Influence of deep Louann structure on the evolution of the northern Gulf of Mexico

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

ABSTRACT

Three aspects of basement structure and rift-related salt distribution have especially influenced the evolution of the deep-water northern Gulf of Mexico: (1) creation of a basement high (Toledo Bend flexure), separating a chain of interior basins from the central Louann salt basin, (2) segmentation of the central Louann salt basin by the Brazos transfer fault into eastern and central domains, and (3) salt provinces formed during basin opening.

The Toledo Bend flexure was reactivated as a hinge during the Cenozoic uplift of the North American craton. This uplift triggered gravity gliding, forming fold belts in the seaward parts of the continental margin. The geometry of the Toledo Bend flexure influenced the position of these fold belts.

The Brazos transfer fault separates the west sector of the study area from the central and east sectors. Most of the salt in the deep-water northern Gulf of Mexico lay in the central sector, which sourced most of the Sigsbee salt canopy. The western sector was narrower and was subdivided by the East Breaks basement high.

Splitting the Callovian salt basin in two as the gulf opened created a southward-thinning wedge of salt at the seaward end of the northern Gulf of Mexico. We divide this wedge into a series of provinces on the basis of the geometry of the base of the deep salt. Original salt thickness influenced diapir location, the geometry of the Sigsbee canopy, the geometry and style of later compressional fold belts, and petroleum systems.

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INTRODUCTION

For the past 40 yr, most regional syntheses and plate restorations of the Gulf of Mexico basin have agreed that the Callovian salt was originally deposited in a single broad basin, which eventually split in two as the gulf widened (e.g., Humphris, 1978; Buffler et al., 1980, 1981; Hall et al., 1982; Pindell and Dewey, 1982; Buffler and Sawyer, 1985; Pindell, 1985; Salvador, 1987, 1991; Winker and Buffler, 1988; Buffler, 1989; Sawyer, 1991; Buffler and Thomas, 1994; Marton and Buffler, 1994; Watkins et al., 1995; Dobson and Buffler, 1997; Pindell and Kennan, 2001). More recently, several articles have discussed the style of salt flow that accompanied basin widening and separation (e.g., Imbert and Philippe, 2005; Pindell and Kennan, 2007; Hudec et al., 2013). However, the systematic study of how this rift history influenced the subsequent salt-tectonic evolution of the basin has been minimal. Most palinspastic restorations in the Gulf of Mexico deep water (e.g., Peel et al., 1995; McBride, 1998; Trudgill et al., 1999; Hall, 2002; Rowan et al., 2004; Mount et al., 2007) assume that the salt was originally flat lying but make little attempt to relate the original salt geometry to basin formation.

Our goals are to describe the salt geometry inherited from rifting then discuss how this geometry influenced diapir location and style, the location of fold belts, the geometry of the Sigsbee salt canopy, and hydrocarbon prospectivity in primary basins. We focus on the deep-water northern Gulf of Mexico (Figure 1) because it has the most data and broadest interest for industry. We contend that rifting has exerted such a primary control on subsequent salt tectonics that the structural and stratigraphic evolution of the basin is impossible to understand without considering this inherited fabric.

SALT GEOMETRY INHERITED FROM JURASSIC RIFTING

A companion article (Hudec et al., 2013, this issue) proposes a new synthesis of the Mesozoic evolution of the Gulf of Mexico. This synthesis posits that salt deposition was followed by protracted crustal extension (Figure 2). During this extension, the salt stretched and thinned, allowing the top of salt to subside well below sea level. Sea-floor spreading then began, first in the west and then in the east. For unclear reasons (probably related to the isostatic properties of hyperextended transitional crust), normal oceanic crust in the Gulf of Mexico is now shallower than adjoining transitional crust so that the limit of normal

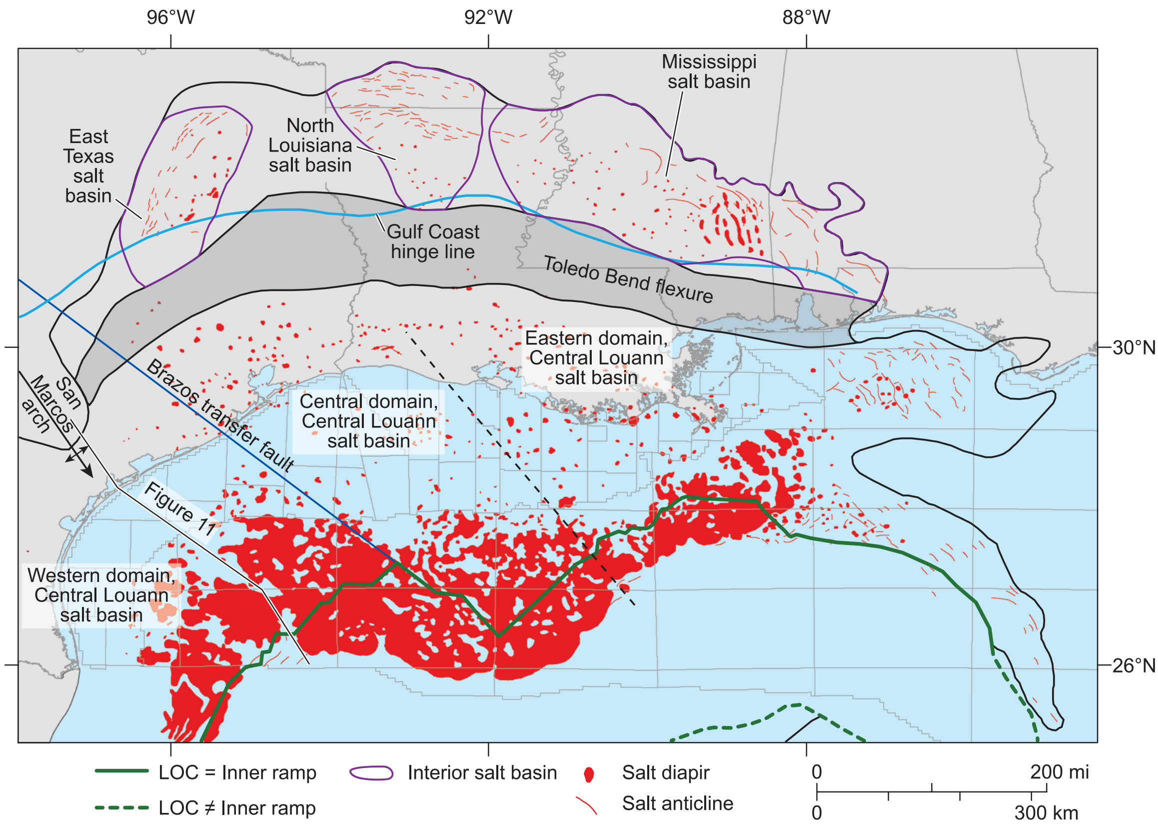

Figure 1. Key structures in the northern Gulf of Mexico and US Gulf Coast. The green line at the seaward end of the basin shows the limit of normal oceanic crust. In some places, this limit is marked by a major ramp in the base of the salt (the inner ramp), but in other places, it is not. Modified from Anderson (1979), Martin (1980), Simmons (1992), Diegel et al. (1995), Lopez (1995), and Jackson et al. (2011).

oceanic crust (LOC) is marked by a 1- to 4-km (0.6–2.5-mi)-high ramp up onto the oceanic crust.

In most of the northern Gulf of Mexico, seaward-flowing salt was able to flow up the ramp onto new oceanic crust (Figure 2C). Salt continued to flow seaward over the oceanic crust until it was buttressed against sediments aggrading in front of it. In places, this aggradation ended salt advance, and the salt was buried in place. Elsewhere, presumably as a result of greater salt supply, the salt inflated until it broke out over the top of the aggrading sediments and advanced as an allochthonous sheet even farther seaward. Eventually, this sheet was buried and stopped advancing.

Three aspects of this rift and salt-flow history especially influenced the subsequent evolution of the basin (Figures 1– 3): (1) creation of a basement high (Toledo Bend flexure) separating a chain of interior basins from the central Louann salt basin; (2) segmentation of the central Louann salt basin by the Brazos transfer fault into western and central domains; and (3) evolution of different salt provinces, controlled by the history of seaward salt advance.

Toledo Bend Flexure

Anderson (1979) defined the Toledo Bend flexure as a broad basement high separating the interior salt basins of the northern Gulf of Mexico basin from the central Louann salt basin (Figure 1). A near absence of salt structures suggests that salt is thin or absent across the flexure. The north boundary of the Toledo Bend flexure is drawn tangent to the southern edge of the interior salt basins. The southern boundary is based on the northern limit of salt diapirs in the central Louann salt basin and the northernmost Cenozoic growth faults.

Because it was a structural high during salt deposition, the Toledo Bend flexure is probably cored by thicker or less-extended crust. We argue subsequently

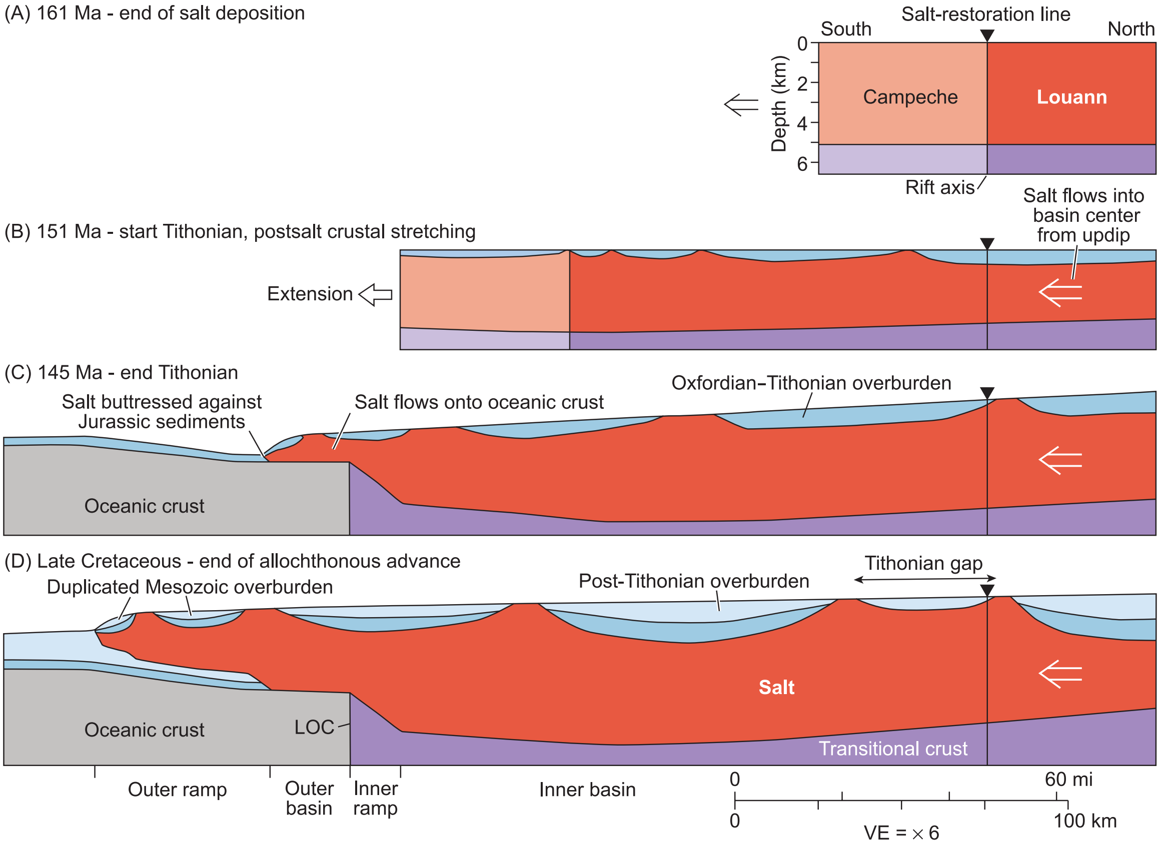

Figure 2. Schematic restoration of salt advance, north-central Gulf of Mexico. (A) End of Callovian salt deposition. (B) The basin widened by crustal extension. As salt stretched and flowed toward the basin center, salt was replenished by flow from updip. Extending overburden was rafted with the flowing salt. (C) Sea-floor spreading separated the Louann Salt from its equivalent in the Bay of Campeche. Salt in the north-central Gulf of Mexico was thick enough to climb the inner ramp and flow out onto the oceanic crust. (D) Eventually, Jurassic sediments on the oceanic crust thickened enough to impede the extruding salt, forcing salt to climb over abyssal-plain sediments as an allochthonous sheet. Allochthonous salt carried rafts of Upper Jurassic–Cretaceous overburden, which created extensional gaps farther updip. Restoration was constructed using the LithoTect software. LOC = limit of normal oceanic crust; VE = vertical exaggeration.

that its boundaries were zones of inherited crustal weakness, in that its northern margin was reactivated during Cenozoic craton uplift.

