Understanding the kinematics of salt-bearing passive margins:  
A critical test of competing hypotheses for the origin of the Albian Gap, Santos Basin, offshore Brazil

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ABSTRACT
Thick-skinned gravitational gliding and spreading drive deformation on salt-bearing passive margins. Such margins typically have an updip extensional domain kinematically linked to a downdip contractional domain. However, calculating magnitudes of extension and shortening in salt-bearing margins is difficult because the initial widths of diapirs are uncertain. Extension and shortening may be cryptic, being hidden in widening or shortening of diapirs. This uncertainty can lead to controversy in regional analysis. The Santos Basin, offshore Brazil, contains a prime example of this uncertainty in the form of an enigmatic structure known as the "Albian Gap," a zone up to 75 km wide within which the Albian section is missing. The Albian Gap has been variably interpreted as the product of post-Albian extensional faulting (the extension model) or as an Albian salt structure evacuated in response to loading by post-Albian sediments (the expulsion model). We evaluate these two models by: (1) structurally restoring a regional seismic-reflection profile across the Albian Gap using both models; (2) quantitatively analyzing the geometry of the Upper Cretaceous rollover overlying the Albian Gap; and (3) synthesizing and critically evaluating arguments previously advanced in support of extension or expulsion. We propose a revised model for the evolution of the Albian Gap that invokes Albian thin-skinned extension and post-Albian salt expulsion. Our approach shows that critical analysis of geological observations from borehole-constrained seismic-reflection data can be used to assess the relative roles of the key processes in the deformation of salt-bearing passive margins.

INTRODUCTION
Passive margins typically respond to thick-skinned, gravity-driven deformation above a salt- or shale-rich detachment by developing an updip extensional domain and a downdip contractional domain. These may be separated by a domain of translation in which neither extension nor shortening occurs (Fig. 1A; Hossack, 1995; Marton et al., 2000; Brun and Fort, 2004, 2011; Hudec and Jackson, 2004; Rowan et al., 2004; Butler and Paton, 2010; Brun and Fort, 2011). Downdip shortening accommodates updip extension, although resolving this kinematic link may be complex due to broadening or narrowing of these domains over time, and to overprinting of structural styles (Marton et al., 2000; Gottschalk et al., 2004; Hudec and Jackson, 2004). Critical factors in our understanding of salt-bearing passive-margin kinematics are the methods we apply to measure updip extension and downdip shortening. For example, a common assumption is that all map-view and cross-sectional stratigraphic gaps are related to extension, so that the magnitude of updip extension can be constrained by summing fault heaves and diapir widths in the updip domain (Fig. 1A; e.g., Rouby et al., 1993, 2002; Cartwright and Jackson, 2008; Butler and Paton, 2010). Likewise, measuring stratigraphic overlaps in thrust faults and shortening accommodated by buckle folds can provide an estimate of the magnitude of downdip contraction (Fig. 1A; e.g., Coward and Stewart, 1995; Stewart, 1996; Butler and Paton, 2010).

However, using these techniques to constrain the magnitude of updip extension and downdip shortening on salt-bearing passive margins can be problematic for several reasons (Hossack, 1995; Rowan and Ratliff, 2012). First, apparent extensional gaps may be related to syndepositional passive diapirs rather than faults, which would overestimate the magnitude of extension. Second, gravity-driven shortening may be principally accommodated by diapir squeezing (Fig. 1B; Vendeveille and Nilsen, 1995; Gottschalk et al., 2004; Dooley et al., 2009, 2015a). Constraining the initial width of diapirs and spacing of minibasins may therefore be problematic. Third, minibasin strata can be folded simply by sinking into autochthonous or allochthonous salt (Rowan et al., 2004; Hudec et al., 2009). Thus, folding is not necessarily diagnostic of shortening. Fourth, salt at the seaward edge of the basin may be thrust over oceanic crust but may not be recognized as a contractual feature (Fig. 1C; Hudec and Jackson, 2004). Fifth, salt overhangs at minibasin flanks can simply be the result of salt rise rate exceeding sediment accumulation rate and may not be diagnostic of shortening (Jackson et al., 1994; Giles and Rowan, 2012). Because of these limitations in analytical techniques and uncertainty in the significance of observations from seismic-reflection profiles, our understanding of the kinematics and therefore dynamics of salt-bearing passive margins is incomplete. Understanding the kinematics and dynamics of such margins is important because they underpin our ability to balance deformation on passive margins and, therefore, determine the geodynamics of their evolution.

This paper focuses on the Santos Basin, offshore SE Brazil, which is one of several salt-bearing passive-margin basins that flank the South Atlantic (Fig. 2). There is abundant evidence for thin-skinned, gravity-driven extension at the updip margin of the Santos Basin, and some evidence for shortening further downdip (Cobbold et al., 1995). Some authors have noted an apparent imbalance between the magnitude of extension and shortening (e.g., Quirk et al., 2012). The Santos Basin is...
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In this section, we review the regional tectono-stratigraphic framework for our detailed analysis of the Albian Gap. The Santos Basin initially formed as a rift basin during the Early Cretaceous as the South Atlantic began to open (e.g., Meisling et al., 2001; Modica and Brush, 2004; Karner and Gambôa, 2007; Mohriak et al., 2008; Contreras et al., 2010). Grabens and half grabens were filled by largely Barremian, fluvial-lacustrine deposits (Picarras and Itapema Formations), which are over lain by an early-to-middle Aptian, carbonate-dominated succession (Barra Velha Formation). During the Aptian, the number of active upper-crustal faults and the rate of slip on these faults a particularly appropriate location to study passive-margin kinematics because of the presence of the “Albian Gap,” a suite of enigmatic, large-scale salt-related structures near the basin margin (Figs. 2 and 3A). The aim of our study is to test two end-member hypotheses for the origin of the Albian Gap: thin-skinned extension and progradation-driven salt expulsion. To achieve this aim, we: (1) evaluate the kinematic linkage between extensional and contractional structures by structurally restoring a regional seismic-reflection profile across the Albian Gap and associated structures; (2) geometrically analyze the Upper Cretaceous rollover overlying the Albian Gap; and (3) synthesize and critically evaluate arguments previously advanced in support of the extension or expulsion hypotheses. We demonstrate that critical analysis of geological observations from borehole-constrained seismic-reflection data can be used to assess the relative roles of the key processes deforming salt-bearing passive margins.

TECTONO-STRATIGRAPHIC FRAMEWORK OF THE SANTOS BASIN

In this section, we review the regional tectono-stratigraphic framework for our detailed analysis of the Albian Gap. The Santos Basin initially formed as a rift basin during the Early Cretaceous as the South Atlantic began to open (e.g., Meisling et al., 2001; Modica and Brush, 2004; Karner and Gambôa, 2007; Mohriak et al., 2008; Contreras et al., 2010). Grabens and half grabens were filled by largely Barremian, fluvial-lacustrine deposits (Picarras and Itapema Formations), which are overlain by an early-to-middle Aptian, carbonate-dominated succession (Barra Velha Formation). During the Aptian, the number of active upper-crustal faults and the rate of slip on these faults...
decreased and, during the latest Aptian, a thick (~2.4–2.6 km; De Freitas, 2006; Davison et al., 2012), salt-rich succession was deposited (Ariri Formation; Figs. 2 and 4).

