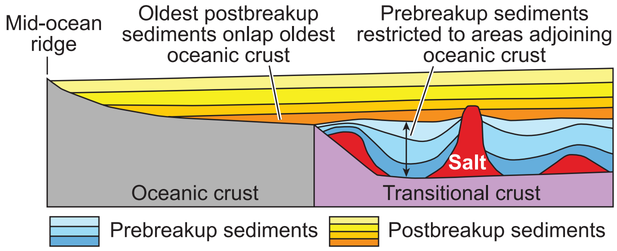


Figure 15. Schematic section showing the relationship of sediment distribution to the onset of sea-floor spreading. Sediments deposited prior to breakup are restricted to basins on either side of the oceanic crust, whereas sediments deposited after breakup onlap the oceanic crust.

the onset of sea-floor spreading therefore involves dating the oldest sediments deposited on the normal oceanic crust near the LOC.

In the eastern Gulf of Mexico, the oldest seismic horizon that we can map confidently onto the normal oceanic crust is the base of the blue megasequence in Figure 10. Well data indicate that the base of the blue megasequence is Tithonian (uppermost Jurassic). Sea-floor spreading must therefore have begun prior to the end of the Jurassic. We estimate from the minimal thickness of the sub-blue-megasequence section on the oceanic crust that the oldest units deposited on the oceanic crust in Figure 10 are also Tithonian or only slightly older. Figure 10 also suggests that a significant thickness of Jurassic overburden is absent on the oceanic crust. Comparing Figures 10 and 15, we therefore argue that sea-floor spreading began significantly after salt deposition (Figure 1).

In our plate-tectonic reconstruction, continental rifting continued for 7 to 12 m.y. after salt deposition ended (Figures 1; 13C–E; 14B, C). We refer to this tectonic phase as postsalt crustal stretching. Because LOCs in the northern and southern Gulf of Mexico are not parallel, continental separation must have been diachronous, beginning in the Kimmeridgian in the eastern and western Gulf of Mexico, but not until the early Tithonian in the central Gulf of Mexico. During postsalt crustal stretching, most of the salt basin widened by 100 to 250 km (62–155 mi), and extension tapered to zero at the edges of the basin (Figure 13C–F). Salt stretched and thinned as it flowed laterally to fill the widening basin, as much as water would flow laterally in a swimming pool whose walls were moving apart (Figure 14B, C).

Another way to picture the kinematics is to envision salt flowing continuously into the gap between the two halves of the basin as they drew apart. Pindell and Kennan (2007) termed the imaginary line separating the Louann and Isthmian salt basins at the end of salt deposition the "salt-fit isochron." We follow this concept but prefer the term "salt-restoration line," which we think is more descriptive. A single salt-restoration line at the end of salt deposition separates into two lines as the severed halves of the basin move apart (Figures 13C–F; 14B–D). Any salt seaward of these lines must have flowed laterally to its current position because the underlying basement did not exist at the time of salt deposition. Because salt-restoration lines are theoretical constructs, they are not mappable on seismic data. They are best considered as a reference framework useful for displaying the lateral flow of salt required by a given plate model. Because their positions are artifacts of the plate restoration, they are subject to the same uncertainties. Considerable latitude to move salt-restoration lines in both time and space as plate restorations change exists, and assumptions about the age of salt deposition evolve.

During postsalt crustal stretching, salt advanced across newly formed basement as fast as this basement was created. Because there was no time to deposit sediments on top of the new crust before it was buried beneath spreading salt, salt rests directly on the acoustic basement. Salt resting on the basement seaward of the salt-restoration line is thus parautochthonous, in that it rests on the basement but lies beyond the limits of the autochthonous salt basin ("para" derived from the Greek for beside, near, or beyond). This usage also meets the American Geological Institute Glossary definition of parautochthonous: "said of a rock unit that is intermediate in tectonic character between autochthonous and allochthonous" (Jackson, 1997).

A period of postsalt crustal stretching helps explain two puzzling observations. The first puzzle is the ability of the inner ramp to block seaward salt flow in approximately half of the basin (Figures 6C–D; 8; 10; 12). Data from the Woodlark basin (Taylor et al., 1999), Red Sea (Bonatti, 1985; Martinez and

Cochran, 1988), and Gulf of California (Lizarralde et al., 2007) suggest that, in the absence of a hot spot, the tip of a propagating mid-ocean ridge forms approximately 2.6 km (1.6 mi) below sea level. Therefore, the top of the inner ramp, which is now level with the top of the oceanic crust, originally lay at least 2 km (1.2 mi) below sea level. The top of the salt must have been deeper than this to pond behind the ramp. How could the top of salt, which was originally near sea level, subside to more than 2.6 km (1.6 mi) in depth before sea-floor spreading began? We follow Pindell and Kennan (2007) in inferring that salt deposition ceased long before the inner ramp formed so that the salt thinned as the underlying transitional crust stretched before the inner ramp formed (Figure 14).