Brazos Transfer Fault

In the central part of the study area, the LOC forms a major southward promontory, which we term the "Walker Ridge salient" (Figure 3). The western margin of the Walker Ridge salient trends southeast, parallel to the movement of the Yucatán block during the early stages of plate separation. This parallelism led Hudec et al. (2013, their figure 13) to suggest that the sea floor spreads in the Gulf of Mexico in two segments, separated by a northwest-trending transform fault (see also Watkins et al., 1995; Barker and Mukherjee, 2011). Offset of the LOC across this transform fault formed the Walker Ridge salient.

The landward continuation of our suggested transform fault coincides with the Brazos transfer fault, a northwest-trending structure proposed by Simmons (1992). He suggested that the Brazos structure acted as a transfer fault during rifting, sepvarating blocks with opposite polarities of crustal extension (Simmons, 1992; Bradshaw and Watkins, 1995; Watkins et al., 1995; Huh et al., 1996). The Brazos transfer fault has been invoked in many other regional studies (e.g., Bradshaw and Watkins, 1995; Watkins et al., 1995; Huh et al., 1996; Karlo and Shoup, 1998; Stephens, 2001, 2009; Adams, 2009), although its existence and significance have been questioned (e.g., Buffler and Thomas, 1994; Pindell

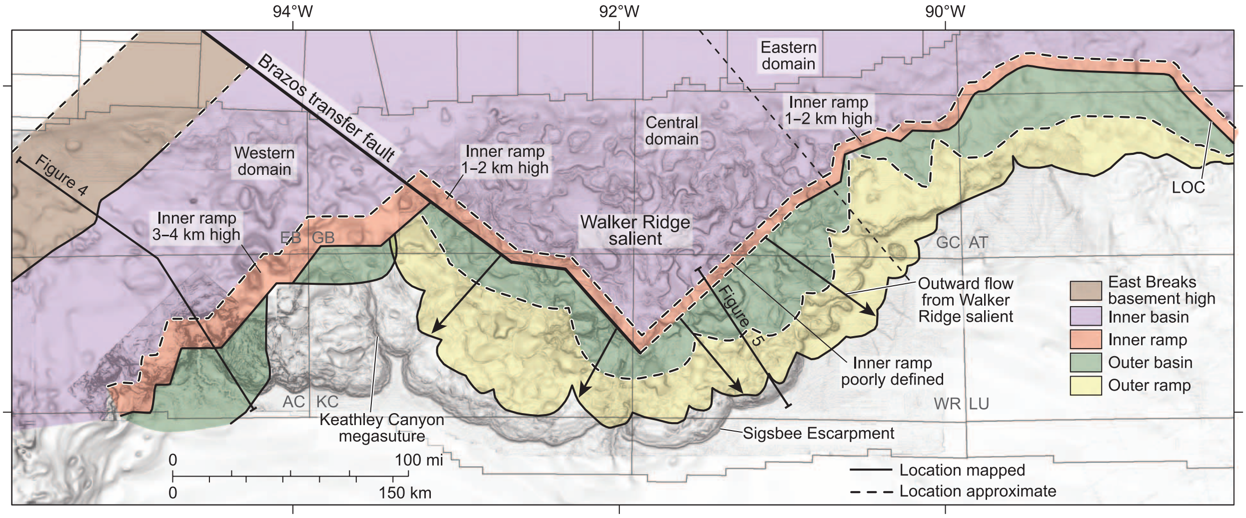

Figure 3. Map of deep-salt provinces in the study area. The limit of normal oceanic crust (LOC) is interpreted to lie at the boundary between the inner ramp and the outer basin. Parautochthonous salt in the outer basin and allochthonous salt in the outer ramp are interpreted to have flowed seaward across the LOC from the inner basin. Diverging arrows show inferred flow vectors of the deep salt. Approximate locations shown for Figures 4 and 5. Protraction area abbreviations on this and other maps: AC = Alaminos Canyon; AT = Atwater Valley; EB = East Breaks; GB = Garden Banks; GC = Green Canyon; KC = Keathley Canyon; LU = Lund; WR = Walker Ridge. The background image is a dip map on bathymetry, mostly from Bryant and Liu (2000).

and Kennan, 2001, 2007, 2009; Imbert and Philippe, 2005). We have drawn the Brazos transfer fault as a straight line in map view (Figures 1, 3). This depiction is an oversimplification, but we lack data to draw a more complex shape.

In addition to offset in the LOC, we identify two major changes across the Brazos transfer fault in the study area.

First, the crust below the inner basin is deeper west of the Brazos transfer fault than to the east (Figures 4, 5). In the western domain, the top of the acoustic basement (the base of deep salt) at the seaward end of the inner basin lies 16 to 18 km (10–11 mi) below sea level (e.g., Figure 4; Hudec et al., 2013, their figure 6B). East of the Brazos transform, the top of the acoustic basement at the seaward end of the inner basin is typically only 14 to 16 km (9–10 mi) deep (e.g., Figure 5; Hudec et al., 2013, their figure 6A). Why the crustal depths are different is unclear, but perhaps the transitional crust west of the Brazos transfer has different isostatic properties than the transitional crust to the east. Because of the difference in crustal elevation, the inner ramp in the central domain (east of the Brazos transform) has relief much lower than that in the west (Figures 3–5). To the east of the Brazos transform, relief on the inner ramp is 1 to 2 km (0.62–1 mi), as opposed to 3 to 4 km (1.8–2.4 mi) to the west.

Second, we propose the existence of a new structure, the East Breaks basement high, forming the updip limit of the inner basin west of the Brazos transfer fault (Figures 3, 4). The existence of this high is supported by three lines of evidence. First, the acoustic basement on the East Breaks basement high is 2 to 3 km (1–2 mi) higher than in the inner basin (Figure 4). Second, we have not mapped any diapirs sourced from the deep-salt layer atop the East Breaks high, although the basement high is overlain by a salt canopy sourced from diapirs to the north of the structure. Finally, Mesozoic strata above the basement high are relatively flat lying. All of these observations suggest that salt was originally thin on the East Breaks high. We are unaware of evidence of this basement high east of the Brazos transfer fault, so the Brazos transfer fault probably forms its northeastern boundary.

Both of these changes across the Brazos transfer fault suggest that the fault was active during Triassic–Jurassic rifting. Many major continental faults reactivate older trends, and it is certainly possible that the Brazos transfer had an earlier history. Similarly,

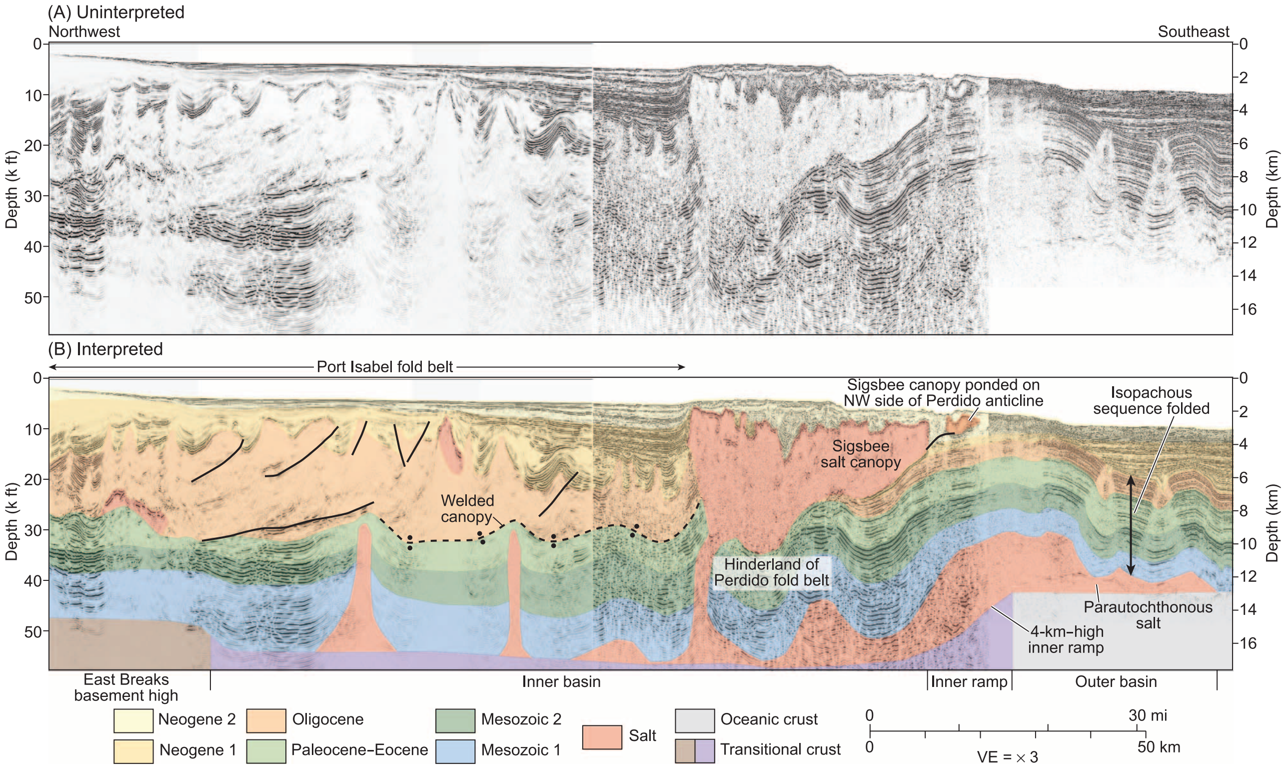

Figure 4. Regional seismic section across the western domain, showing deep-salt provinces. See Figure 3 for location. The top of the acoustic basement is not imaged in the inner ramp, but the overlying ramp in salt and sediment is well imaged. Mesozoic units in the inner basin lie several kilometers below their level on the outer basin, leading us to interpret a 4-km (2-mi)-high step in the basement at the inner ramp. The Sigsbee salt canopy appears to have ponded on the northwestern side of the largest Perdido anticline. Seismic data courtesy of Petroleum Geo-Services (PGS) and CGGVeritas. VE = vertical exaggeration.

the fault may have been reactivated after rifting, most notably during Cenozoic craton uplift (see below).