During the Albian, in response to postrift, thermally induced subsidence and eustatic rise of sea level, fully marine conditions became established in the Santos Basin, and a carbonate-dominated succession was deposited (Itanhaem Formation; Fig. 4). Subsidence was focused toward the center of the oceanic spreading center and resulted in southeastward tilting of the basin, downdip salt flow (Davison et al., 2012; Fiduk and Rowan, 2012; Quirk et al., 2012), and an array of thin-skinned, gravity-driven, predominantly seaward-dipping normal faults that stretched the Albian carbonate platform and locally formed extensional rafts at the basin margin (Fig. 3B; see also “zone of extension” in Fig. 2A; Demercian et al., 1993; Rouby et al., 1993; Cobbold et al., 1995; Mohriak et al., 1995; Guerra and Underhill, 2012; Quirk et al., 2012). These faults accommodated up to 40 km of extension (Quirk et al., 2012). Based on uppermost Albian growth wedges in the hanging walls of these faults (Fig. 3B), these structures were likely active during the latest Albian (see following). The Albian Gap, the focus of our study, is located at the seaward end of this domain of extension (Figs. 2 and 3A). A detailed description of the Albian Gap and associated structures and an outline of competing hypotheses for its origin are presented in subsequent sections.

A rapid rise in eustatic sea level during the Cenomanian–Turonian abruptly increased water depth in the Santos Basin, drowning the Albian carbonate system as a fine-grained, clast-dominated succession accumulated (lower part of the Itajai-Acu Formation; Fig. 4). During the Coniacian–Maastrichtian, the eastern coastal ranges of Brazil were uplifted and eroded and the basin margin prograded southeastward into the Santos Basin (upper part of the Itajai-Acu Formation; e.g., Cobbold et al., 2001; Meisinger et al., 2001; Medica and Brush, 2004). Most faults in the Albian carbonate platform became inactive by the end of the Albian (Fig. 3B).

WHAT IS THE ALBIAN GAP?

A curious and much debated feature of the Santos Basin is the Albian Gap. This spectacular and enigmatic structure is defined as a plan-view and cross-sectional “gap” in an ~1-km-thick succession of Albian carbonates (Itanhaem Formation) that directly overlies an evaporite-dominated, latest APTian succession (Ariri Formation; Figs. 2 and 3A; Demercian et al., 1993; Cobbold et al., 1995; Mohriak et al., 1995; Gemmer et al., 1997; Modica and Brush, 2004; Gemmer et al., 2005; Davison et al., 2012; Guerra and Underhill, 2012; Quirk et al., 2012). Most publications indicate that the Albian Gap trends NE, is up to 75 km wide, has a strike length of at least 200 km and possibly as much as 320 km, and is up to 7700 km² in map-view area (Figs. 2 and 3A). The exact map-view geometry of this gap is uncertain: Some authors suggest the feature is relatively continuous along strike (Fig. 2A; e.g., Mohriak et al., 1995), whereas others interpret at least three discrete segments (e.g., Modica and Brush, 2004; Guerra and Underhill, 2012; Quirk et al., 2012). Above the Albian Gap, Upper Cretaceous deep-water clastics (Itajai-Acu Formation) are folded into a largely un faulted southeastward-facing rollover that is, in some ways, more striking than the stratigraphic gap itself (Figs. 2B and 3A). Upper Cretaceous strata above the gap either overlie thin, residual evaporites or are welded to subsalt, Aptian carbonates (Barra Velha Formation). Along much of its length, the Albian Gap marks the southeastern, downdip limit of the basin-margin extensional province (Fig. 2A). The gap is bounded on its southeastern (seaward) margin by a large diapir, up to 10 km wide, that has relief of up to 4 km. Further seaward are large diapirs and minibasins (Figs. 2B and 3A; Demercian et al., 1993; Cobbold et al., 1995; Mohriak et al., 1995; Modica and Brush, 2004; Davison et al., 2012; Fiduk and Rowan, 2012; Guerra and Underhill, 2012; Quirk et al., 2012).

HOW DID THE ALBIAN GAP FORM?

Dynamic versus Kinematic Models

Discussion on the evolution of the Albian Gap has revolved around two types of models. The first are dynamic models, focusing on the driving forces that created the Albian Gap. The second are kinematic models, focusing on the timing and geometry of sedimentary structures in and around the Albian Gap. Some misunderstandings have arisen by confusing the two types of models.

At the broadest scale, passive-margin dynamics are a result of varying combinations of gravity gliding and gravity spreading. As summarized by Peel (2014, synthesizing earlier work by Ramberg [1967, 1977, 1981a, 1981b] and De Jong and Scholten [1973]), gravity gliding refers to deformation in which gravitational potential energy is lost by lowering the center of gravity due to movement along an inclined detachment, whereas spreading describes deformation in which energy is released by lowering the center of gravity due to thinning of the material. There has been considerable debate of the relative roles of these two processes in passive-margin dynamics (e.g., Brun and Fort, 2011, 2012; Rowan et al., 2012; Peel, 2014).

Quirk et al. (2012) proposed a hybrid dynamic model for South Atlantic passive margins that they termed “salt drainage.” In salt drainage, deformation of postsalt sediments is caused by downslope, viscous flow of the underlying salt. Drainage would thus involve both gliding and spreading; the key diagnostic feature is that salt flow drives deformation, and sediments are merely carried along. As advocated by Quirk et al. (2012), drainage may be important on passive margins, especially early in margin evolution when postsalt strata are thin and weak.

In this paper, however, we focus on distinguishing between kinematic rather than dynamic models for the origin of the Albian Gap. Two such models predominate in the literature. In the first, sediments on either side of the Albian Gap were joined at the end of the Albian, and their current separation is a result of post-Albian extension (the “extension model”). In the second, the Albian Gap had already formed by the end of the Albian and was occupied by a broad, low-amplitude salt wall that in post-Albian time was expelled seaward from beneath a prograding wedge of sediments (the “expulsion model”). We emphasize that these two kinematic models may be considered in isolation from a debate on margin dynamics; spreading, gliding, and drainage may have occurred in either kinematic model. We now turn to a more detailed description of the two kinematic models.

