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STRUCTURAL DYNAMICS OF SALT SYSTEMS

Martin P. A. Jackson, Bruno C. Vendeville, and
Daniel D. Schultz-Ela

Bureau of Economic Geology, The University of Texas at Austin, Austin, Texas 78713

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INTRODUCTION

The structural dynamics of salt systems—salt tectonics—encompasses any deformation involving salt or other evaporites. It includes halokinesis, in which deformation is driven primarily by salt upwelling and withdrawal, not by regional tectonics. A salt-tectonic system is composed of a source layer of rock salt or other evaporites (collectively referred to as "salt"), overlain by sedimentary overburden, and overlying a basement or subsalt strata. The salt acts as a lubricant, aiding decoupling of the overburden, and accommodates the potential gaps and overlaps between shifting fault blocks in the deforming overburden. This review focuses on the structural dynamics of such salt systems, with an emphasis on diapir formation in both extensional and contractional settings. We also consider subsiding diapirs, salt welds, and allochthonous salt sheets.

BASIC INTERACTIONS

Because the weight of overlying sediments tends to expel salt upward, salt tectonics is largely confined to shallow crust above the ductile-brittle transition zone some 8–15 km deep. In this brittle domain, the overburden deforms not by creep flow—as has been widely assumed by modelers of salt diapirism—but by frictional slip along faults or penetrative slip surfaces (e.g. see compilations in Brace & Kohlstedt 1980 and Weijermars et

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al 1993). In marked contrast, salt typically deforms as a viscous or power-law fluid with viscosity typically ranging between 10^{17} and 10^{19} Pa s, depending mainly on grain size and water content (van Keken et al 1993). Salt contains shear zones, but faults within it form only during rapid strain rates accompanying seismic or certain igneous intrusive events.

Except in salt glaciers, dry salt is constrained to flow slowly. However, unconfined, damp salt can flow unusually fast (Table 1, Talbot & Rogers 1980, Talbot & Jarvis 1984). Because of the low viscosity of damp salt, processes controlling salt mechanics mostly depend on (*a*) pressure (pressure in salt, friction in overburden rocks) rather than time, or (*b*) parameters independent of salt rheology (e.g. rates of regional shortening, extension, and sediment progradation or aggradation). These multiple parameters can be simply combined into three ratios: V/B, P/B, and P/V, where V, P, and B are stresses due to salt viscosity, salt pressure, and overburden brittle strength, respectively (Vendeville & Jackson 1993).

V/B reflects the relative strengths of salt and overburden. A system containing a very thin overburden or that is deformed rapidly has a high V/B. The salt layer thins uniformly and stretches the pervasively faulted overburden skin. Viscous forces also increase markedly if the source layer is greatly thinned by extension or salt withdrawal. In contrast, a system comprising a thick overburden above moderately thick, slowly deformed salt has a low V/B because of salt's low viscosity. Here, salt tends not to stretch its overburden, which is decoupled from its basement. The overburden layer faults and tilts above the lubricating salt.

P/B determines whether a diapir can autonomously pierce its overburden. Salt flow driven by pressure gradients can overcome the resisting brittle strength of the overburden (high P/B) if the diapir roof does not exceed a critical thickness (see *Active Piercement*).

Table 1 Strain rates and velocities in salt compared with other tectonic systems[a]

Type of flow	Strain rate s^{-1}	Velocity mm a^{-1}	Velocity
Lava flow	10^{-5} to 10^{-4}	5×10^{11} to 3×10^{13}	1 to 60 km hr^{-1}
Ice glacier	10^{-10} to 5×10^{-8}	3×10^{5} to 2×10^{7}	1 to 60 m day^{-1}
Salt glacier	1×10^{-11} to 2×10^{-9}	2×10^{3} to 2×10^{6}	10 to 100 km Ma^{-1}
Mantle currents	10^{-14} to 10^{-15}	10 to 1×10^{3}	2 m a^{-1} to 5 m day^{-1}
Salt tongue spreading (<30 km wide)	8×10^{-15} to 1×10^{-11}	2 to 20	2 to 20 km Ma^{-1}
Salt tongue spreading (>30 km wide)	3×10^{-16} to 1×10^{-15}	0.5 to 3	0.5 to 3 km Ma^{-1}
Salt diapir rise	2×10^{-16} to 8×10^{-11}	1×10^{-2} to 2	10 m to 2 km Ma^{-1}

[a] From Jackson & Talbot 1991.