Salt Provinces

Synrift advance produced a seaward-thinning wedge of salt, with major thickness changes at ramps in the base-salt surface (Figure 2). These ramps divide deep-salt structure into four provinces (from landward to seaward): inner basin, inner ramp, outer basin, and outer ramp (Figures 2D, 3; Hudec et al., 2013, this issue). For convenience, we have interpreted the LOC at the boundary between the inner ramp and the outer basin, although this crustal transition is likely to be gradational.

Because province boundaries mark changes in original salt thickness, Figure 3 is similar to previous maps of subcanopy structural style (e.g., Hall, 2002; Fillon et al., 2005; Seitchik et al., 2007; Zarra, 2007; Kilsdonk et al., 2010; Pilcher et al., 2011). However, our map differs in that it is derived from the geometry of the base of deep salt, which is, in turn, controlled by rift architecture.

Two aspects of province geometry appear to have strongly influenced the evolution of the deepwater northern Gulf of Mexico.

First, the inner basin is much less extensive in the western domain than in the central domain (Figure 3). This smaller size has two inferred causes: (1) the LOC and inner ramp are offset roughly 175 km (109 mi) to the northwest on the western side of the Brazos transfer fault, and (2) the East Breaks basement high forms the northern boundary of the inner basin west of the transfer fault. Near-isopachous Mesozoic strata atop the East Breaks basement high suggest a negligible flow of salt southward across the high (Figure 4), so we infer that the volume of salt to make structures in the western Gulf of Mexico was less than to the east of the Brazos transfer fault.

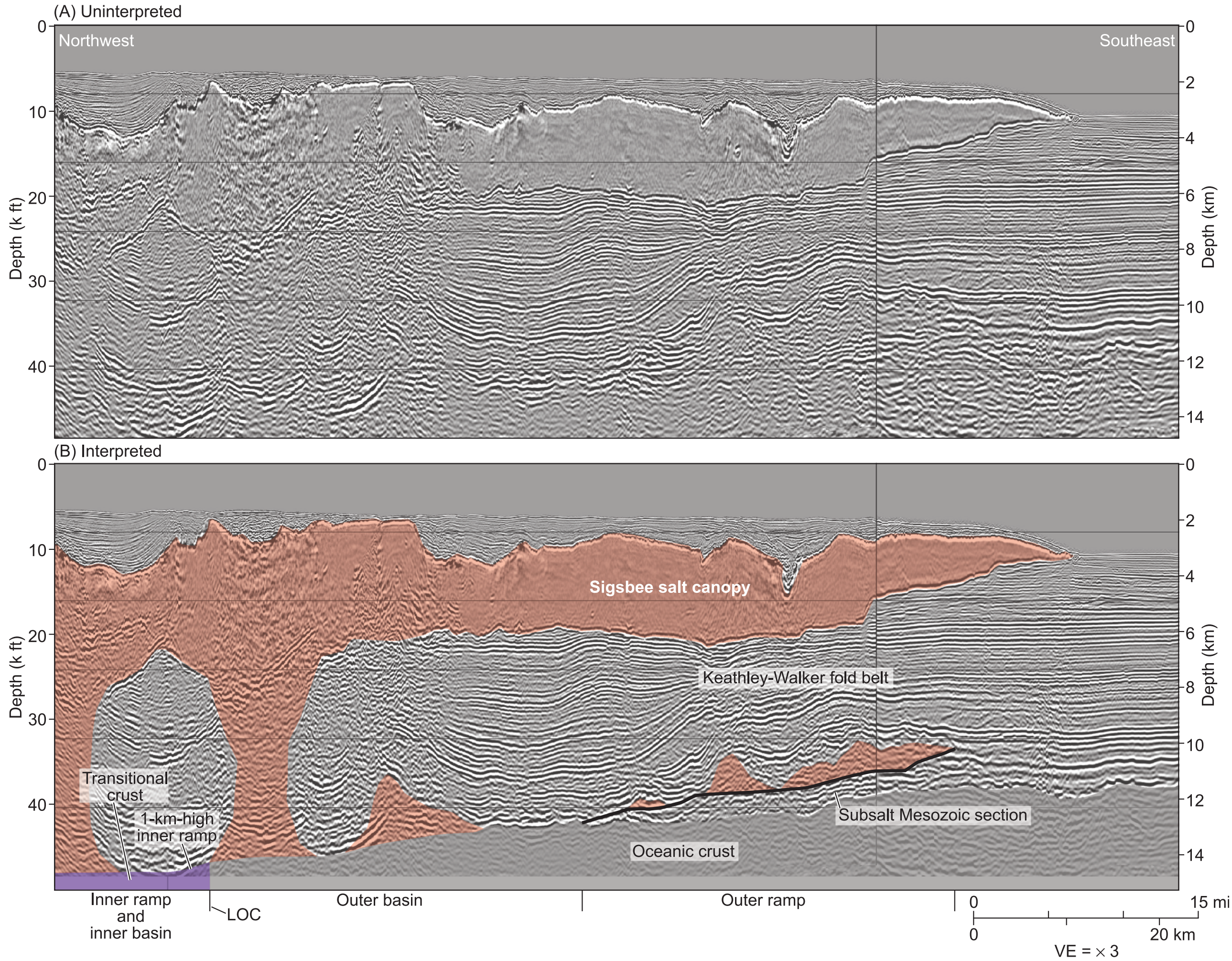

Figure 5. Regional seismic section across the central domain, showing deep-salt provinces. See Figure 3 for location. Basement dips gently northwest across the area, with no obvious inner ramp. The limit of normal oceanic crust (LOC) is drawn from its position on the regional map (Figure 3). The folds in the Keathley-Walker fold belt did not have enough relief to block advance of the Sigsbee salt canopy. Seismic data courtesy of CGGVeritas. VE = vertical exaggeration.

Second, the outer basin and the outer ramp are much more extensive around the Walker Ridge salient than farther west, indicating that salt flowed outward from the salient (arrows in Figure 3). By contrast, very little salt flowed onto the oceanic crust in the western Gulf of Mexico—the outer basin is discontinuous, and the outer ramp does not exist. We interpret this lateral variation in seaward salt flow to reflect (1) the lesser height of the inner ramp around the Walker Ridge salient, making the ramp a less effective barrier to seaward salt flow; and (2) the greater volume of salt in the inner basin updip of the Walker Ridge salient, which allowed it to surmount the low barrier easily.

INFLUENCE OF DEEP LOUANN STRUCTURE ON THE SUBSEQUENT EVOLUTION OF THE DEEP-WATER GULF OF MEXICO

We infer that basement structure and salt isochore patterns resulting from rifting and basin separation fundamentally controlled the style, size, and location of salt and sedimentary structures in the deep-water northern Gulf of Mexico. We illustrate this influence by discussing how salt distribution and basement structure governed the location of salt diapirs, the location and style of fold belts, the geometry of the Sigsbee salt canopy, and subcanopy hydrocarbon prospectivity and play types.

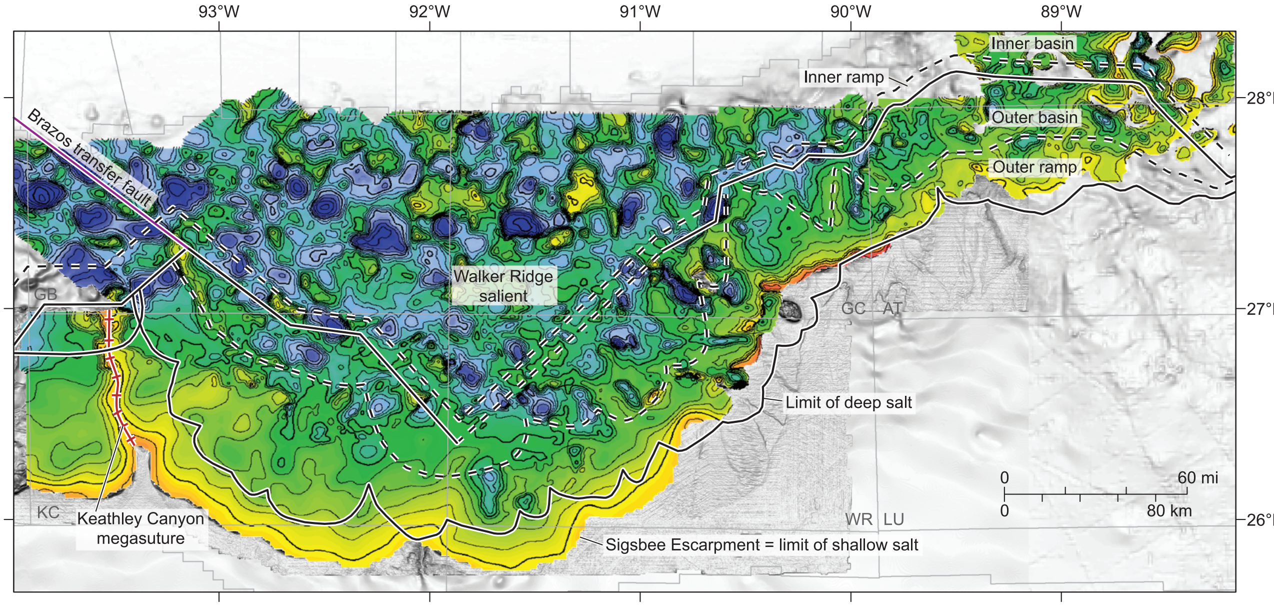

Figure 6. Deep-salt province boundaries superimposed on a contour map of the top of primary basins. Blue areas mark structural lows, which represent salt feeders. Most large feeders are in the inner basin because salt was thicker there. Additional large feeders in the outer basin on the eastern side of the Walker Ridge salient are attributed to an indistinct inner ramp that was less effective in blocking seaward salt flow. Contour map from Pilcher et al. (2011). AT = Atwater Valley; GC = Green Canyon; LU = Lund; WR = Walker Ridge.

Diapir Location

A map of the top of primary (subcanopy) basins by Pilcher et al. (2011) has depressions marking the geometry and location of diapirs feeding the overlying Sigsbee salt canopy. Superimposing their map onto our map of salt provinces (Figure 6) shows that feeders are more numerous and larger in the inner basin and inner ramp, as would be expected had salt been originally thicker there. In contrast, most salt structures on the outer basin are small diapirs, salt-cored folds, and relict pods of salt. Even less salt was available on the outer ramp, where most structures are salt-cored folds.

Fold-Belt Location and Style

Most major hydrocarbon discoveries in the deep-water Gulf of Mexico are related to the salt-cored folds near the seaward edge of the deep salt. These folds form three belts differing in timing, location, and structural style: the Atwater, Perdido, and Keathley-Walker fold belts (Figures 7, 8). Another diffuse, poorly defined fold belt that we name the "Timbalier fold belt" (after the South Timbalier protraction area that hosted the Blackbeard discovery well) lies farther updip, mostly beneath the present continental shelf. We suggest that these fold belts were strongly influenced by basement geometry and salt distribution inherited from rifting. Next, we briefly describe each fold belt and then infer how each was influenced by rift architecture.

Atwater Fold Belt

The Atwater fold belt (parts of which have also been called the "Mississippi Fan fold belt") extends 15 to 50 km (9–31 mi) updip from the seaward edge of the deep salt in the eastern domain (Figure 7). Publications discussing Atwater structures include those of Worrall and Snelson (1989), Weimer and Buffler (1989, 1992), Wu et al. (1990), Peel et al. (1995), Rowan (1997), Rowan et al. (2000, 2004), and Grando and McClay (2004).