Figure 2 (on following page). (A) Regional map showing the main basement-involved structures and salt-related structural domains of Santos Basin, offshore Brazil (modified from Davison et al., 2012; Modica and Brush, 2004). Map shows locations of Figures 2B and 3A–3B, the regional line used for the restoration in Figure 10, and the three-dimensional (3D) seismic reflection survey from which the seismic profiles in Figure 9 were taken. (B) Geoseismic section across Santos Basin and the Albian Gap, modified from Carminatti et al. (2008), de Melo Garcia et al. (2012), and Davison et al. (2012). The approximate locations of four of the boreholes used to constrain the seismic interpretation are projected onto this line (723C and 532A). CFF—Cabo Frio fault. UTM 24 South. Datum—WGS84.
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Figure 2.
Figure 3. (A) Interpreted seismic section through the center of the Albian Gap, which is clearly visible as a cross-sectional gap in the ~1-km-thick Albian carbonate succession (Itanhaém Formation). Red dots indicate the Albian cutoffs defining limits of the gap. Paired black dots indicate salt welds. The huge seaward-facing rollover in the Upper Cretaceous succession is underlain by a salt weld and by small salt diapirs. Turonian to Santonian strata display bulk seaward thinning (unit labeled “A”), whereas Campanian to Eocene strata display bulk seaward thickening (unit labeled “B”). Seismic data are courtesy of Fugro. (B) Geoseismic section through the northern Santos Basin illustrating the structural style of the extensional domain (modified from Guerra and Underhill, 2012). TWT—two-way traveltime. Note the small salt rollers formed by Albian extension, a seaward-facing Upper Cretaceous rollover that onlaps a salt diapir, and the lack of a major landward-facing fault such as the Cabo Frio fault. The Upper Cretaceous rollover has bipartite stratigraphy: Albian to Maastrichtian strata thin seaward (unit labeled “A”), whereas Maastrichtian to Miocene strata thicken seaward (unit labeled “B”). For location of both cross sections, see Figure 2.
Extension Model

The extension model envisages that the Albian Gap formed mostly in post–early Albian time by gravity-driven extension of overburden, which formed the gap in a previously continuous Albian carbonate succession (Fig. 5A; see also Figs. 1A–1C; Demercian et al., 1993; Cobbold et al., 1995; Mohriak et al., 1995; Davison et al., 2012; Fiduk and Rowan, 2012; Guerra and Underhill, 2012; Quirk et al., 2012). The extension model implies that the contact between Upper Cretaceous strata and Aptian salt along the southeast margin of the Albian Gap is a listric fault system (Figs. 2B and 5A), comprising either a single through-going segment (Mohriak et al., 1995) or several soft-linked segments (Fig. 2A; Modica and Brush, 2004; Quirk et al., 2012). This “Cabo Frio fault” is envisaged as a seaward-dipping listric normal fault system having at least 3 km of throw, and a total strike length of at least 200 km. It merges downdip into a subhorizontal salt weld that extends up to 55 km landward from the southwest edge of the Albian Gap, thus defining a fault-related separation of the Albian strata of comparable width. Kinematically, the extension model implies that the entire footwall of the Cabo Frio fault was translated up to 55 km basinward during the main period of Late Cretaceous extension (Fig. 5A), gradually separating from the hanging wall to create the Albian Gap. Progradation of basin-margin clastic wedges during the Late Cretaceous was broadly synchronous with faulting, and these sediments filled the deep-water basin above the Albian Gap, landward of the Cabo Frio fault (Fig. 5A). Slip on the listric Cabo Frio fault formed a seaward-facing extensional rollover in Upper Cretaceous strata and resulted in apparent downlap of these units against subsalt strata where salt is absent (Fig. 5A; Demercian et al., 1993; Mohriak et al., 1995; Szatmari et al., 1996; Quirk et al., 2012; Rowan and Ratliff, 2012). Because salt is significantly thinned or even locally absent in the Albian Gap, the extension model also involves salt thinning and expulsion.

Expulsion Model

In contrast to the extension model, the expulsion model infers that the Albian Gap already existed by the end of the Albian, when it was occupied by an extremely wide, low-relief salt wall (or series of salt walls) in contact with Albian sediments on either side. The monocline sediment rollover above the Albian Gap formed during the Late Cretaceous by seaward expulsion of salt driven by progradation of basin-margin clastic wedges across the salt wall (Fig. 5B; cf. Fig. 1D). The area of missing Albian strata in the gap marks the location of the former salt wall, so this model requires that the low, broad diapir covered at least the same area in map view (i.e., 7700 km²) as the present Albian Gap itself, and that the diapir was as tall as the Albian strata flanking it where thick (~1 km; Fig. 5B).

In the expulsion model, only limited post-Albian extension of the overburden need have occurred to form the Albian Gap (as modeled by Ge et al., 1997; Gemmer et al., 2005; Albertz and Ings, 2012). Instead, salt flow was driven by the difference in density between sediments within the seaward-dipping, progradational clastic wedge that drove expulsion and adjacent seawater (see “sedimentary topographic loading” of Hudec and Jackson, 2007). This density difference would be present even in entirely unconsolidated sediments. A critical difference between the expulsion and extension models is that the former implies: (1) the contact between Upper Cretaceous strata and Aptian salt is a stratigraphic onlap surface rather than a fault; and (2) overburden on the seaward side of the gap did not need to undergo a significant amount of post-Albian basinward translation to form the gap, although the model entails massive seaward flow of salt (Fig. 5B). In the expulsion model, the broad Upper Cretaceous monocline could be interpreted as being purely structural, resulting from basinward rotation of gently dipping strata that define depositional clinoforms (Ge et al., 1997; Gemmer et al., 2005; Krézsek et al., 2007; Albertz and Ings, 2012). However, Hadler-Jacobsen et al. (2010) used three-dimensional seismic-reflection and borehole data to demonstrate that Upper Cretaceous strata in the rollover display a consistent downdip transition from shelf to slope to basin-floor environments, implying that, during basin margin progradation, a primary depositional dip (as clinoforms) was present, which was later steepened by removal of underlying salt.

The expulsion model for the Albian Gap thus assumes that by the end of the Albian, a wide diapir existed there, from which salt was expelled during Late Cretaceous progradation. What could have formed such a broad diapir? It could have grown by simple passive diapirism, but alternatively it could have been initiated as a reactive diapir during Albian extension (see following section titled “An Alternative Model for the Origin of the Albian Gap”). If so, the expulsion model would incorporate large magnitudes of Albian extension of overburden, followed by post-Albian salt expulsion. By contrast, the extension model requires large magnitudes of post-Albian extension of overburden. Both models, then, would involve extension-driven opening of the Albian Gap, but at different
times. The terms “extension model” and “expulsion model” are thus somewhat misleading. We keep the terms for historical consistency, but stress ages of deformation in the text for clarity.

KINEMATIC SCENARIOS FOR GRAVITY-DRIVEN LINKED STRUCTURAL SYSTEMS

The Albian Gap lies within a domain in which both thin-skinned extension and salt expulsion could conceivably have occurred, hence the competing hypotheses for its origin. To set the scene for the analysis that follows, we outline here the key kinematic scenarios that can occur along salt-bearing passive margins. For each of the two end members (thin-skinned extension and salt expulsion), we focus on the style and timing of deformation in different parts of the system and the key diagnostic observations. For simplicity, we illustrate systems that have: (1) a single isopachous, prekinematic layer that records finite strain throughout the system; and (2) a single synkinematic wedge that provides a stratigraphic record of the style, location, and timing of deformation (Figs. 1A–1D).