P/V controls the rise rate of passive diapirs emergent at the sea floor or land surface (see *Passive Piercement*). Driving pressure is proportional to diapir height and to the salt-overburden density ratio. Viscous forces are proportional to salt viscosity and inversely proportional to diapir width and source-layer thickness. A high P/V (high density contrast, low salt viscosity, wide diapir, or thick source layer and overburden) favors rapid salt upwelling. Conversely, low P/V increases viscous drag and retards diapir rise.

Regional extension can initiate reactive diapirs (see *Reactive Diapirism*). Their rise rate is controlled by the regional extension rate. Early in the reactive stage the brittle strength of the graben floor still exceeds the underlying diapiric pressure (low, but gradually increasing P/B) and prevents diapiric breakout. Further extensional thinning of the graben floor increases P/B, eventually allowing the diapir to actively pierce and perhaps emerge at the surface.

Diapir subsidence rate depends inversely on P/V (see SUBSIDING DIAPIRS). Thinning of the source layer by extension and withdrawal lowers P/V. Rapid widening of diapirs during extension also lowers P/V.

Most of the concepts summarized here originated or were tested by modeling. Weijermars et al (1993) reviewed the history, methodology, and dynamic scaling laws of modeling salt tectonics.

MODES OF DIAPIRIC PIERCEMENT

Salt diapirs have discordant contacts with overlying strata, as opposed to concordant structures such as salt pillows and salt-cored anticlines. To become discordantly encased in the overburden, a diapir faces a "room problem." Three modes of piercement solve the room problem: active, passive, and reactive diapirism (Figure 1).

Early in this century, different researchers argued whether diapirs rose because of density inversion, crystallization forces, or lateral contraction. Regardless of the elusive driving force, most agreed that the diapir was forcefully intruded by the process now called *active piercement* (Nelson 1991). Such a diapir forces its roof upward and sideways, thereby solving the room problem (Figure 1, *middle panel*).

This dogma was shattered by Barton's (1933) concept of a downbuilding or *passive diapir* (Nelson 1991, Jackson & Talbot 1991) that remains at or near the surface while sediments accumulate discordantly around it. The diapir grows taller because its base sinks relative to the surface. No overburden is displaced, so there is no room problem (Figure 1, *bottom panel*).

A third mechanism of piercement is *reactive diapirism* (Vendeville & Jackson 1992a). Unlike the other two modes, reactive piercement requires