Shortening in the Atwater fold belt deformed precursor diapirs, withdrawal basins, and low-relief salt anticlines (e.g., Weimer and Buffler, 1989, 1992; Wu et al., 1990; Rowan, 1997; Rowan et al., 2000, 2004; Hall, 2002). Preexisting diapirs have a major influence on fold belts, typically localizing shortening and nucleating folds (e.g., Vendeville, 2000; Grando and McClay, 2004; Letouzey and Sherkati, 2004; Rowan et al., 2004; Roca et al.,

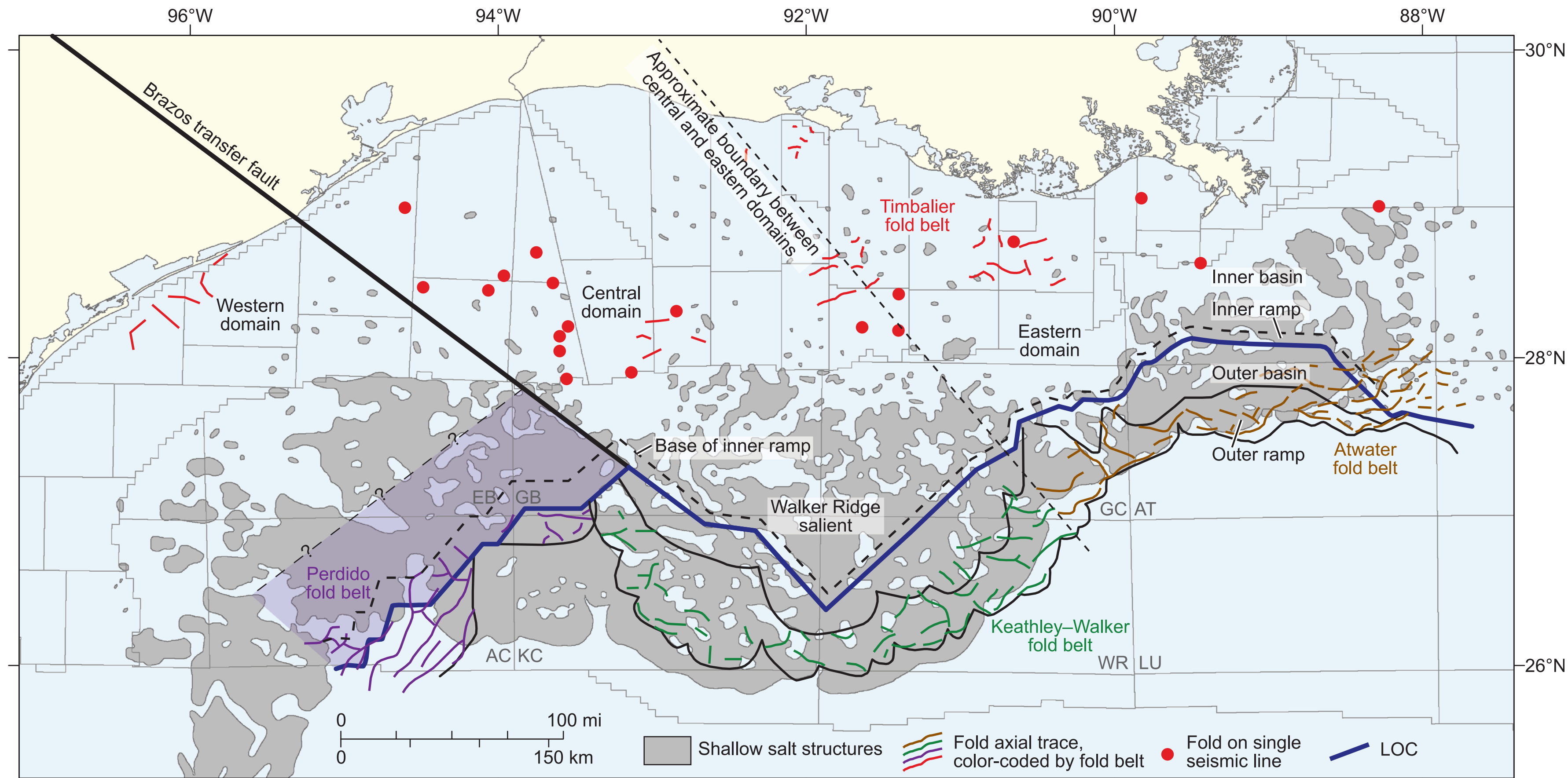

Figure 7. Map of fold belts and province boundaries in the northern Gulf of Mexico. Anticlinal axial traces modified from Krueger (2010) and Jackson et al. (2011). Salt outlines from Diegel et al. (1995). AC = Alaminos Canyon; AT = Atwater Valley; EB = East Breaks; GB = Garden Banks; GC = Green Canyon; KC = Keathley Canyon; LU = Lund; WR = Walker Ridge.

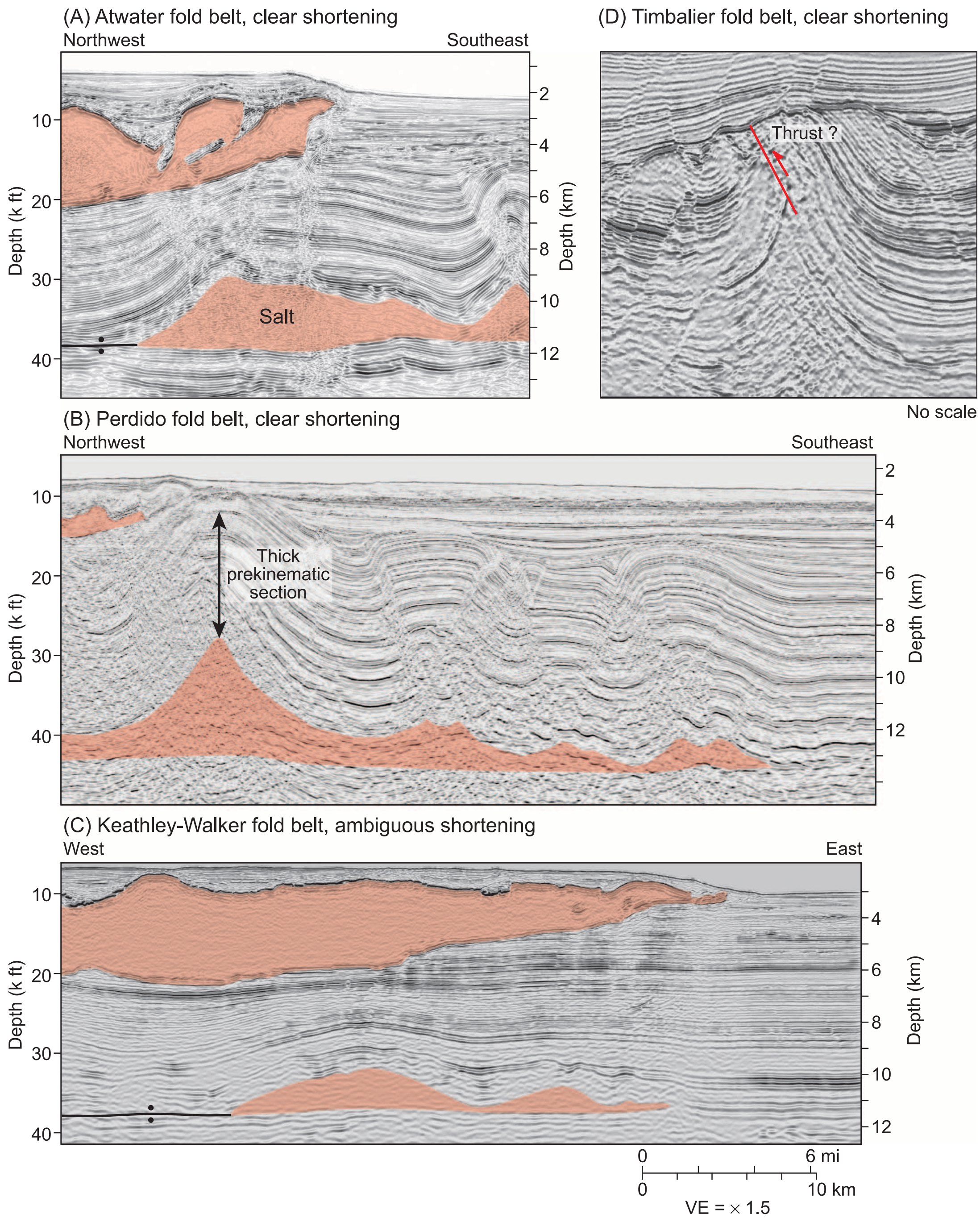

Figure 8. Seismic sections across the Gulf of Mexico fold belts. (A) Atwater fold belt. The authors thank BP, BHP Billiton, Chevron, and WesternGeco for access to and permission to show seismic data. (B) Perdido fold belt. Seismic data courtesy of CGGVeritas. (C) Keathley-Walker fold belt. More than 5 km (3 mi) of stratigraphy thins over the crest of these anticlines, suggesting a prolonged folding atypical of compressional deformation. The function of shortening in this folding is therefore unclear. Seismic data courtesy of CGGVeritas. (D) Timbalier fold belt. The compressional origin of this fold is indicated by crestal erosion and a probable crestal thrust fault. Seismic data reprinted from Veritel (2008). VE = vertical exaggeration.

2006; Rowan and Vendeville, 2006; Callot et al., 2007, 2012; Dooley et al., 2007, 2009). Mesozoic low-relief salt anticlines may also have influenced the structural style of superimposed Tertiary compressional structures (Rowan et al., 2000).

A variety of ages have been assigned to the Atwater fold belt. Most early articles propose a middle to late Miocene age for shortening (e.g., Wu et al., 1990; Feng and Buffler, 1991; Weimer and Buffler, 1992). Some later articles expand this range to late