The second puzzle explained by postsalt crustal stretching is the imbalance between salt-detached Jurassic extension and shortening. If the entire basin widened by 100 to 250 km (62–155 mi) during postsalt crustal stretching, then the postsalt section should also be stretched without the need for any compensating shortening (Figure 14C). The first bed above the salt should exhibit the full 100 to 250 km (62–155 mi) of extension unbalanced by shortening; younger sediments would be less extended. In the interior basins, salt was tectonically isolated from the salt in the central Louann salt basin and so remained perched while the salt in the central basin stretched and thinned (Figure 14).

To date, salt-detached Oxfordian–Kimmeridgian extension uncompensated by downdip shortening has been seismically imaged only in the northeastern Gulf of Mexico and the Yucatan salt basin. However, similar provinces should exist around the rest of the rim of the central Louann salt basin if indeed the crust continued to stretch after salt had been deposited. Because uncompensated extension should have the same magnitude as the distance from the salt-restoration line to the LOC, it should vary around the basin rim.

Phase 4: Sea-Floor Spreading (154–149 to 137 Ma)

A striking feature of our plate reconstruction is that the LOCs on the northern and southern sides of the Gulf of Mexico are not mirror images (Figures 2, 7, 8, 13). In particular, the prominent Walker Ridge salient has no corresponding indentation on the Mexican side of the basin (Figure 8). Asymmetric plate boundaries imply that oceanic crust began forming at different times across the basin. Our reconstruction accounts for the asymmetry between the northern and southern LOCs with the postulate that sea-floor spreading propagated in from both the east and west (Figure 13E). This type of geometry in which the limits of normal oceanic crust on conjugate margins are not isochronous is also seen in margins such as Australia-Antarctica (Direen et al., 2012) and Galica-Flemish Cap (Jagoutz et al., 2007), and may be common to many margins but below the resolution of current data.

We thus infer that the onset of sea-floor spreading was diachronous in the Gulf of Mexico, starting in the western and eastern parts of the basin before the basin center broke apart (Figure 13E, F). In our model, a crustal bridge persisted in the Walker Ridge salient until roughly 149 Ma. Delayed breakup of the Walker Ridge salient means that postsalt crustal stretching lasted longest there, so this area should have the most uncompensated extension updip.

In approximately half of the basin, salt ponded behind the inner ramp, indicating that the top of salt was lower than the ramp crest. In the other half of the basin, salt overflowed the top of the ramp seaward onto the oceanic crust, forming the parautochthonous outer basin. Overflow onto oceanic crust may have been the result of a lower inner ramp. Alternatively, continued seaward salt flow resulting from thermal subsidence may have surmounted the inner ramp in some areas.

Where salt overtopped the inner ramp, salt continued to flow out onto the oceanic crust as the Louann and Isthmian salt basins separated, steadily enlarging the outer basin. Because little or no sediment appears to be beneath the salt in the outer basin, the oceanic crust must have been covered by salt soon after the crust formed.

Except in the north-central Gulf of Mexico, the salt in the outer basin continued to flow seaward until the supply was exhausted, leaving a thin flat layer of parautochthonous salt resting on the oceanic crust. This salt veneer was then buried by Upper Jurassic and Cretaceous sediments. Around the Walker Ridge salient, however, salt flowed into the outer basin even after its toe was pinned by sediment. Flow continued probably because this area (1) has the lowest relief on the inner ramp and (2) faces the widest part of the central Louann salt basin, where tilting supplied the most elevation head to drive seaward salt flow. Salt therefore continued to advance beyond the Walker Ridge salient, climbing upsection across sediments accumulating in front of the salt on the oceanic crust. This climb formed an allochthonous fringe of salt in the outer ramp (Figure 9).

The age of the outer ramp allochthon varies across the north-central Gulf of Mexico. In the east, the allochthon continued to advance until near the end of the Cretaceous. Farther west, advance stopped during the middle to Late Cretaceous.

IMPLICATIONS AND CONCLUSIONS

Most of the key elements of this tectonic hypothesis —a bisected salt basin, existence of an inner ramp, and seaward flow of rifted salt to form parautochthonous and allochthonous fringes at the seaward end of the basin—have been proposed before. Our contributions are mapping of the LOC and salt provinces using seismic data and proposal of a long period of crustal extension after salt deposition. These changes have several important implications.

First, mapping shows that LOCs on the northern and southern sides of the basin are not parallel, implying diachronous rifting. In our reconstruction, the Gulf of Mexico has two spreading segments, separated by the Brazos transform (Figure 13E). The eastern segment opened from east to west, leaving the Walker Ridge salient as the final area to separate. The most extended crust and broadest salt basin should therefore be near the salient. Conversely, the least extended continental and/or transitional crust, and oldest oceanic crust, should be on the eastern and western margins of the basin.