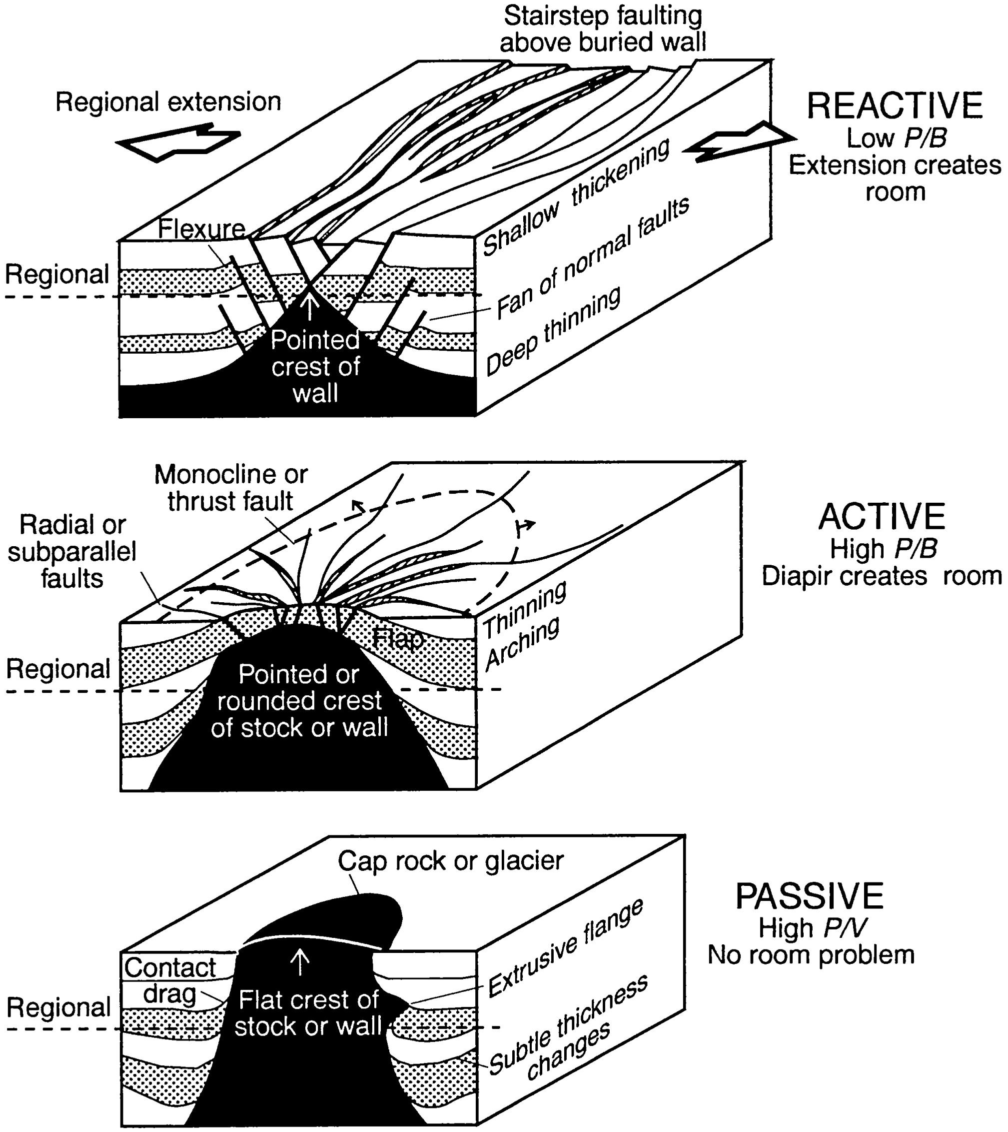


Figure 1 Three piercement modes for salt diapirs (*black*) and their characteristic structures. Regional datum (*dashed line*) is base of upper stippled layer. P, V, and B refer to stresses due to salt pressure, salt viscosity, and overburden brittle strength, respectively.

regional extension caused by rifting or by thin-skinned gravity spreading or gliding. Regional extensional faulting creates space for this type of diapir (Figure 1, *top panel*).

Initiation of Diapirism

Initiation of diapirs would pose no problem if their overburden were a fluid lacking yield strength. Diapirs would rise from any subtle bulge of salt, such as that created by facies changes. The only possible impediment would be a geologically insignificant rise rate (Jackson & Talbot 1986). But the finite strength of brittle sedimentary rocks impedes penetration driven by the weak differential pressures in a subtle bulge of salt. Without regional extension, a realistic brittle overburden can only be pierced if it is thin and flanked by much thicker sediments. This sedimentary differential loading could create enough pressure at the crest of a bulging pillow for active breakthrough and emergence.

Until the 1990s the role of thin-skinned extension in initiating and promoting diapiric rise was generally ignored. Grabens associated with diapirs were attributed to intrusion, withdrawal, or dissolution of salt diapirs. Regional faults lacing around the diapirs were regarded as resulting from, or superposed on, an independent mechanism of diapiric rise. This view became suspect when the first computerized structural restorations from the Gulf of Mexico were published (Worrall & Snelson 1989).

Physical modeling by Vendeville & Jackson (1992a) showed that regional extension can initiate and promote diapirism regardless of the thickness, lithology, and density of the overburden. The overburden necks to form grabens or half grabens, and pressurized salt rises as diapiric walls into these structurally-thinned zones. Faulting and loading of the overburden drive salt flow rather than vice versa. Tectonic differential loading by extension initiates salt diapirism more effectively than does sedimentary differential loading because extension (*a*) pervades rifts and divergent margins where salt typically accumulates, (*b*) weakens the overburden by fracturing and thinning, and (*c*) by creating open-cast structures, differentially loads salt much more effectively than do sedimentary facies changes, which are more gradual and of lower density contrast. Regional extension provides the only means for tabular salt to pierce thick or thin overburdens of initially uniform thickness. A survey of 18 of the world's salt basins documents a close, consistent temporal link between the onset of diapirism and regional extension (Jackson & Vendeville 1994). Extending salt basins typically develop salt structures, whereas nonextending basins typically do not. In some basins containing thick salt (e.g. SW Iran), diapirism was delayed as long as 400 Ma until the basin extended regionally. In other salt provinces (e.g. French Maritime Alps), episodic growth of salt diapirs correlates with episodic regional extension. Once initiated, salt diapirism can continue if contraction or quiescence follows regional extension. Thus, even in salt basins overprinted by inversion or orogenic contraction (e.g. Morocco, Lusitania, Basque-Cantabrian, North Sea), diapirs were initiated during extension on divergent continental margins or in intracontinental rifts.