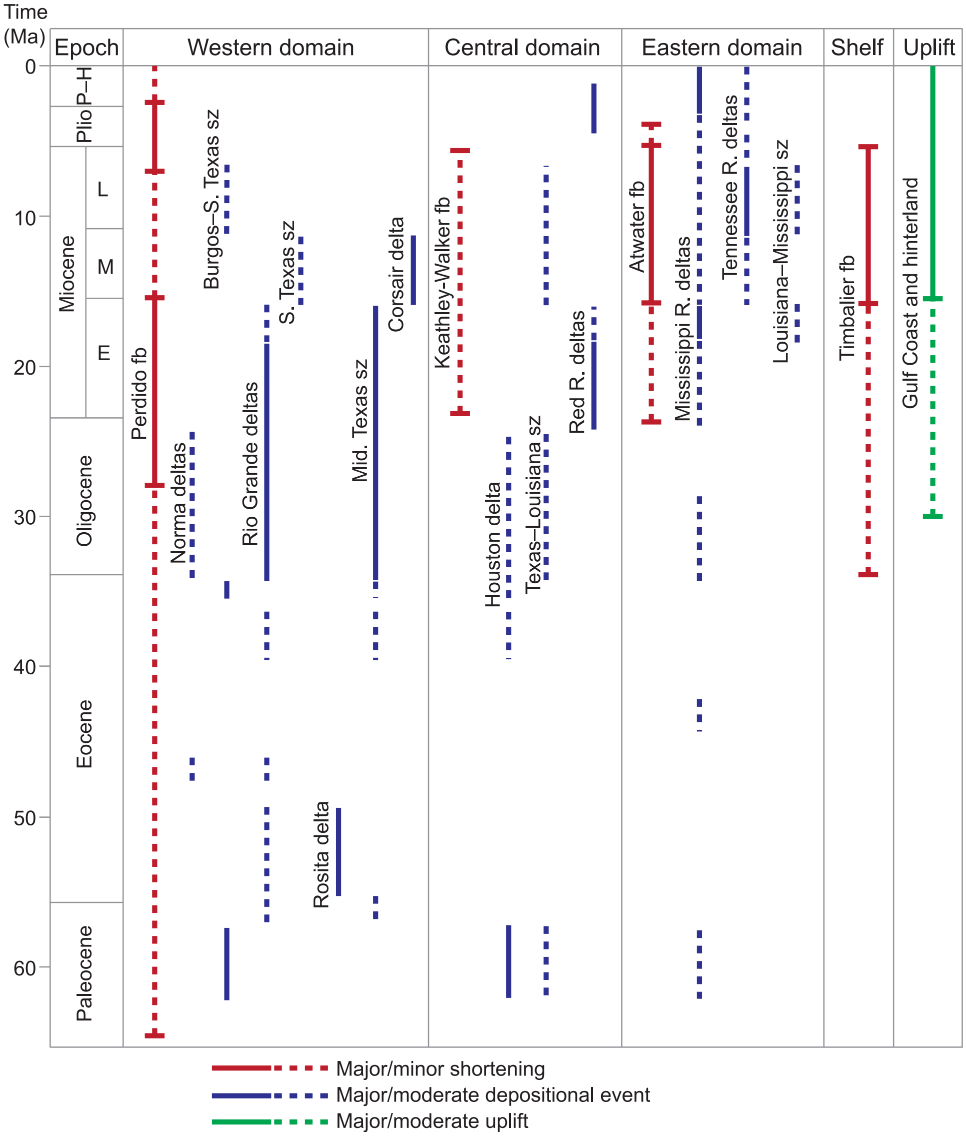

Figure 9. Timing chart for depositional events (blue), fold belts (red), and crustal uplift (green) in the northern Gulf of Mexico. Timing of depositional events from Galloway et al. (2000) and Galloway (2005). fb = fold belt; sz = shore zone; S. Texas = South Texas; Mid. Texas = Middle Texas; R. = River; Plio = Pliocene; P-H = Pleistocene-Holocene.

Oligocene to Pleistocene (Rowan, 1997; Rowan et al., 2004). Still others propose a late Miocene to early Pliocene age (Grando and McClay, 2004; Grando et al., 2009). Two main reasons account for these discrepancies. First, later studies use better seismic data and more well control. Second, folding typically started earlier landward and to the west, so the age of deformation varies with location. Recent seismic data showing lower Miocene reservoirs thinning across the crest of folds in the northwestern part of the Atwater fold belt indicate that shortening there began in the early Miocene (Mount et al., 2007). Regionally, shortening accelerated and spread during the middle to late Miocene then rapidly declined in the early Pliocene (Figure 9).

Perdido Fold Belt

The Perdido fold belt, in the western part of our study area (Figure 7), has also been the subject of numerous recent articles (e.g., Fuqua, 1990; Mount et al., 1990; Hall et al., 1993; Peel et al., 1995; Trudgill et al., 1995, 1999; Novoa et al., 1998; Rowan et al., 2000; Camerlo and Benson, 2006; Waller, 2007). Gradmann et al. (2009) provided an excellent summary of literature on the fold belt and discussed the factors influencing its development.

The Perdido fold belt has two parts. Most published maps (Trudgill et al., 1995, 1999; Fiduk et al., 1997, 1999; Rowan et al., 2000; Meyer et al., 2005; Camerlo and Benson, 2006; Waller, 2007) show only the frontal part of the fold belt above the outer basin, mostly outboard of the Sigsbee Escarpment (Figure 7). This part of the fold belt is limited by the lobate edge of the outer basin; beyond this edge, no salt to lubricate the basal detachment is observed. In this frontal part of the fold belt, no precursor structures were observed, as indicated by the isopachous, 4.5-km (2.8-mi)-thick folded Jurassic–lower Cenozoic units (Figures 4, 8B).

The Perdido hinterland is more poorly known. Despite the lack of published maps, most workers agree that most of the Perdido fold belt may lie in the inner basin, hidden beneath the Sigsbee canopy (Peel et al., 1995; Hall, 2002; Rowan et al., 2004; Mount et al., 2007; Radovich et al., 2007a, b; Kilsdonk et al., 2010; Figure 4). The lateral boundaries of the Perdido hinterland are not clear. The fold belt continues southward into Mexican waters and may extend eastward as far as the Brazos transfer fault (Figure 7). The hinterland of the fold belt formed above much thicker salt than did the distal end of the system, and numerous precursor diapirs, pillows, and withdrawal basins appear to have existed. We anticipate that these precursor structures will prove to have controlled folding, as in the Atwater fold belt.

Ages for shortening have been published for only the frontal part of the Perdido fold belt. The main phase of shortening here is late Oligocene to early Miocene (e.g., Trudgill et al., 1995, 1999; Fiduk et al., 1997; Winker, 2007), as indicated by the age of synkinematic strata (Figure 4). However, some folds nearer the land affect the sea floor and grew during a much wider time interval (Fiduk et al., 1999; Waller, 2007; Winker, 2007). Waller (2007) suggested that some folds started to grow in the Paleocene, peaked in the late Oligocene to the early Miocene, and then had a secondary peak in the late Miocene to the Pliocene (Figure 9). Given that the earliest initiated parts of the frontal Perdido fold belt are nearest the hinterland, the subcanopy parts of the Perdido fold belt may also have begun shortening during the Paleocene to the Eocene.

Keathley-Walker Fold Belt

An arc of folds rims the seaward end of the deep salt between the Atwater and Perdido fold belts (Figure 7; Seitchik et al., 2007; Krueger, 2010). Little has been published on this fold system, probably because it is almost entirely obscured by the Sigsbee salt canopy. We term this region of shortening the "Keathley-Walker fold belt," after the two protraction areas in which it occurs.

Despite its continuity in map view (Figure 7), the age and style of the anticlines in the Keathley-Walker fold belt vary. In the east, folds are clearly compressional and of Miocene age, leading Kilsdonk et al. (2010) to consider it a westward continuation of the Atwater fold belt. Farther southwest, in Walker Ridge, the folds are older and less clearly compressional (Figure 8C). Peel (2003) suggested a phase of Cretaceous–Paleogene shortening in this area, although their style suggests that some of these folds could also be halokinetic. Farther west in Keathley Canyon, unpublished data suggest Miocene shortening. We consider the only clear shortening in the Keathley-Walker fold belt to be Miocene in age, but this deformation is much milder and more diffuse than in the Atwater fold belt farther east (Figure 9).

Most axial traces in the Keathley Walker fold belt trend subparallel to the edge of the deep-salt layer, but many do not (Figure 7). Seitchik et al. (2007) interpreted some axial traces to form semicircles or closed polygons, suggesting that the shortening may have reactivated preexisting polygonal salt ridges formed by halokinesis. If so, preexisting structures created a structural template for the Keathley-Walker fold belt, just as in the Perdido and Atwater fold belts on either side.

Timbalier Fold Belt

Patchy information from beneath the salt canopy north of the deep-water fold belts indicates that a diffuse zone of compressional folding may extend as far north as the present coastline (Figure 7; Philippe et al., 2005; Guerin et al., 2006; Hunsdale, 2009; McDonnell et al., 2009; Hudec, 2010; Kilsdonk et al., 2010; Jackson et al., 2011; Dooley et al., 2013, this issue). Many of the structures in the current deep-shelf and ultradeep-shelf plays are four-way closures above these compressional folds. Most of the known Timbalier fold belt underlies the present continental shelf, although we cannot rule out the presence of subcanopy folds farther downdip. Folds

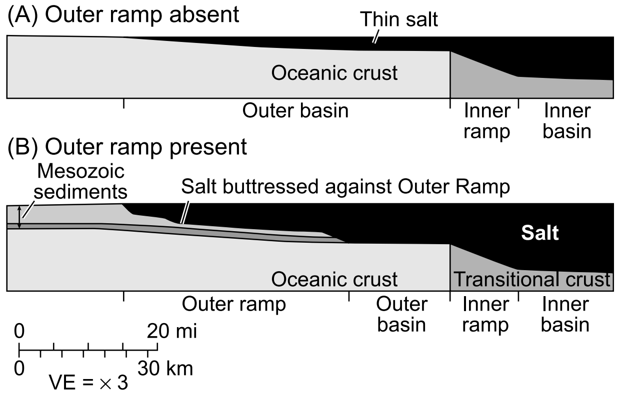

Figure 10. Thickness of deep salt as a function of basin geometry. (A) Where an outer ramp is absent, salt spreads across the outer basin in a relatively thin sheet, which is unlikely to form large salt structures. (B) Where an outer ramp is present, advancing salt ponds against it and thickens. This thicker salt can form more plentiful and larger salt structures. VE = vertical exaggeration.

belonging to the Timbalier fold belt have been observed across most of the width of the United States continental shelf and lie updip of the Perdido, Keathley-Walker, and Atwater fold belts (Figure 7).

Jackson et al. (2011) and Dooley et al. (2013, this issue) suggested that the Timbalier fold belt is a pillow fold belt, which they defined as a structural province containing mildly shortened salt pillows. Pillow fold belts are thus characterized by an early phase of halokinetic pillow growth, followed by a mild phase of shortening that amplifies the folds. The compressional phase is easily recognized where amplification is significant or where thrusts form, but it can be missed where shortening is subtle.

Consistent with this variability, some folds in the Timbalier fold belt have a clear compressional stage (Figure 8D), whereas others are more ambiguous. Where shortening is clear, it is not everywhere the same age. Philippe et al. (2005) and Guerin et al. (2006) interpreted a broad compressional domain active from the Mesozoic to the Miocene. McDonnell et al. (2009) reported a shortening of Oligocene age updip of the Perdido fold belt in Texas coastal waters. Jackson et al. (2011) reported a mid-Miocene shortening on the Louisiana shelf. Our own observations suggest that Miocene shortening is mostly widespread and involves the greatest shortening (Figure 9). Because shortening varies in time and space, however, the Timbalier fold belt is likely to be a composite of several deformational belts. We use the name purely as a convenience and look forward to further clarification as data improve.

Influence of Rift Structure on Fold Belts

The most obvious control of rift structure on fold belts is through salt distribution. Areas of thicker salt are more likely to have prefolding salt structures, which were critical in nucleating folds, controlling fold spacing, and influencing fold-belt structural style (Rowan et al., 2000, 2004; Jackson et al., 2011; Callot et al., 2012).

The thickest salt in the deep-water Gulf of Mexico accumulated where salt was buttressed by the inner or outer ramp during rifting (Figure 10). Thus, the Atwater, Timbalier, and Keathley-Walker fold belts and the hinterland of the Perdido fold belt all formed above thick salt containing abundant precursor structures. Salt was much thinner in the frontal Perdido fold belt, where no outer ramp to dam the salt existed. Much salt flowed southward from the inner basin into frontal Perdido fold cores during the Oligocene–Miocene shortening, but structural restorations suggest that the salt there was thin before shortening began (e.g., Trudgill et al., 1999; Mount et al., 2007). Initially, thin salt in the frontal Perdido fold belt explains the absence of precursor salt structures and, thus, the isopachous preshortening stratigraphy.