Second, the delayed onset of sea-floor spreading allowed the salt and its overburden to thin and stretch before the normal oceanic crust formed. This stretching induced basinwide salt-detached extension, which is likely to have initiated reactive diapirs, even in unusual places (e.g., the modern deep water; Figure 14C). Furthermore, extensional thinning of the salt also caused the overburden to subside. This salt-induced subsidence would have locally amplified the effects of the marine transgression that followed Norphlet deposition, causing water depths to increase more rapidly than by marine influx or eustatic rise alone. This effect would have been especially pronounced where salt was thickest before extension.

Finally, the province map (Figure 8) provides insights about seaward salt flow as the basin opened. Seaward salt flow was greatest near the Walker Ridge salient because (1) the inner ramp was lower there and (2) abundant salt was available because of the wide basin. Furthermore, the province map is a guide to original salt thickness in the deepwater Gulf of Mexico. Salt was thickest in the inner basin, of intermediate thickness in the outer basin, and thinnest in the outer ramp. Because salt thickness is a primary control on structural style in salt basins, the imprint of rifting and basin separation should fundamentally control structural, stratigraphic, and exploration trends in the deep-water Gulf of Mexico. These controls and their results are the subject of our companion article (Hudec et al., 2013, this issue).

APPENDIX: PLATE RECONSTRUCTIONS

The motion of Yucatan relative to North America, which formed the Gulf of Mexico, is only partly constrained by direct observations of spreading ridge and fracture zones in the eastern Gulf of Mexico (Imbert, 2005; Imbert and Philippe, 2005). Much of our reconstruction has therefore to be derived from the motion of surrounding plates and local geometric and timing constraints. The prerift fit of Yucatan against North America is poorly constrained because it requires a quantitative restoration of the several hundred kilometers of crustal extension.

Rifting in the Gulf of Mexico region started in the Late Triassic (Salvador 1991b), at the same time as the rifting between North America and Africa (Schouten and Klitgord, 1994; Kneller et al., 2012). We follow the most common hypotheses for the opening of the Gulf of Mexico (Marton and Buffler, 1994; Watkins et al., 1995; Pindell and Kennan,

Table 1. Rotation Poles

Age	Pole Latitude (°)	Pole Longitude (°)	Angle (°)
Yucatan to North America			
137	0.00	0.00	0.00
146	−22.48	96.94	20.04
153	−23.25	95.42	36.61
161	−24.42	96.14	38.92
195	−25.47	96.79	41.83
210	−26.49	97.18	43.09
South America to North America			
137	−31.63	−90.37	18.86
146	−36.32	−94.59	20.25
153	−39.14	−99.58	22.48
161	−40.44	−102.59	24.19
195	−45.03	−112.93	29.98
210	−46.05	−115.33	30.55

2001, 2007, 2009; Bird et al., 2005; Imbert and Philippe, 2005) and initially move Yucatan in tandem with South America. Yucatan moves with South America until it impinges on the western side of Florida and begins to spin counterclockwise about a pole near the present-day Straits of Florida. The primary constraints on plate vectors during this rotation are the Tamaulipas margin of Mexico, which is presumed to be a transform margin (Pindell, 1985), and possible ridge axes in the eastern Gulf of Mexico (Imbert, 2005; Imbert and Philippe, 2005). These possible ridge axes are consistent with the counterclockwise motion of Yucatan relative to North America and are on strike with rift structures in the Florida Straits, which were active during the rotation of Yucatan (Yang et al., 2011). Rotation poles for Yucatan and South America relative to North America are given in Table 1.

Following the general rift-to-rotation scheme for Yucatan outlined previously, plate positions were reconstructed in several stages. Stepping backward in time, the first constraint is extent of normal oceanic crust. This crust needs to be removed by moving Yucatan relative to North America via a route consistent with transform motion along the Tamaulipas margin and rifting in the Florida Straits. Because the limits of normal oceanic crust on the conjugate North American and Yucatan blocks do not make a simple mirror image (Figure 2), the oceanic crust must not have initiated synchronously alongstrike. We account for this lack of parallelism between margins by inferring that sea-floor spreading propagated in from both the east and west (Figure 13E). The next-oldest step is to restore the original edges of salt. Because the seaward boundaries of original salt deposition are unknown, this reconstruction is constrained by placing Yucatan in such a position so as to close the surrounding salt edges (Figure 13C). A further constraint is to allow Yucatan to move with respect to North America without moving toward South America while the Gulf of Mexico is opening. In all reconstructions, Yucatan is moved away from North America at a rate slower than or equal to the rate of motion of South America. The initial prerift fit is constrained by simply allowing Yucatan to fill the space between North and South America (Figure 13A).

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