Reactive Diapirism

A reactive diapir progressively pierces an initial half graben or graben. As extension proceeds, the fault block slides into the source layer until supported by the pressure of salt and by flexural resistance of the block. Continued extension generates new faults from older fracture zones that activate successively inward above the diapir crest. Each new fault slices smaller pieces off the diminishing graben floor. Faulted strata above the diapir crest diverge over time down the flanks of the rising diapir. These supercrestal strata lie below regional datum (Figure 1), which is the reference line connecting undeformed (apart from compaction) points on a horizon (see Figure 6). Strata drop below regional datum during extension, reactive diapirism, salt withdrawl, or salt dissolution; strata rise above regional datum during contraction or active diapirism.

Extension thins the overburden and promotes underlying diapirism, whereas sediment accumulation in the graben thickens the overburden and retards diapirism. The ratio between the rates of sediment aggradation, $\dot{A}$, and regional extension, $\dot{\varepsilon}$, controls structural style. Rapid extension and slow deposition (low $\dot{A}/\dot{\varepsilon}$) promote reactive diapir rise by progressively thinning the diapir roof, thereby increasing P/B. Conversely, slow extension and rapid deposition opposes diapirism by thickening the overburden and decreasing P/V, leading to normal growth faulting and salt rollers—sharp-crested low diapirs (Vendeville & Jackson 1993). The rate of regional extension controls the rate of reactive diapirism. Extension is typically much slower than unconstrained salt flow, so deformation is virtually independent of salt viscosity. Whenever regional extension ceases, reactive diapirs stop growing.

Active Piercement

If a diapir becomes sufficiently tall and its roof sufficiently thin, pressure at the diapiric crest impels it to the active stage. The diapir lifts its roof above regional datum, then rotates, breaks through, shoulders aside, and disperses its remains while dragging up adjoining strata (Figure 1). Driven by diapir pressure, an active diapir forcefully intrudes its overburden (Schultz-Ela et al 1993). During reactive piercement, in contrast, the strength of the overburden is overcome by the tangential forces driving regional extension.

Active diapirism is enhanced by: (*a*) high density contrast between overburden and salt, (*b*) weak overburden, (*c*) large diapiric height and width relative to regional overburden thickness, (*d*) elongated rather than circular planform, and (*e*) topographic relief rather than salt relief. For a density ratio of 1.1, and a height/width diapiric ratio of 1, a salt wall must be encased by roughly three-quarters of the overburden thickness before it can intrude actively. Active diapirism becomes progressively more difficult for diapiric crests whose cross-sectional shape range from a rectangle, a round-cornered rectangle, a semicircle, to an upright triangle.

Modeling suggests the following evolution for active piercement (Vendeville & Jackson 1992a, Schultz-Ela et al 1993): The overburden roof begins to arch upward as the diapir transmutes from reactive to active. Forceful intrusion extends the roof's outer arc to produce a central graben flanked by upward and outward rotating flaps. At their outer ends the flaps hinge by (*a*) bending, which is enhanced by layer-parallel slip, or (*b*) either inward-dipping reverse or outward-dipping normal faults. Reverse faults curve outward as they propagate upward from an initially near-vertical initiation. Meanwhile, normal faults in the central graben propagate downward with new faults created outward (in contrast, reactive faults propagate inward). Antithetic fault slip maintains a flat crest on the graben. Thin, wide roofs develop paired grabens separating an undeformed flat roof from adjoining wedge-shaped flaps. Active diapirs are surrounded by radial faults if regional extension has ceased (Figure 1) or by subparallel normal faults if extension accompanies piercement (Withjack & Scheiner 1982).

As the diapir nears emergence, its pointed crest rounds and widens. Erosional thinning and weakening of the roof stimulate further active rise, while adjoining redeposition increases the pressure for salt breakout. The displaced flaps oversteepen and slump onto the flanking overburden as chaotically resedimented, recumbently folded, imbricately thrusted strata. Finally, the flaps overturn or are entrained by outward salt flow or are removed by erosion. The diapir then emerges as a narrow crest then with its full width. The diapir can also propagate along strike of the graben that initiated it reactively. A density inversion promotes emergence but is not essential if the graben floor is below the pressure head of the source layer.

The overburden load on the source layer can pump salt from the diapiric vent to form an exposed bulge of salt ranging from a low mound (Jackson et al 1990) to a mountain 1 km high (Talbot & Jarvis 1984). The cross-sectional shape of the salt bulge and whether it extrudes glacially depend on the relative rates of upward salt flow and dissolution of its crest.