We contend that fold-belt timing and growth were also influenced by the Cenozoic craton uplift that reactivated old rift trends. This uplift is clearly indicated by the 60-m.y. span of offlapping outcrops along the US Gulf Coast and by seismic data showing large-scale erosional truncation at the present topographic surface (Dooley et al., 2013, this issue). Together, these suggest several kilometers of uplift of the continental interior. Jackson et al. (2011) provided stratigraphic evidence that cratonic uplift began in the early Oligocene (~30 Ma), accelerated in the middle Miocene (~15 Ma), and is ongoing (Figure 9).

Dooley et al. (2013, this issue) documented two hinge lines in the uplift of this part of the North American craton, both of which reactivate rift structures. First, decreasing outcrop widths suggest that uplift steepens near the updip depositional limit of Louann salt, near the edge of full-thickness

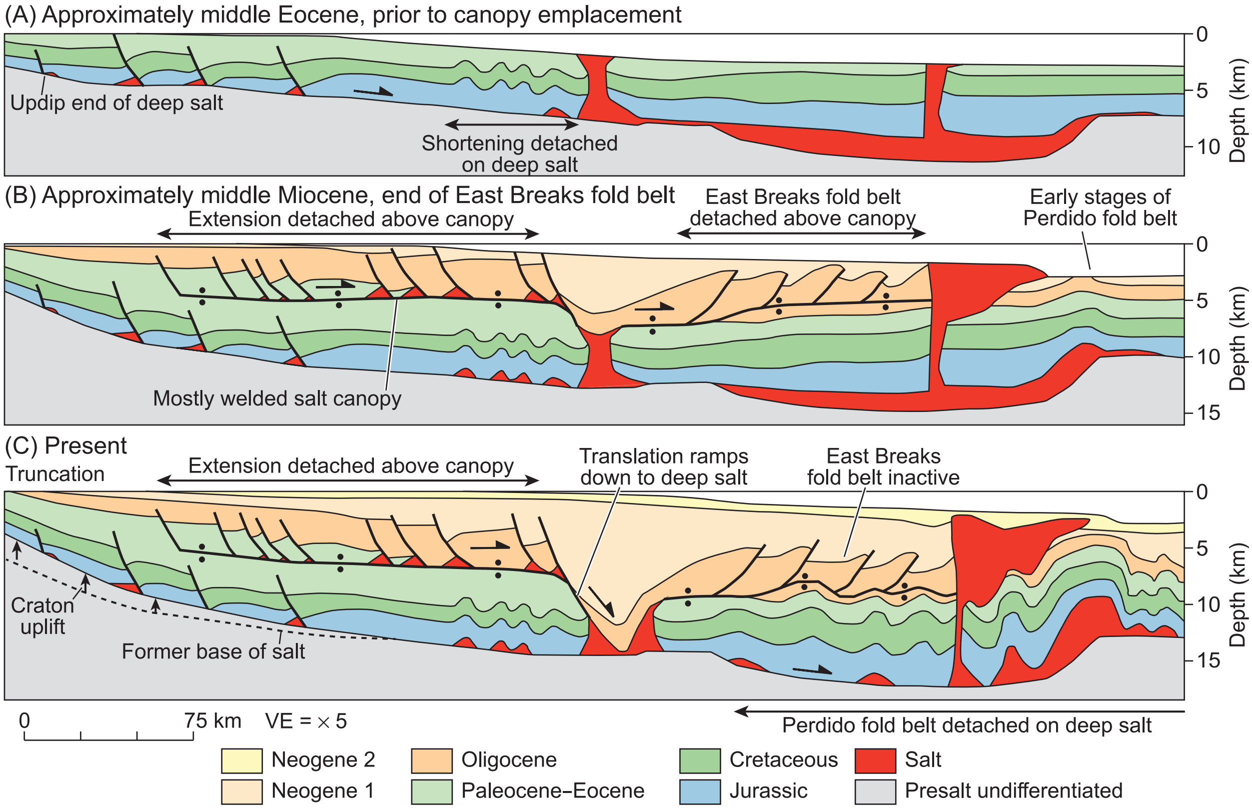

Figure 11. Schematic restoration of a simplified geologic cross section across the western Gulf of Mexico, illustrating fold-belt evolution. Approximate line location shown in Figure 1. (A) Initially, extension and shortening were detached on the deep-salt layer. (B) Following the emplacement of a continuous canopy in the Eocene, translation was detached on allochthonous salt, forming the East Breaks fold belt near the downdip end of the welded canopy. (C) Oligocene–Holocene craton uplift has tilted the margin, reactivating parts of the deep-salt detachment and shifting shortening to the Perdido fold belt at the seaward end of the margin. Geology on the cross section is simplified from Figure 4 (downdip half) and from an unpublished ION–GeoVentures GulfSPAN Merge interpretation report (courtesy of Adrian McGrail, ION Geophysical). VE = vertical exaggeration.

continental crust (Sawyer et al., 1991; Galloway, 2008). Jackson et al. (2011) defined a second prominent hinge (Gulf Coast hinge line; Figure 1) near the landward edge of the Toledo Bend flexure. The location of these hinge lines suggests that uplift-exploited zones of crust weakened during basin formation.

Given the existence of this uplift and seaward tilting of the northern Gulf of Mexico margin, how were fold belts affected? Most publications attribute the Perdido and Atwater fold belts to gravity spreading (e.g., Wu et al., 1990; Peel et al., 1995; Rowan et al., 2000; Gradmann et al., 2009). In this model, depocenters on the continental margin failed extensionally updip and shortened downdip as a fold belt. Gravity spreading explains the timing and location of the Atwater and Keathley-Walker fold belts, in that they began shortening in the Miocene as major updip depocenters were initiated (Figure 9; Winker, 1982; Galloway et al., 1991, 2000; Galloway, 2001, 2005, 2008). However, the link between gravity spreading and shortening is more complex in the Perdido and Timbalier fold belts.

Large-scale clastic deposition began updip of the Perdido fold belt during the Paleocene and peaked during the early Oligocene (Figure 9; Galloway et al., 2000). However, the main phase of Perdido shortening did not begin until the late Oligocene, during a systems-tract retreat (Galloway et al., 2000). Peak shortening in the Perdido fold belt was thus out of phase with depositional loading. Why was Perdido shortening delayed, and where was shortening accommodated during gravity spreading prior to the initiation of the Perdido fold belt?

Most workers agree that Cretaceous–Eocene gravity spreading was accommodated by Louann-detached shortening somewhere updip of the present frontal Perdido fold belt (Figure 11A; Peel et al., 1995; Trudgill et al., 1999; Radovich et al., 2007a; Gradmann et al., 2009). This inferred that early fold belt is obscured by the overlying salt canopy. Emplacement of a 100-km (62-mi)-long canopy in the late Eocene provided a higher-level detachment for gravity spreading. As a result, most Oligocene–early Miocene shortening was in the shallower Port Isabel–East Breaks fold belt at the seaward end of the canopy (Figure 11B; Rowan et al., 2004, 2005; Mount et al., 2007; Radovich et al., 2007a). The Perdido fold belt in the late Oligocene to the early Miocene therefore entailed an abrupt shift of at least some of the slip back down to the deep-salt detachment and an equally abrupt shift all the way to the seaward end of the salt layer. Rowan et al. (2005) argued that displacement descended onto the autochthonous detachment because sediment thickened above the Port Isabel fold belt, stiffening the fold belt until it could no longer shorten. In contrast, Gradmann et al. (2009) suggested that resumption of slip on the deep-salt detachment was a natural result of margin progradation, which reactivated the deeper detachment regardless of the presence of a shallower detachment.

We propose a third hypothesis: slip resumed on the autochthonous detachment in the late Oligocene because the uplift of the North American craton tilted the whole margin seaward (Figure 11C). Consistent with our inference of tilting, continental uplift began at the same time as shortening in the frontal Perdido fold belt (Figure 9). Furthermore, tilting displaced shortening away from the uplift to the seaward end of the deep salt. We agree that sedimentary loading, margin progradation, and burial of the Port Isabel fold belt may all have contributed to the transfer of translation back onto the deep-salt detachment but argue that the timing and location of the Perdido fold belt both suggest that margin tilting was the main trigger.

As noted, the age of shortening in the Timbalier fold belt is poorly known and variable. Scanty evidence suggests that most of the shortening is Oligocene to Miocene, an age consistent with a gravity-spreading hypothesis. However, the shortening is anomalously far updip, near the paleoshelf edge. Furthermore, structures in the central domain formed beneath a mostly continuous Paleogene salt canopy, above which most subsequent slope failure was detached (Peel et al., 1995; Rowan et al., 2009; Rowan, 2011). Why then did slip shift back down to the autochthonous level?

Jackson et al. (2011) and Dooley et al. (2013, this issue) presented physical and conceptual models of a two-level detachment system on a tilted passive margin. Their results suggest that tilting causes slip simultaneously on both detachments and that shortening above the deep-salt detachment can occur far updip of the toe of the slope. They therefore inferred that the Timbalier fold belt was triggered by craton uplift, as argued earlier for the Perdido fold belt. Jackson et al. (2011) suggested that the diffuse nature of the Timbalier shortening in the central and eastern domains was caused by the greater width of the salt basin there. Continental uplift was much closer in the Perdido fold belt because (1) basement uplifts are much farther south on the western side of the Brazos transfer fault (as shown by wrapping of the updip pinch-out of salt around the San Marcos arch in Figure 1) and (2) the LOC jumps 175 km (109 mi) to the northwest on the western side of the Brazos transfer fault.

We therefore infer that the Perdido and Timbalier fold belts were caused by both gravity spreading and gravity gliding. The presence of major depocenters updip destabilized the margin and began gravitational failure, typically detached above shallow-salt canopies. As cratonic uplift tilted the passive margin, some of the seaward translation shifted down to the deep-salt detachment, forming the fold belts. The age of shortening in the Atwater and Keathley-Walker fold belts leaves open the possibility that they were caused by gravity spreading alone, but because the margin was also tilting then, gravity gliding probably was an important factor (Figure 9).