Thin-Skinned Extension

Thin-skinned extensional systems are widely recognized in proximal parts of salt-bearing passive margins. Extension is typically accommodated by an array of seaward-dipping and, more rarely, landward-dipping listric normal faults, which detach downward onto a seaward-dipping salt layer (Figs. 1A–1C). Small, typically triangular salt diapirs termed “salt rollers” form in the footwalls of these normal faults (e.g., Jackson and Cramez, 1989; Duval et al., 1992; Vendeville and Jackson, 1992; Brun and Mauduit, 2009). After mild extension, prekinematic fault blocks remain in contact with one another and are only weakly rotated (known as “prerafts”; Jackson and Cramez, 1989). The autochthonous salt layer is generally not completely welded below prerafts. However, after extreme extension, fault blocks are strongly rotated and may become completely separated to form “rafts,” which are separated from their neighbors by extensional gaps (Burollet, 1975; Duval et al., 1992; Gaullier et al., 1993; Chimney and Kluth, 2002). Salt flow may create a weld that separates presalt and suprasalt strata. Overburden translation exerts local shear traction on the salt, which causes the salt to shear and flow in Couette fashion. This combines with a component of squeezing flow exerted by the load of the overburden (Figs. 1A–1C) to yield a hybrid type of salt flow. Thickness

Figure 5. Restorations modified from Rowan and Ratliff (2012) illustrating two end-member hypotheses for the origin of the Albian Gap. (A) Sequential restoration assuming an origin as a landward-dipping fault (CFF—Cabo Frio fault). (B) Sequential restoration of an expulsion rollover. No vertical exaggeration. Paired black dots indicate salt welds. Note that the final margin-scale geometries are identical (compare them to Fig. 3A) despite markedly different kinematics (i.e., extension vs. expulsion). Restorations, which are based on seismic interpretation by Mohriak and Szatmari (2008), were conducted in LithoTect® using vertical-simple shear and method of Rowan (1993; for more details, see Rowan and Ratliff, 2012).
variations in synkinematic strata, and the locations of the upper tips of the faults, document the timing of extension in the updip domain (Figs. 1A–1C).

A key feature of the extension model is that it results in distal shortening if salt is constrained downdip by a topographic barrier. Even if the distal margin of the salt basin is unconstrained, shortening still occurs there but takes the form of an overflowing salt nappe, the overburden of which may be carried seaward to duplicate the stratigraphy of subsalt strata. Shortening can be accommodated in several ways by thin-skinned buckle folding and related thrusting (Fig. 1A), diapir squeezing, inversion of preexisting extensional structures, and emplacement of an overthrust salt nappe (cf. “source-fed thrust” of Hudec and Jackson, 2004) out beyond the limits of the salt basin (Fig. 1C). The timing of this distal deformation is recorded by synkinematic strata that thin across compressional folds or are erosionaly truncated as they are uplifted (Figs. 1A and 1C). If a thin-skinned system is kinematically self-contained (sensu Butler and Paton, 2010), the magnitudes of distal shortening and proximal extension should balance. However, many systems are not discernibly balanced. In particular, the magnitude of shortening visible in the distal domain is typically less than the magnitude of extension in the proximal domain. For example, based on line-length balancing of two-dimensional seismic-reflection data from offshore Namibia, Butler and Paton (2010) documented an 18%–25% longitudinal strain shortfall in the shortening domain of an overpressured mudstone-based, thin-skinned system. Henry et al. (2003) documented lower but still significant values (12%) of horizontal subseismic shortening based on core samples from the Nankai subduction-accretion thrust system.

The missing strain in large linked systems has been accounted for by one or more of the following mechanisms. First, lateral compaction of weakly lithified, shallowly buried sediment in the system absorbs shortening by the collapse of pore space and dewatering (“distributed volume loss”; sensu Butler and Paton, 2010). For example, in the eastern Mediterranean, proximal extension began in the Pliocene, but distal shortening only began in the Pleistocene. Cartwright et al. (2012) attributed this lag to the buffering effect of lateral compaction, which absorbed initial translation between the two domains. Second, convergent or divergent extension and shortening cause a progressively larger amount of out-of-plane movement of material with increasing distance seaward. Thus, some shortening strain is not captured in the plane of analysis (Cobbold et al., 1995).

Third, strain can be cryptic in kinematically linked systems if shortening occurs by diapir squeezing (Fig. 1B). Diapir squeezing is real but can be hidden if the diapirs were exposed at the surface and so had no roof to record the deformation. Squeezed passive diapirs become narrower by expelling salt upward to be dissolved or flow onto the free surface, or even downward back into the source layer (Vendeville and Nilsen, 1995; Gottschalk et al., 2004; Dooley et al., 2009). Finally, overthrusting of a salt nappe can involve considerable shortening (20–30 km in the case of the Lower Kwanza Basin, offshore Angola; Hudec and Jackson, 2004) in thin-skinned systems that are not kinematically self-contained (Fig. 1C).

Salt Expulsion

Although they are more poorly documented than extensional systems, large expulsion rollover or rollover systems can also develop in salt-bearing basins in response to differential sedimentary loading (Figs. 1D and 6; Ge et al., 1997; Rowan and Inman, 2005, 2011; Hudec and Jackson, 2007; Krézsek et al., 2007; Trudgill, 2010; Bruns and Fort, 2011; Allen and Beaumont, 2012; Rowan and Ratliff, 2012). One way to trigger salt flow in them is by sedimentary topographic loading during progradation of clastic wedges from the basin margin. Juxtaposition of a wedge of relatively dense (>1700–2000 kg/m³) clastic sediment against water (density 1000 kg/m³) at the basin margin results in a lateral pressure gradient that may expel most or all of the salt from beneath the wedge (Hudec and Jackson, 2007; Hudec et al., 2009). Any expelled salt flows basinward and, if contained within its original depositional basin, undergoes intrasalt horizontal shortening and thickens to form an inflated salt plateau or broad salt wall (Figs. 1D, 6A, and 6B; Ge et al., 1997; Hudec and Jackson, 2004; Albertz and Ings, 2012).

The structural style of expulsion-related systems differs strongly from that of purely extensional systems. First, little or no extension need occur; sedimentary wedges simply sink into the salt and are rotated to form a basinward-facing expulsion rollover as salt is expelled progressively seaward (Figs. 1D, 6A, and 6B; Ge et al., 1997; Gemmer et al., 2005; Krézsek et al., 2007; Trudgill, 2010; Albertz and Ings, 2012). Minor outer-arc extension associated with rollover folding may form small normal faults at the rollover apex (Ge et al., 1997; Rowan et al., 1999). Second, numerical modeling (Gemmer et al., 2005; Albertz and Ings, 2012) and physical modeling (Ge et al., 1997) show that the former onlap strata steepen seaward and become apparent downlaps once a basal weld has formed.

Third, basinward expulsion of salt and rise of the inflated salt plateau lift the prekinematic overburden. If present, synkinematic layers above the plateau are thin, being the distal equivalent of the basin-margin wedges. This uplift can be achieved without regional shortening of overburden above the rising salt plateau (Figs. 1D, 6A, and 6B; Hudec and Jackson, 2004). In contrast, intrasalt shortening is inherently widespread and locally very intense within the rising salt plateau (Fig. 6B; Albertz and Ings, 2012; Cartwright et al., 2012).