Passive Piercement

A diapir becomes passive when it emerges (Figures 1, 2). Strata accumulate around the diapir and thin over the crest. A thin roof has little mechanical influence on a passive diapir and is periodically destroyed by active breakout. This lack of vertical constraint has two important implications. First, there is no room problem because flanking strata are not displaced. Faulting and folding are thus negligible around passive diapirs, apart from a narrow drag zone along their contacts (Figure 1). Thickness changes in synkinematic strata (those deposited during the flow of underlying salt) may not be apparent if the overburden resists flexure, in which case salt is withdrawn from wide surroundings. Peripheral sinks are deeper if the overburden can flex and the diapir taps mainly nearby salt. Second, accumulating sediments directly contact salt. Changes in sedimentation rate can profoundly influence diapiric shape (Vendeville & Jackson 1993). The balance between rates of (*a*) aggradation and (*b*) net diapiric rise (gross rise diminished by dissolution and erosion) determines the cross-sectional shape of a passive diapir. If the sediment aggradation rate equals the diapiric rise rate, the diapiric margins grow vertically. However, if the diapir rises faster than the surrounding sediment aggrades, the laterally unsupported salt repeatedly overflows its margins, thus widening upward with time. Conversely, if aggradation is relatively fast, onlapping strata encroach the diapir, which thus narrows upward with time.

Sedimentation rate also controls the planform of a passive diapir. A diapir's flux increases with increasing width. At the diapir's widest part, salt flow rates are high relative to aggradation rates (Figure 2). Conversely, contact drag affects the narrow ends of an elliptical diapir more than wider parts, creating *relatively* high aggradation rates. Thus, salt spreads at the widest part of the diapir and retreats from its elliptical ends. Accordingly, the originally-elongated salt wall evolves toward a string of pluglike passive stocks rising from the deeply buried reactive ridge. The number of exposed plugs declines as they exhaust their supplies and become buried (Lin 1992).

The gross flow rate of passive diapirs reflects the balance between driving

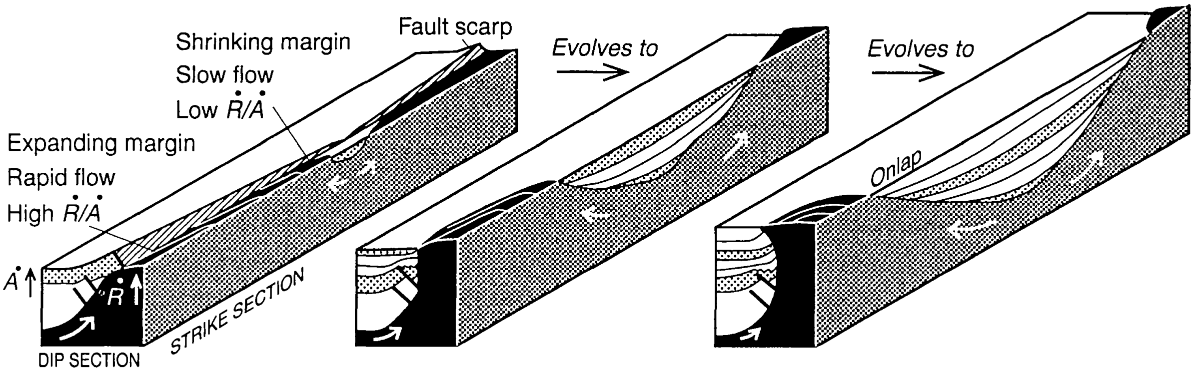


Figure 2 Newly emergent salt wall (*left*, transforming from active to passive) evolves to separate passive stocks (*right*) by along-strike flow of salt. Stocks become more rounded as they spread at their wider, more-vigorously flowing zones and shrink at their narrower, more sluggish zones as strata onlap across them. $\dot{A}$ is local aggradation rate, and $\dot{R}$ is net rise rate of diapiric salt (qualitatively contoured in white on diapir crest).

forces, retarding forces, and salt supply (Vendeville et al 1993b). The flow rate of salt increases with diapir width and also changes over time (Figure 3). As the passive diapir initially grows taller and the overburden thickens, increased pressurization of the source layer increases the flow rate proportionally to overburden thickness. Later, the flow rate is controlled by source-layer thickness because as this layer is depleted, resistance to flow climbs steeply in inverse proportion to the cube of its thickness. Rise rates in nature are much lower than the scaled peak experimental rates, suggesting that either (*a*) much salt must dissolve from passive diapirs or (*b*) most passive diapirs would rapidly extrude to form salt sheets unless dissolved. The long phase of flow deceleration results in low flow rates like those in nature (Figure 3).