Thus, we argue that rift architecture governed the location, timing, and structural style of fold belts in the deep-water Gulf of Mexico. Crustal boundaries during rifting were reactivated during the Cenozoic uplift of the North American craton, which influenced fold-belt location and timing. As the severed

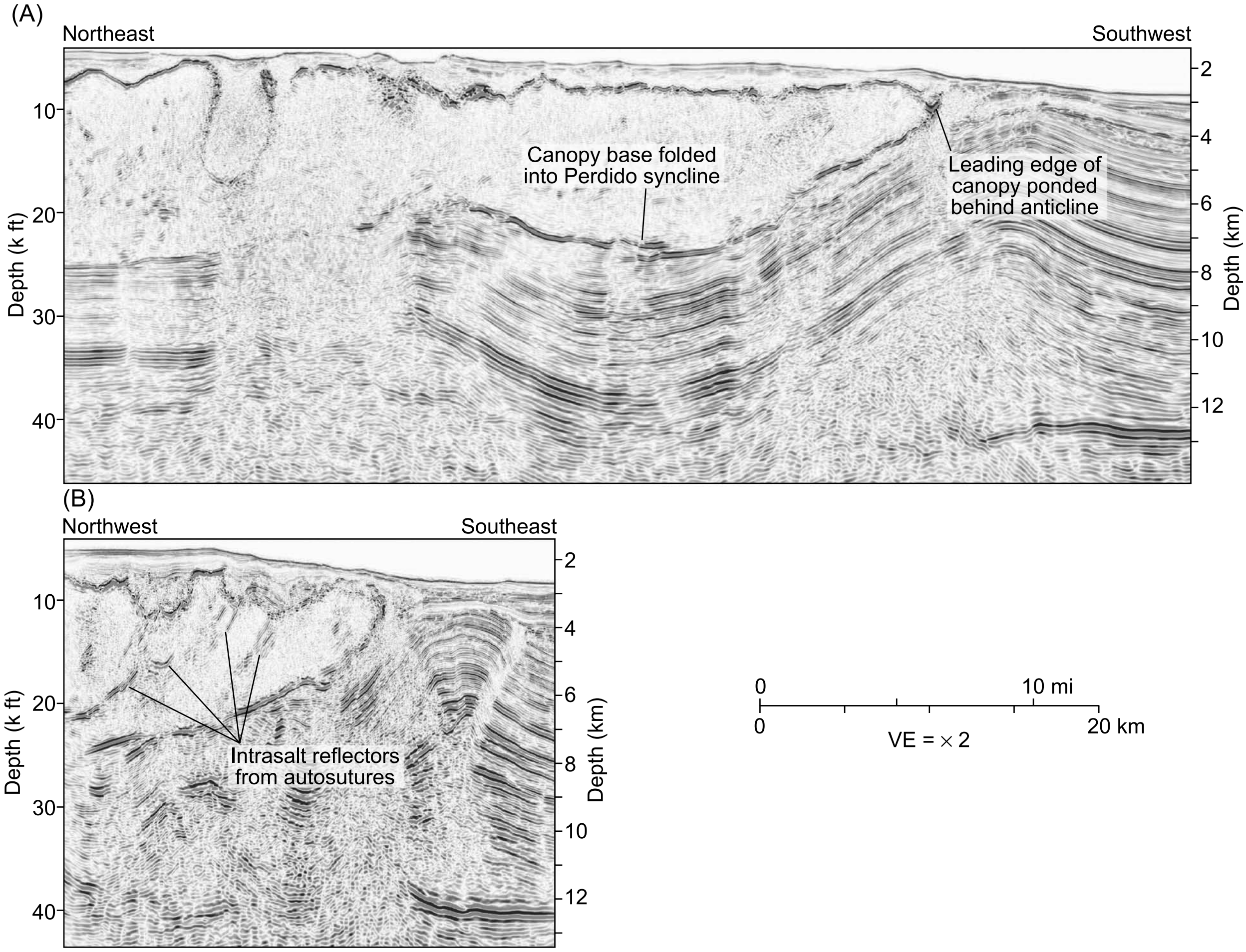

Figure 12. Seismic sections showing the interaction between the Sigsbee salt canopy and the Perdido fold belt. Sections are at the same scale. (A) Canopy deformed by infilling Perdido folds. (B) Canopy ponded behind the Perdido anticline. The rising anticline has blocked advance of the canopy, causing the canopy to thicken, shorten, and ruck up to produce autosutures. Seismic data courtesy of CGGVeritas. VE = vertical exaggeration.

salt basins separated, seaward flow of salt redistributed the salt, controlling where preshortening salt structures could form. These precursor structures, in turn, influenced fold-belt structural style.

Geometry of the Sigsbee Salt Canopy

Seaward salt flow during rifting left the thickest salt in the inner basin. As a result, most of the salt currently in the Sigsbee canopy rose up feeders in the inner basin and then flowed south for as much as 100 km (62 mi) to its present location (Figure 6; Hall, 2002; Pilcher et al., 2011). Two main source areas for the Sigsbee canopy are separated by the Brazos transfer fault: the Walker Ridge salient (central domain) and the area updip of the frontal Perdido fold belt (western domain).

Allochthonous salt from these two source areas merged along the present Keathley Canyon to form the Keathley Canyon megasuture (Figures 3, 6). The Keathley Canyon megasuture does not lie midway between the two canopy source areas. Instead, it is nearly 100 km (62 mi) west of the midpoint, implying that the canopy derived from the central domain spread much farther than the canopy derived from the western domain. We suggest two reasons for this differential advance, both related to rift architecture.

First, the map extent of the Sigsbee canopy derived from the two source areas is similar to that of the deep salt in the outer basin and the outer ramp (Figure 3), that is, deep salt derived from the central domain also advanced much farther out onto the oceanic crust than did salt from the western domain.

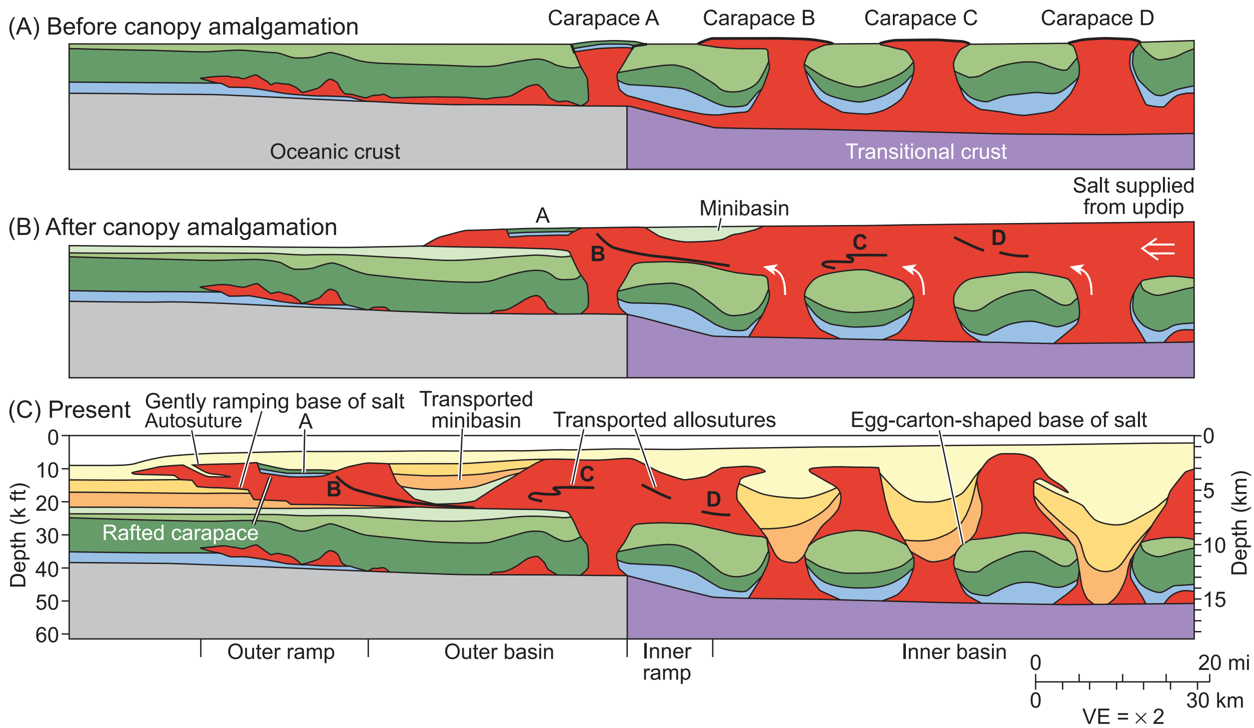

Figure 13. Schematic restoration showing the inferred evolution of Sigsbee salt canopy in the central domain. Most salt was supplied to the canopy via feeders in the inner basin then flowed seaward to its current position. This lateral allochthonous flow transported sutures, carapace blocks, and minibasins. The diverse fates of the four carapaces are merely illustrative and do not imply any spatial relationships. Restoration constructed using the LithoTect software. VE = vertical exaggeration.

This greater advance implies that much more salt was available in the central domain to feed advance of both the deep salt and the Sigsbee canopy. Abundant salt in the Walker Ridge salient most likely resulted from more downslope flow of salt from the landward parts of the basin. Because the central Louann salt basin is widest updip of the Walker Ridge salient, more salt was available there to flow downdip and feed the Sigsbee salt canopy. Furthermore, even the lesser volume of salt in the updip parts of the western domain would have been blocked from flowing downdip by the East Breaks basement high.

Second, rising folds in the Perdido fold belt appear to have hindered advance of the Sigsbee canopy in the western domain (Gradmann et al., 2009; Kilsdonk et al., 2010). Thus, the Sigsbee Escarpment lies just landward of the crest of major Perdido anticlines over the large parts of the area, and the base of the Sigsbee canopy is deformed by Perdido folds (Figures 4, 12A). The rise of the Perdido fold belt therefore buttressed the overlying Sigsbee canopy at the seaward end of the deep salt, which we tie to the proximity of rift structures reactivated by crustal uplift in the northwestern Gulf of Mexico. No folds of this scale existed in the central domain, so the Sigsbee canopy was free to flow southward unimpeded (Figure 5).

Differential advance of the Sigsbee salt canopy in the north-central and northwestern Gulf of Mexico has important implications for the internal structure of the canopy in both areas.

In the western domain, ponding of salt on the updip side of Perdido anticlines abruptly thickened salt in the Sigsbee canopy updip of the fold belt. In contrast, the canopy above the Walker Ridge salient tends to taper more gently (cf. Figures 4, 5). Buttressing of canopy advance by the rising Perdido folds created numerous autosutures, which incorporated roof sediments into the canopy (Figure 12B; Dooley et al., 2012). These sedimentary inclusions complicate the interpretation of the canopy base and may be one reason for the poor subsalt imaging in the area (see Hudec et al., 2013, their figure 3, this issue, for the nomenclature used to state position with respect to salt).

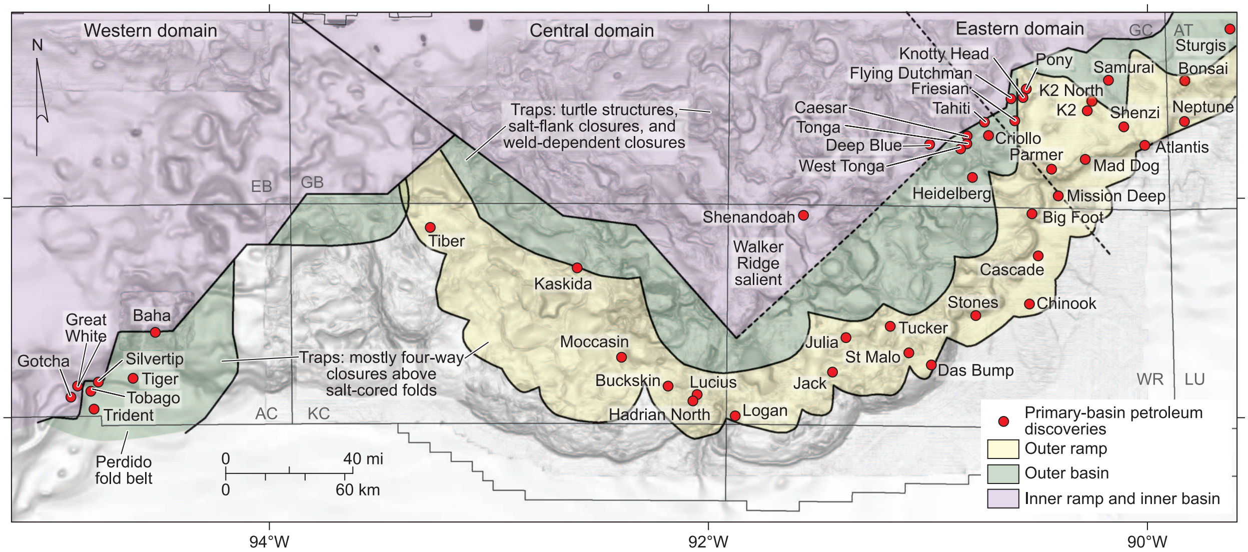

Figure 14. Map showing the discoveries beneath and in front of the Sigsbee canopy in the study area. Discovery locations courtesy of BHP Billiton.