CRITICAL EVALUATION OF GEOLOGICAL OBSERVATIONS

Most authors have inferred a purely post–early Albian extensional origin for the Albian Gap and associated rollover (Fig. 5A; Demercian et al., 1993; Mohriak et al., 1995; Szatmari et al., 1996; Modica and Brush, 2004; Davison et al., 2012; Fiduk and Rowan, 2012; Guerra and Underhill, 2012; Quirk et al., 2012; Rowan and Ratliff, 2012); a minority favors an expulsion origin (Fig. 5B; Ge et al., 1997; Gemmer et al., 2005; Krézsek et al., 2007). The paper of Quirk et al. (2012) is notable as the most thorough attempt to systematically evaluate a series of observations to assess the origin of this remarkable structure (building on a previous discussion by Mohriak et al., 1995). Motivated by Quirk et al. (2012), we critically evaluate the key geological features of the Albian Gap, including the criteria of Quirk et al. (2012) and other proponents of the extension model, augmented by new observations. In the following six subsections, “positive evidence” refers to geological features actually observed; these criteria may or may not be diagnostic of a particular genesis. “Negative evidence” refers to geological features predicted by the extension or expulsion model but that are not seen; by inference, this missing evidence is considered to favor the other model. Within this framework, we identify the following: (1) observations that support both models, (2) observations that support the extension model but not the expulsion model, and (3) observations that support the expulsion model but not the extension model.

Positive Evidence Compatible with both Post-Albian Extension and Post-Albian Expulsion

Location of the Albian Gap

The Albian Gap is located in the proximal domain of a salt-bearing passive margin (Fig. 2A), where thin-skinned, gravity-driven
Figure 6. Examples of rollovers generated by salt expulsion. (A) Physical model from Ge et al. (1997); geometry is similar to Figure 3B. (B) Numerical model from Albertz and Ings (2012). (C) Gorleben salt wall, Germany; expulsion rollovers are in the southeast (see Ge et al., 1997). (D) Paradox Basin, Utah; expulsion rollovers are in the northeast (Trudgill, 2010). (E) Pre-Caspian Basin, Kazakhstan; expulsion rollovers are in the east (image courtesy of Condor Petroleum). Note the complex intrasalt deformation in B in response to seaward expulsion and thickening of salt without extension of the proximal overburden. Paired black dots indicate salt welds. Seaward thinning and thickening wedges in expulsion-related rollovers are labeled in sections A–D; these wedges are also apparent in section E but are too small to label on this regional profile. VE—vertical exaggeration.
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Salt expulsion is also likely in such a setting. During the Late Cretaceous progradation was subsequently evacuated by overburden of a former low, wide salt wall from which salt an extensional origin. Thus, the large for thin-skinned extension on a passive Cretaceous (Fig. 5A). This heave is unusually ward translation of its footwall during the Late created by slip on the Cabo Frio fault and sea-

extension model, this area is the extensional gap

Areal Extent of the Albian Gap

As mentioned already, the Albian Gap cov-
ers an area of up to 7700 km² (Fig. 2A). In the extension model, this area is the extensional gap created by slip on the Cabo Frio fault and sea-
ward translation of its footwall during the Late Cretaceous (Fig. 5A). This heave is unusually large for thin-skinned extension on a passive margin, but it is certainly possible. Thus, the areal extent of the Albian Gap is consistent with an extensional origin.

In the expulsion model, the gap is the area of a former low, wide salt wall from which salt was subsequently evacuated by overburden loading during Late Cretaceous progradation (Fig. 5B). The Albian salt required by the expulsion model could have formed either by passive diapirism or by Albian extension. The 7700 km² dimensions are large but plausible for either origin. In the case of a passive diapir, this extremely broad salt structure would be comparable in area to salt walls in the Whale Basin, offshore Newfoundland (Balkwill and Legall, 1989), or Precaspian Basin, Kazakhstan (Volozh et al., 2003). For an extensional origin, scaled physical models indicate that very broad and long salt walls can also form during thin-skinned gravity-driven extension (Fig. 7).

Seaward Thickening of Upper Cretaceous Strata in Rollover

Upper Cretaceous to Lower Tertiary strata dip and thicken seaward within the suprasalt rollover above the Albian Gap (Fig. 3A). Most authors interpret the overall geometry of the rollover and the detailed seismic-stratigraphic architecture of the Upper Cretaceous interval to indicate that the Albian Gap is the inactive hanging wall of a landward-dipping, listric normal fault defining its southeastern margin (Fig. 5B; the Cabo Frio fault; Demercian et al., 1993; Mohriak et al., 1995; Modica and Brush, 2004; Davison et al., 2012; Quirk et al., 2012; Rowan and Ratliff, 2012). Such seaward-facing extensional rollovers are common in thin-skinned, gravity-driven systems (Figs. 3B and 8; Lundin, 1992; Roubly et al., 1993, 2002; Diegel et al., 1995; Anderson et al., 2000; Cartwright and Jackson, 2008; Brun and Mauduit, 2009). However, physical modeling (Fig. 6A; Ge et al., 1997) and numerical modeling (Fig. 6B; Gemmer et al., 2005; Albertz and Ings, 2012) indicate that seaward-facing rollovers can form equally well during salt expulsion. Thus, a rollover is not unique to either model, although we argue later herein that the geometry of the rollover in the Albian Gap is more consistent with expulsion than with extension.

Thin Salt in the Albian Gap

Aptian salt is very thin in the Albian Gap. Where salt is inferred to be locally absent, sub-salt Aptian and suprasalt Santonian strata are in direct contact across a salt weld (Figs. 2B and 3A). The lack of salt in the Albian Gap and the observed stratigraphic juxtaposition have been taken as evidence that most of the Aptian salt flowed seaward by the middle Santonian, and little or no salt flowed after this time (Quirk et al., 2012). However, the presence of thin or welded salt is also compatible with the expulsion model because progradation of a basinward-thinning clastic wedge would also drive massive seaward expulsion of salt simply owing to sedimentary topographic loading, even in the absence of a density inversion (Figs. 5B, 6A, and 6B). Thus, we consider the thinness of salt in the Albian Gap to be nondiagnostic.

Thickness of Overburden when Deformation Began

Quirk et al. (2012) estimated that the Upper Cretaceous clastic wedge was only 1000–1800 m thick when it began to subside into the Albian Gap. They argued that this wedge was too thin and thus not dense enough to sink into the salt and expel it. However, simple mechanical arguments summarized by Hudec et al. (2009) indicate that “sedimentary topographic loading” (the juxtaposition of a relatively dense sedimentary wedge against a less dense water column) would initiate salt expulsion, irrespective of overburden thickness and density. We therefore conclude that the thin overburden at the onset of rollover formation is consistent with both extension and expulsion models.

Thickest Upper Cretaceous Shelf Succession Deposited West of the Albian Gap

The thickest (>2.5 km) succession of Upper Cretaceous shelf strata was deposited southwest of the center of the Albian Gap (Fig. 2A; Quirk et al., 2012). This anticorrelation between the thickest shelf succession and the Albian Gap led Quirk et al. (2012) to argue that basin-margin progradation could not have driven salt expulsion and that the Albian Gap more likely formed by extension. However, our seismic data, and almost all published seismic interpretations, show up to 4 km thickness of Upper Cretaceous shelf and slope strata immediately overlying the Albian Gap (Figs. 2B and 3A; see also fig. 14 in Quirk et al. 2012). We infer that the combined thickness of basin-margin strata, including both shelf and slope deposits, would clearly be sufficient to drive salt expulsion by sedimentary topographic loading. The location of the Albian shelf depocenters is therefore not conclusive evidence for either model.