Superposing constant sedimentation rate on this change in salt flow rate creates a characteristic diapiric shape (Figure 3, *bottom panel*). Its base narrows upward as salt initially flows slowly and sediments encroach; the diapir then widens upward as salt flow accelerates and diapiric overhangs spread over the sediments. Finally, deceleration of salt flow from the depleting source layer tapers, then completely buries the diapir's shoulders. This evolution produces a diapiric stem without necking or otherwise deforming the overburden. Salt flows along the diapiric wall from below depressions into intervening culminations that evolve into increasingly pluglike, rounded stocks (Figure 2).

Evolutionary Paths of Diapirs

During constant moderate aggradation, a salt diapir would evolve from reactive to active to passive. This progression can be changed by either depletion of source layer or variations in regional extension or aggradation rates. Various evolutionary paths are possible:

- A diapir could appear to be initiated passively. If the overburden is thin and unevenly deposited, the reactive and active stages of growth are brief. Strata faulted during these early stages could be too thin to be seismically visible.
- A diapir could appear to have been continuously passive, although it actually passed through countless cycles. Each cycle could comprise an episode of passive growth during slow aggradation, followed by brief burial during a pulse of high aggradation, and terminated by brief active breakout. Each active stage destroys the domed sedimentary veneer by local extension, entrainment, slumping, erosion, and dissolution collapse.
- An active diapir can partially break through its roof but never emerge, for any of three reasons. First, the source layer may become depleted (Figure 3, *right*). Second, piracy of salt by an emergent segment of a wall