Around the Walker Ridge salient, large-scale advance of the Sigsbee canopy from areas of thicker salt created structures very different from those farther west (Figure 13). This advance history influenced canopy structure in three main ways.

First, most of the Sigsbee canopy in the inner basin and the inner ramp is locally sourced, rising up feeders directly beneath the canopy. Salt extruding from these feeders climbed upsection and sutured with salt from nearby feeders, creating a salt-stock canopy having an irregular base like an egg carton (Kilsdonk et al., 2010). We agree with Pilcher et al. (2011) that the complex and steeply dipping base of salt produced by coalescing salt stocks is a major reason that subsalt seismic imaging is so poor in the inner basin and the inner ramp. In contrast, most of the salt within the Sigsbee canopy above the outer basin and the outer ramp is sourced from feeders located tens of kilometers to more than 100 km (62 mi) updip. This salt climbed gradually upsection as it flowed southward, producing a gently ramped base of salt. This comparatively simple base-salt geometry is the primary reason that subsalt seismic imaging is relatively good in the outer basin and the outer ramp.

Second, the southward flow of canopy salt would have carried any minibasins or carapace blocks many kilometers south of their original locations (Figure 13). We suggest that most, if not all, minibasins on the outer fringe of the Sigsbee canopy were transported seaward. Many of these minibasins include carapace blocks containing anomalously old (Paleogene to Cretaceous) strata that were originally deposited above feeder diapirs before being rafted off (e.g., McGuinness and Hossack, 1993; Hart et al., 2004; Mount et al., 2007, Pilcher et al., 2011).

Third, the southward flow of salt within the Sigsbee canopy would have created and transported a large number of sutures. Some were allosutures, formed when salt advancing from northern feeders overrode less-vigorous diapirs and smaller salt sheets sourced from thinner salt farther south. Others were autosutures, formed when landward parts of the canopy overrode pinned frontal parts of the same sheet. In either type, suturing would have trapped fragments of roof within the canopy. These fragments were then transported southward within the spreading canopy to accumulate as flotsam near the southern margin of the canopy (Figure 13).

Subcanopy Hydrocarbon Prospectivity

Because province architecture is a primary control on original salt thickness (Figure 10), it should come as no surprise that province geometry influences many aspects of petroleum geology in the deepwater northern Gulf of Mexico. We identify four

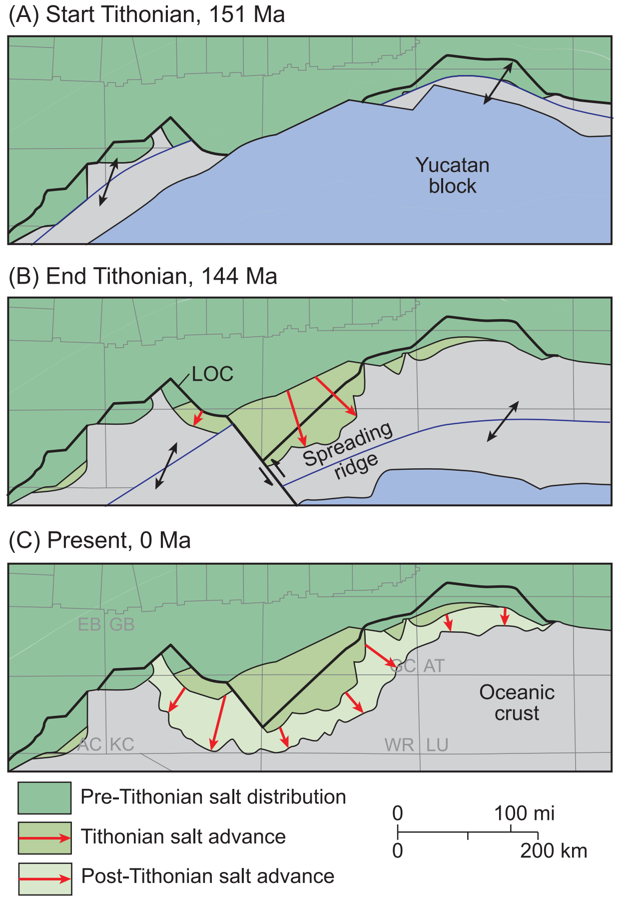

Figure 15. Plate restorations showing Tithonian and post-Tithonian advance of the deep-salt layer in the study area. Advancing salt carried Tithonian source rocks, creating an extensional gap of Tithonian units somewhere updip. Extension during Tithonian salt advance (A-B) thinned the Tithonian sequence, in which lowermost units were extended more than the upper ones. Post-Tithonian advance (B-C) further stretched the Tithonian interval, producing extensional gaps whose shapes and locations are as yet unknown. Restorations constructed by Ian Norton using PaleoGIS software. See Hudec et al. (2013) for details of plate reconstruction. LOC = limit of normal oceanic crust.

aspects of petroleum geology in which this influence is especially pronounced: trap style, reservoir deposition, source rock distribution, and field distribution (Figures 14, 15).

First, salt thickness exerts a primary control on salt structural style and, thus, trap style. Areas of thick original salt formed large salt stocks and salt walls (Figure 6), whereas thinner salt farther outboard tended to form salt-cored anticlines (Figure 8). The outermost salt province, where salt was thinnest (outer ramp in the north-central Gulf of Mexico, outer basin in the northwest), is thus characterized by four-way closures above salt anticlines. In the thicker-salt provinces farther landward, turtle structures, salt-flank closures, and weld-dependent traps dominate (e.g., Pilcher et al., 2011).

Second, variations in salt thickness across the study area also influenced reservoir deposition. Near the thin seaward edge of the deep salt, drilled Paleogene reservoirs have proven remarkably continuous over long lateral distances (e.g., Zarra, 2007). This lateral continuity lessens landward, where the underlying salt layer was thicker and minibasins dominate (Pilcher et al., 2011). We therefore expect reservoir risk in the Paleogene section to increase northward in the deep-water Gulf of Mexico. This risk may be less severe in the Miocene section because some deep salt may have welded before the Miocene.

Third, a key implication of our hypothesis for the evolution of the deep salt is that Turonian source rocks were transported along with seaward-flowing salt. This transport duplicated Turonian units in the outer ramp as allochthonous source rocks were rafted by the advancing canopy above autochthonous equivalents overlying the oceanic crust (Figure 2). Allochthonous Turonian source rocks carried by the advancing salt will be less mature than their deeper autochthonous equivalents and most likely to laterally vary in thickness because of their allochthonous setting above unstable salt. The source rocks may even have been deposited far enough away to have a geochemical signature different from that of autochthonous source rocks. Another consequence of rafting is that the duplicated source rocks must be kinematically balanced by an extensional Turonian gap located somewhere updip (Figure 2). Turonian source rocks may be extensionally thinned or completely absent in this gap. We expect the Turonian gap to be widest updip of the Walker Ridge salient, where post-Turonian salt advance was greatest. Our hypothesis suggests as much as 100 km (62 mi) of salt advance during the Tithonian, followed by an additional 30 to 60 km (19–37 mi) of advance (Figure 15). If so, the Tithonian source rocks could be thin or absent beneath parts of the modern shelf and upper slope in the north-central Gulf of Mexico.

Finally, a map of deep-water Gulf of Mexico discoveries in primary basins (where the reservoirs lie below the Sigsbee canopy or outboard of it) shows a strong correlation to province and domain boundaries (Figure 14). We discuss two aspects of this correlation. First, around the Walker Ridge salient and in the Perdido fold belt, discoveries are concentrated in the outermost salt province (outer ramp and outer basin, respectively), with only two discoveries in more northern provinces. We attribute this cluster of discoveries to the simple trap geometries and good seismic imaging characteristic of the outermost parts of the deep-salt layer. Because most of the simple four-way dip closures in these areas have now been drilled, subcanopy exploration is beginning to move into more complex structural domains to the north. Success in this work should extend the belt of discoveries northward in the future. Second, subcanopy discoveries extend much farther north in the eastern domain than in the central domain. Part of this northerly extension is caused by the presence of Atwater fold-belt discoveries in the eastern domain, but not all of the discoveries there are in the fold belt. Another observation is that subcanopy seismic imaging improves to the northeast in this area, offering an alternative explanation for greater exploration success. More work is needed to explain the pattern of discoveries in the northeastern part of the study area.

SUMMARY AND CONCLUSIONS

Cretaceous and Cenozoic evolution of the deep-water northern Gulf of Mexico cannot be understood apart from its Jurassic rift history. The Brazos transfer fault divided the central Louann salt basin into unequal halves having much less salt in the west. The Toledo Bend flexure isolated the chain of interior basins from the main salt basin during Jurassic salt deposition but was reactivated as a hinge during Cenozoic craton uplift. Salt flow during basin separation formed a seaward-thinning wedge of salt. This salt distribution controlled the subsequent evolution of the study area in the following respects.

1. Diapir location and style: Most big diapirs are in the inner basin, where deep salt was thickest. Farther seaward, most salt structures are smaller diapirs or salt-cored folds. In some areas, seaward-leaning diapirs are rooted at the crest of the inner ramp, along the LOC.
2. Fold-belt location and style. Cenozoic fold belts were caused partly by gravity gliding related to cratonic uplift. The rift-aged Toledo Bend flexure influenced fold-belt location when it was reactivated during this uplift. Where the Toledo Bend flexure was close to the deep-water area, to the west of the Brazos transfer fault, shortening was focused at the seaward end of the deep salt in the Perdido fold belt. Furthermore, fold-belt structural style was controlled partly by the abundance and style of precursor salt structures, which are tied to original salt thickness.
3. Geometry of the Sigsbee salt canopy: Most of the salt in the Sigsbee canopy originated in the inner basin, where deep salt was thickest. Once up in the canopy, salt flowed southward, locally more than 100 km (62 mi), carrying a flotsam of minibasins and sutures. Much more salt was available to supply the Sigsbee canopy east of the Brazos transfer fault than to the west. Furthermore, the rising Perdido fold belt, which was displaced seaward by the proximity of basement uplift, blocked advance of the Sigsbee canopy west of the Brazos transfer fault. In the central domain, the canopy was able to advance unimpeded. As a result, the Keathley Canyon megasuture between the two asymmetric parts of the canopy is approximately 150 km (93 mi) west of the transfer fault.
4. Subcanopy hydrocarbon prospectivity: Salt thickness controls salt structural styles and, therefore, structural trap styles. Four-way dip closures above salt-cored folds dominate above the thin salt in the outer ramp. Turtle structures and diapir-flank traps become more common as the deep salt thickens updip. Reservoir thickness also becomes more variable updip as the depositional setting above thicker salt becomes more akin to a minibasin. The influence of salt thickness on petroleum systems is reflected in a map of subcanopy discoveries, which follows province boundaries. Early advance of salt carried Tithonian source rocks above the salt, duplicating the source rocks in parts of the outer basin and leaving Tithonian-free

extensional gaps somewhere updip of the Walker Ridge salient.

Mapping of salt provinces in the deep-water northern Gulf of Mexico thus provides a template to improve our understanding of structural, stratigraphic, and exploration trends. Our maps will doubtlessly benefit from revision as newer and better data are acquired. However, our primary theme—that salt configuration inherited from rifting controlled subsequent evolution—should remain robust.

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