Flip-Flop Salt Walls

Inherent in the extension model is that the SE margin of the Albian Gap is bounded by a landward-dipping, listric normal fault (Fig. 5A; the Cabo Frio fault). However, Quirk et al. (2012) illustrated relatively small, seaward-dipping, supradiapir faults—which were only active during the late Albian and early Cenomanian, and which are now crosscut by the much larger, landward-dipping structure—occur locally along the SE margin of the Albian Gap (fig. 8 in Quirk et al., 2012). They argued that these so-called “flip-flop” salt walls and associated normal faults are evidence that the Albian Gap has an extensional origin. However, the hypothetical precursor diapir inherent in the expulsion model could also have undergone late Albian–to–early Cenomanian extension before loading and seaward salt expulsion began (see following). Thus, late Albian–early Cenomanian extension is consistent with both models.

Balance between Updip Extension and Downdip Shortening

The extension model requires that post-Albian extension across the Albian Gap must pass downdip into a domain of kinematically linked, post-Albian shortening of the overburden (e.g., Fig. 1A; Demercian et al., 1993; Cobbold et al., 1995; Mohriak et al., 1995; Fiduk and Rowan, 2012; Guerra and Underhill, 2012; Quirk et al., 2012). In support of this hypothesis, several authors have pointed to local evidence for intrasalt shortening (Fig. 9A; Demercian et al., 1993; Cobbold et al., 1995; Davison et al., 2012; Fiduk and Rowan, 2012; Jackson et al., 2014). However, physical and numerical modeling of expulsion rollovers shows that the amount of strain in a salt layer may be much more than that for coeval deformation in the overburden (Ings et al., 2004; Gemmer et al., 2005; Albertz
and Ings, 2012; Cartwright et al., 2012). This is because salt that is pumped downslope from beneath the expulsion rollover thickens and shortens the salt in downslope diapirs, forming intrasalt shortening structures that have nothing to do with regional translation of the overburden. The intrasalt structures cannot therefore be used to balance regional updip extension of Albian and younger overburden.

Only overburden contractional structures can be used to balance overburden extension. The amount of visible shortening in Upper Cretaceous strata downdip of the Albian Gap is modest, principally taking the form of low-amplitude folds (Cobbold et al., 1995). Some of these folds are demonstrably contractional (Fig. 9B), but most are ambiguous minibasin synclines, which could have formed by either shortening or salt withdrawal. Even if all the folding were interpreted as due to shortening, cross sections at 1:1 vertical exaggeration show that the supposed shortening is far too small to balance an extensional Albian Gap, a shortfall also noted by Quirk et al. (2012), who estimated that downdip shortening only accounts for ~50% of the updip extension.

Despite appeals to regional kinematics to support an extensional origin for the Albian Gap, we are unaware of any published basin-scale restorations across the Albian Gap that might help resolve the controversy. This absence is likely due to a combination of factors, including limited data access and difficulty in correlating horizons away from extensive

**Figure 7.** Overhead view of physical models of broad salt walls developed during (A) mild, (B) moderate, and (C) extreme thin-skinned extension (images used with the permission of Tim Dooley). Mild extension of a prekinematic overburden triggers the growth of salt-detached normal fault systems and reactive salt walls (A), which become passive but continue to widen as extension continues (B). During the latter stages of extension, very wide extensional gaps form, which are filled by large, flat-topped salt walls (C). For model setup, a 15-mm-thick salt analogue (silicone polymer, SGM36; Weijermars, 1986) was overlain by a 10-mm-thick overburden. Basement dipped 3.5° seaward. Siliciclastic overburden was simulated using brittle, dry, granular materials consisting of blue silica sands and ceramic microspheres having a bulk density matching that of the model salt (for further details, see Dooley et al., 2009). To initiate gravity gliding, the model was tilted by 3° to the left.
well control, which is principally restricted to the basin margin. We here present a restoration of a regional seismic profile showing alternative restorations to the end of the Albian for the expulsion (Fig. 10B) and extension (Fig. 10C) models.

A key feature of the regional section is that we have interpreted the Albian section (and indeed all of the Cretaceous) to be restricted to isolated minibasins in the outer half of the margin, based on the observation that most of the section in the minibasins laps out at the minibasin flanks. We lack well control in these minibasins but are confident based on regional thickness patterns that the Albian cannot be much thicker than we have drawn it (cf. Carminatti et al., 2008; de Melo Garcia et al., 2012). If this is correct, then any restoration to the end of the Albian must feature
Figure 9. Seismic profiles from a three-dimensional seismic volume illustrating shortening-related deformation on the São Paulo Plateau. (A) Intralateral thrusts capped by an intralateral sheet (see Fiduk and Rowan, 2012; Jackson et al., 2014, 2015). (B) Salt-cored buckle folds and related thrusts, some of which are intralateral and some of which deform overburden. See Figure 2A for location of both seismic profiles. Key to seismic horizons is shown in Figure 4. TWT—two-way traveltime.
Figure 10. (A) Regional geoseismic profile used in structural restoration. See Figure 2A for location of profile. We lack seismic data for the updip end of the profile, so used data published by Quirk et al. (2012, their fig. 4a) and Modica and Brush (2004, their fig. 6) to illustrate the proximal margin. (B) Restoration to end-Albian time, assuming that the Albian Gap was occupied by a broad, low diapir at that time (expulsion model). In this restoration, most of the lateral translation occurred during the Albian, so minibasins were located close to their present positions. (C) Restoration to end-Albian time, assuming that the Albian Gap did not yet exist but rather opened in post-Albian time (extension model). The large magnitudes of post-Albian extension were accommodated by shortening salt walls in the seaward half of the profile. The point of the restorations is to show the lateral positions of Albian blocks, so no effort was made to decompact the section or restore basement subsidence and deformation. VE—vertical exaggeration.
isolated pods of Albian strata floating in a vast sea of salt (Fig. 10), as inferred for the Angolan margin (Hudec and Jackson, 2004). The isolation of Albian minibasins is not simply a feature of the particular profile chosen for restoration, but rather is typical of the entire deep-water area seaward of the Albian Gap.

The absence of a continuous Cretaceous section in the deep-water Santos Basin means that there is no way to evaluate magnitudes or locations of Cretaceous shortening in the basin. There are no constraints on the original spacing of these minibasins, so Cretaceous shortening during any stage can be accommodated simply by changing the width of the salt walls that separate the Albian pods (compare Figs. 9B and 9C). Albian restorations using the expulsion and extension models are thus equally valid, so regional restoration cannot distinguish between them.