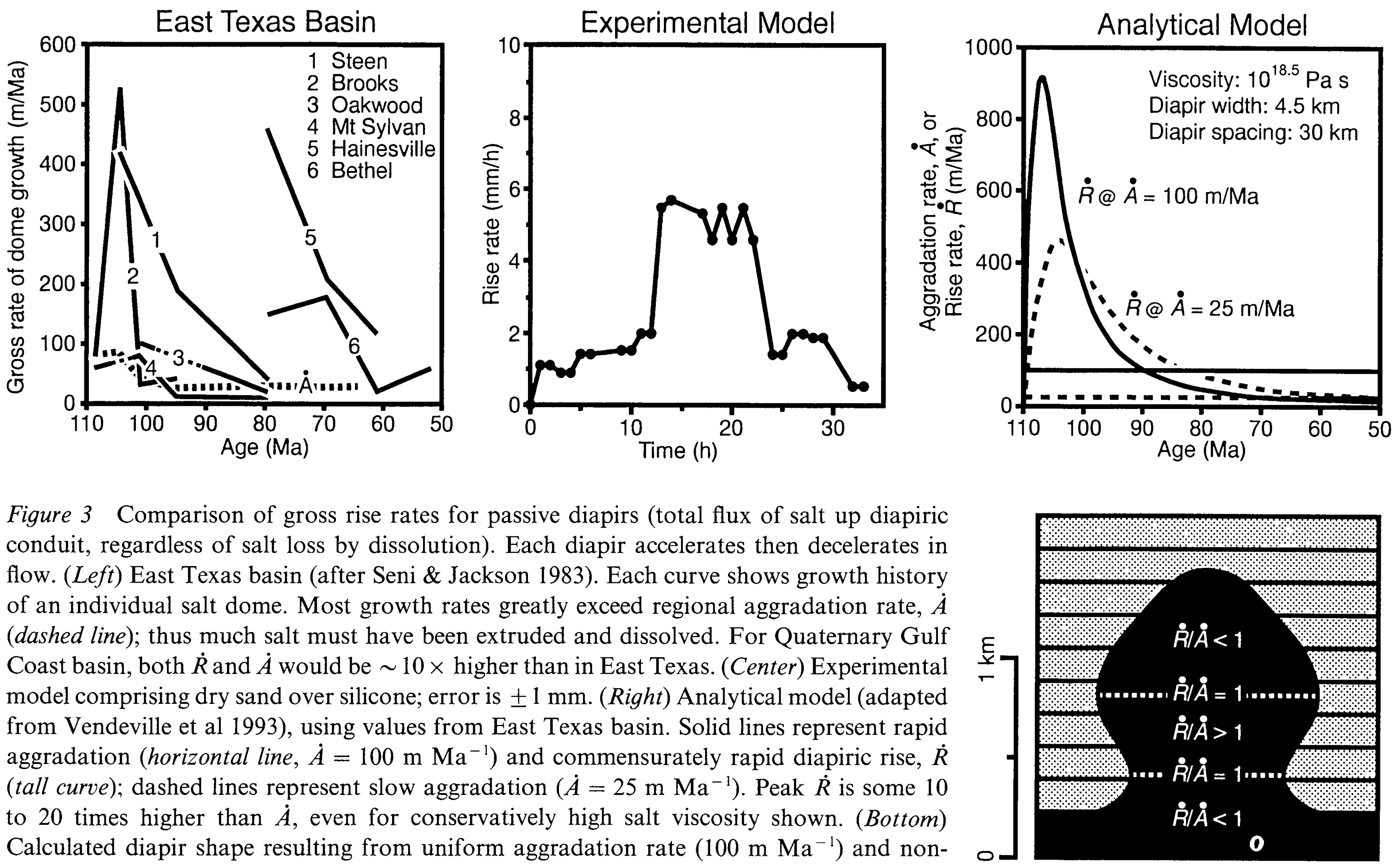


Figure 3 Comparison of gross rise rates for passive diapirs (total flux of salt up diapiric conduit, regardless of salt loss by dissolution). Each diapir accelerates then decelerates in flow. (*Left*) East Texas basin (after Seni & Jackson 1983). Each curve shows growth history of an individual salt dome. Most growth rates greatly exceed regional aggradation rate, $\dot{A}$ (*dashed line*); thus much salt must have been extruded and dissolved. For Quaternary Gulf Coast basin, both $\dot{R}$ and $\dot{A}$ would be ~10 × higher than in East Texas. (*Center*) Experimental model comprising dry sand over silicone; error is ±1 mm. (*Right*) Analytical model (adapted from Vendeville et al 1993), using values from East Texas basin. Solid lines represent rapid aggradation (*horizontal line*, $\dot{A} = 100$ m Ma^{-1}) and commensurately rapid diapiric rise, $\dot{R}$ (*tall curve*); dashed lines represent slow aggradation ($\dot{A} = 25$ m Ma^{-1}). Peak $\dot{R}$ is some 10 to 20 times higher than $\dot{A}$, even for conservatively high salt viscosity shown. (*Bottom*) Calculated diapir shape resulting from uniform aggradation rate (100 m Ma^{-1}) and non-uniform salt flow, derived from analytical model (*right*); width of diapir is unscaled. Dissolution would truncate pointed crest of diapir.

can halt growth of other adjoining sections (Figure 2). Third, numerical modeling suggests that without erosion or deposition, an active diapir only partially penetrates all but the thinnest roofs before reaching equilibrium (Schultz-Ela et al 1993).

THIN-SKINNED EXTENSIONAL SALT TECTONICS

Evaporites accumulate almost exclusively in extensional regimes such as divergent continental margins and rifts. Thin-skinned extension involves stretching of the cover decoupled by a basal layer of salt or other evaporites from an essentially undeformed basement.

An unambiguously extensional structure is a salt roller—a sharp-crested low diapir. One flank is concordant with the overburden, whereas the other discordant flank is the basal segment of a listric growth fault. Structural relief and width of a roller gradually increase over time, as shown by physical modeling (Vendeville & Cobbold 1987, Cobbold et al 1989, Vendeville & Jackson 1992a,b). Palinspastic reconstructions (Worrall & Snelson 1989, Wu et al 1989, Duval et al 1992, Schultz-Ela 1992) and this modeling favor regional extension triggering salt upwelling rather than the upwelling causing the faulting.

Aggradation rates vary in time and space. For example, Paleogene accumulation rates on the stable shelf of the Gulf of Mexico varied by a factor of ten, largely due to a pulsating supply of sediment (Galloway & Williams 1991). Accumulation rates could vary as much as 100-fold on the adjoining slope, where most salt structures are initiated, because structures create accommodation space and concentrate sediment pathways.