**Abrupt Thickening of Salt Immediately Downdip of the Albian Gap**

Seismic profiles and maps indicate that the Aptian salt abruptly thickens downdip of the Albian Gap, and that the structure is bounded on its basinward margin by a large salt wall (Figs. 2B and 3B). Advocates of an extensional origin for the Albian Gap argue that abrupt thickening of the salt and the presence of a diapir are evidence for shortening and thickening of the salt caused by basinward flow of salt related to thin-skinned extension (Fig. 5A; e.g., Quirk et al., 2012). However, as we have argued, abrupt downdip thickening of salt is equally consistent with the expulsion model (Fig. 6A). Almost all of the salt within the hypothetical broad Albian diapir in the expulsion model was expelled downdip by sediment loading, so it would be entirely reasonable to find large volumes of salt just downdip of the Albian Gap. In the same way, physical modeling (Ge et al., 1997) and numerical modeling (Gemmer et al., 2004, 2005; Ings et al., 2004; Albertz and Ings, 2012) of salt expulsion show a similar abrupt seaward thickening of salt in response to thinness of the prograding overburden wedge. We therefore suggest that abrupt thickening of salt immediately downdip of the Albian Gap is not diagnostic of either model (cf. end stages of contrasting restorations in Fig. 5).

**Positive Evidence Compatible with Post-Albian Extension but Not Post-Albian Expulsion**

Many observations have been advanced by other authors as supporting post-Albian extension (see previous). We agree with these authors that the observations are consistent with the extension model, but we maintain that the observations are equally consistent with expulsion. We therefore find no arguments that we believe are truly diagnostic of a post-Albian extensional origin of the Albian Gap.

**Upper Cretaceous Rollover Extends Beyond the Northeast Termination of the Albian Gap**

Seismic data indicate that the Upper Cretaceous rollover extends at least 60 km along strike to the northeast of the end of the Albian Gap (Fig. 3B). Here Albian strata are only weakly deformed by predominantly seaward-dipping, low-displacement, listric normal faults that appear to have been most active in the late Albian to early Cenomanian (Fig. 3B). That the rollover exists even in the absence of the Albian Gap and Cabo Frio fault suggests that the rollover is not fault-related and therefore formed instead by salt expulsion.

Furthermore, the stratigraphic succession in the rollover, which is bounded to the SE by a tall (1500 ms two-way traveltime) salt diapir, has a bipartite stratigraphic architecture; the lower, Late Cenomanian to Maastrichtian packages dip and thin seaward, whereas the overlying Paleocene to Miocene packages dip and thicken seaward (Fig. 3B). Based on these observations, in particular, the seaward, rather than landward, thinning of the entire Upper Cretaceous succession, it is implausible to simply attribute the rollover to a large seaward-dipping, listric normal fault. In contrast, an upward transition from basinward-thinning to basinward-thickening would occur as strata: (1) thinned against a rolling monocline formed by salt expulsion, and then (2) thickened as salt was expelled seaward by the prograding wedge and the rolling monocline migrated seaward.

Ge et al. (1997) applied a new methodology, the construction of rollover plots, to quantitatively analyze the geometry of rollovers to differentiate between those formed by expulsion and those formed by extension (see their fig. 4). Rollover plots are constructed by measuring the maximum stratal dip on synkinematic, hanging-wall beds at progressively deeper structural levels at progressively greater distances from the downdip rollover termination (Fig. 11). In an extensional rollover, this termination is the bounding listric fault; in an expulsion rollover, this termination is the salt-sediment onlap contact (Fig. 11). The maximum dip on the seaward-facing limb of the rollover is measured (numbered white sticks on Fig. 11), not the angular discordance between overburden and salt, or the angle of overburden downlap/onlap immediately adjacent to the salt-sediment interface. The angle of downlap/onlap depends on the steepness of the salt face at the sediment surface, which in turn depends on the topographic relief of the diapir at the downdip end of the expulsion rollover. This relief probably varies through time as a function of sedimentation rate and salt supply.

Ge et al. (1997) argued that this methodology provides an objective and seemingly robust technique to differentiate between extension-
related and expulsion-related rollovers. Using this technique, their investigation of the Albian Gap rollover suggested that it was largely an expulsion-related structure.

We expand the methodology of Ge et al. (1997) by compiling many more rollover plots for: (1) natural examples (e.g., Fig. 8) and scaled physical models of listric fault–related rollovers; and (2) natural examples (e.g., Figs. 6C–6E), and physical (e.g., Fig. 6A) and numerical (e.g., Fig. 6B) models of expulsion-related rollovers; our analysis draws on examples published since the study of Ge et al. (1997). The extension-related rollovers that we analyzed—irrespective of whether they are natural or model examples—have maximum dips that gradually increase from typically <5° to an average of ~60° (maximum of 80°; Fig. 12A; cf. inset graph in Fig. 11A). This dip increase can be relatively regular and roughly linear, or the increase can be relatively abrupt before becoming gentler. Irrespective of the shape of the curve, maximum stratal dip corresponds to flattening of the underlying listric fault. In contrast, plots for natural and model examples of expulsion-related rollovers are strikingly different from those associated with extension. Rather than having a curve that progressively increases to a maximum dip (e.g., Fig. 11A), the expulsion-related plots are strongly asymmetric, steeply increasing to a maximum before gently decreasing to a value that is not much larger than that immediately adjacent to the salt-sediment contact (Fig. 12B; cf. inset graph in Fig. 11B). Our analysis of rollover plots for unequivocally extension-related structures and those related to expulsion clearly indicates that both types of rollover, despite being superficially similar, are fundamentally different. The Albian Gap has a strongly asymmetric rollover plot that indicates an origin as an expulsion rollover.

**Unique Scale and Attitude of the Cabo Frio Fault**

In the extension model, the salt-sediment contact at the southeastern edge of the Albian Gap is interpreted as a landward-dipping, listric normal fault that has several kilometers of throw (Fig. 5A; Mohriak et al., 1995; Szatmari et al., 1996; Quirk et al., 2012; Rowan and Ratliﬀ, 2012). However, to the best of our knowledge, thin-skinned, landward-dipping extensional structures of this scale have never been documented on any other salt-bearing passive margin. Seaward-dipping faults are more common and typically larger than landward-dipping faults in basins dominated by thin-skinned gravity-driven tectonics (Fig. 7; Jackson and Cramez, 1989; Duval et al., 1992; Lundin, 1992; Anderson et al., 2000; Rouby et al., 2002; Hudec and Jackson, 2004; Cartwright and Jackson, 2008). Furthermore, the following four aspects of the Cabo Frio fault have never been explained:
Figure 12. Rollover plots illustrating the geometric variability between natural and modeled (physical and numerical) rollovers generated by (A) extension and (B) expulsion. To compare different size structures, the plots have been normalized for distance (x-axis) and dip (y-axis). Sources of the data are shown in the key. Examples in B labeled with * are derived from time-migrated sections from the Santos Basin, depth-converted using velocities presented by Quirk et al. (2012, their fig. 14). LD—landward-dipping faults; SD—seaward-dipping faults. Examples from Guerra and Underhill (2012, their fig. 4) and Ge et al. (1997, their figs. 16A and 16B) are from the Albian Gap. Compare these data to idealized plots shown on inset graphs in Figure 11.
(1) why each segment should dip anomalously landward along the entire 200 km length of the fault system (Fig. 2A); (2) why it has no equivalent seaward-dipping fault along strike (Fig. 2A); (3) why an extensional zone of such great width should be dominated by, in effect, a single fault system and associated weld (Fig. 2A); and (4) why some seismic profiles across the Albian Gap show the rollover ending seaward in a salt wall, rather than in a fault (Fig. 3B; see fig. 14 in Mohriak et al., 1995). In contrast, the scale and uniform landward dip of the salt-sediment contact at the southeastern edge of the Albian Gap are inherent to the expulsion model (Figs. 6A and 6B).

**Negative Evidence Compatible with Post-Albian Extension and Post-Albian Expulsion**

**Lack of a Prekinematic Layer Capping the Salt**

Seismic data suggest that seaward-dipping Upper Cretaceous strata directly overlie sub-salt strata across the Albian Gap and that a prekinematic layer is absent (Fig. 3A). Quirk et al. (2012) argued that the absence of such a layer casts doubt on the physical (e.g., Fig. 6A; Ge et al., 1997) and numerical (Gemmer et al., 2005) models of salt expulsion. However, the thin prekinematic layer built into these models was merely a strain marker to track variations in extension and shortening and has negligible mechanical significance (see also Albertz and Ings, 2012). Expulsion is driven by synkinematic strata that prograde across the salt, regardless of whether or not a thin prekinematic layer is present.

In the extension model, there is no prekinematic layer because it was pulled apart by post-Albian extension. In the expulsion model, there is no prekinematic layer because the Albian Gap was occupied by a salt wall during Albian time. The absence of the prekinematic layer is therefore not diagnostic.

**Lack of Strike-Slip Faults at the Ends of the Albian Gap and Cabo Frio Fault**

Fiduk and Rowan (2012) suggested that the expulsion model requires high strain gradients at the north and south ends of the Albian Gap because nonextending strata within the gap are juxtaposed against extending units to the north and south. Such high strain gradients, they argued, should have produced strike-slip faults to separate the static Albian Gap from more active adjacent domains. They concluded that the absence of such strike-slip faults indicates that the Albian Gap cannot have formed due to expulsion. The lack of strike-slip faults is not diagnostic of either extension or expulsion, based on the following two arguments. First, the expulsion model suggests that during post-Albian times, there was only limited extension in the Albian Gap. Post-Albian extension is also relatively minor to the northeast and southwest of the Albian Gap, so there is no reason to call on a large lateral strain gradient during post-Albian deformation, and thus no reason to require post-Albian strike-slip faulting. Second, our expulsion model allows for large magnitudes of Albian extension to open the Albian Gap in the first place. This is consistent with major Albian extension to the north and south, again eliminating the need for a large lateral strain gradient and strike-slip faults. Thus, salt expulsion would not need to create strike-slip faults at the lateral boundaries of the Albian Gap, so the absence of such structures is consistent with both models.

**Lack of Large Normal Faulting in the Upper Cretaceous Rollover**

The Upper Cretaceous rollover above the Albian Gap lacks large normal faults (Fig. 3A). Quirk et al. (2012) argued that this lack of faults can be explained by rapid seaward flow of salt during the late Albian to middle Santonian, which meant that the Albian Gap and the overlying deep-water basin formed relatively quickly and before progradation of the Upper Cretaceous shelf margin; extension of and faulting within the clinoforms are thus not kinematically required.

We suggest that a lack of normal faults in the Upper Cretaceous rollover is not diagnostic of either the extension or the expulsion model. Although normal faults are extremely common in extensional rollovers (Figs. 8B and 8C; e.g., Christiensen, 1983; Laberg, 1983; Diegel et al., 1995; Rowan et al., 1999; Brun and Mauduit, 2009), these faults need not be large enough to be seismically resolvable nor pervasive throughout the rollover. Furthermore, because overburden does not extend during progradation and salt expulsion, natural expulsion rollovers (Figs. 6C–6E; Diegel et al., 1995; Schuster, 1995; Trudgill, 2010) and modeled expulsion rollovers (Fig. 6A; Ge et al., 1997) tend to be weakly faulted, except for minor faults related to outer-arc stretching near the crest of the rollover (Ge et al., 1997; Rowan et al., 1999). Because published examples of natural and modeled expulsion rollovers tend to have fewer faults than published examples of extensional rollovers, the lack of faulting in the Albian Gap points to an expulsion origin for the Upper Cretaceous rollover. However, the lack of faulting is not sufficiently diagnostic to be a persuasive point.

**Negative Evidence Compatible with Post-Albian Extension but Not Post-Albian Expulsion**

As we discussed earlier herein and will expand on in the following, Albian extension is part of both the extension and expulsion models. However, we see no diagnostic evidence for post-Albian extension.

**Negative Evidence Compatible with Post-Albian Expulsion but Not Post-Albian Extension**

We see no negative evidence that is compatible with post-Albian expulsion but not post-Albian extension.

**ALTERNATIVE MODEL FOR THE ORIGIN OF THE ALBIAN GAP**

At the start of this paper, we outlined two key models of thin-skinned, gravity-driven structural systems on salt-bearing passive margins: extension and expulsion. We proposed key geological observations to support or oppose each model. We then critiqued arguments in support of either an extension or expulsion origin of the Albian Gap. Our assessment is that much of the evidence put forward in support of the post-Albian extension model: (1) is flawed (e.g., that intra-salt shortening can be used to balance suprasalt extension); (2) cannot account for specific geological observations (e.g., the unique scale and dip direction of the supposed Cabo Frio fault); and (3) is not unique to the extension model and also supports an expulsion origin for the structure (e.g., thin salt in the Albian Gap, or the regional balance between extension and shortening). Furthermore, several lines of evidence favor the expulsion model over the extension model, particularly our qualitative (seismic-stratigraphic) and quantitative analyses of the geometry of the suprasalt rollover that overlies the Albian Gap. Based on the preceding summary, we propose the following revised model for the origin of the Albian Gap. This hybrid, two-stage model involves Albian extension followed by post-Albian expulsion (Fig. 13).

**Stage 1: Tilting, Extension, and Reactive Diapirism**

During the initial stages of seaward tilting of the continental margin in the late Albian, extension was accommodated by predominantly seaward-dipping, listric normal faults that defined extensional Albian rafts, and by opening and widening of a major salt wall that would later become the Albian Gap. The interpretation that
Figure 13. Three-stage synoptic model illustrating our hypothesis for the formation of the Albian Gap: (A) Stage 1—end–early Albian; (B) Stage 2—end-Albian; and (C) Stage 3—Late Cretaceous–present. This model incorporates both thin-skinned extension (Figs. 1A and 1B) and progradation-related salt expulsion (Fig. 1D). Key to stratigraphic units is shown in Figure 4.
this broad, low salt wall formed during the late Albian is consistent with observations that, along much of the length of the Albian Gap, the Albian to Cenomanian thickness thins southward toward its apparent downlap onto the pre-Albian salt related salt depocenter as an oblique rifted margin, Journal of Geophysical Research: Solid Earth, v. 117, B08103.


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