The influence of aggradation rate on the structural style of extensional salt tectonics was first noted by Vendeville & Cobbold (1987) and systematically modeled by Lin (1992), who simulated a tenfold variation in aggradation rate during regional extension. Markedly different structural styles resulted (Figure 4). High $\dot{A}/\dot{\varepsilon}$ (rapid aggradation, slow extension) favors major listric growth faults rather than diapiric growth. Initially, grabens pierced by symmetric, buried diapiric walls are formed. As the overburden thickens, only the downslope-dipping boundary fault of each graben remains active, evolving into a listric growth fault with a deeply buried, low diapiric roller in its footwall. In contrast, low $\dot{A}/\dot{\varepsilon}$ promotes salt diapirism. Under slow aggradation, triangular salt walls penetrate the initial grabens and evolve into tall walls having steep or overhanging, discordant contacts. Their crests are typically shallowly buried or emergent, in which case they become flat-topped and pluglike. In the extensional spectrum, these tall stocks are polar to salt rollers (Figure 4). Intermediate aggradation rates resulted in intermediate diapiric structures (moderately buried, large triangular salt walls).

Where salt walls penetrate only the deep overburden, the deep cover extends mainly by widening of salt walls, whereas the shallow cover extends by faulting in the overburden; extension is partly cryptic (Figure 5, *left*). Where passive salt walls penetrate the entire overburden, the section can extend almost entirely by widening of walls between separating glide blocks of overburden. Lack of visible faults makes even large extensions subtle or cryptic (Figure 5, *right*) (Vendeville & Jackson 1992a).

Walls and faults are typically parallel to each other and perpendicular to the dip of the basin margin, as in the Kwanza basin of Angola (Duval et al 1992). On an irregular margin, downslope extension can be convergent

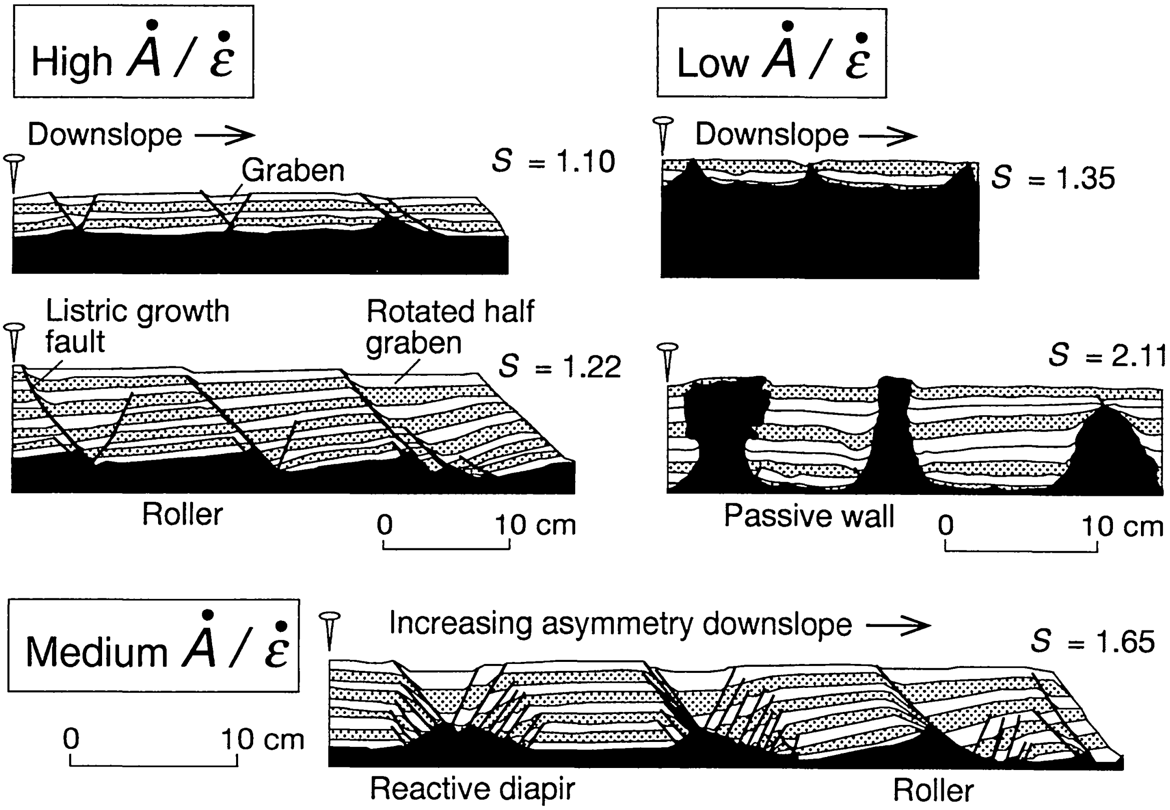


Figure 4 Ratio of aggradation rate to extension rate ($\dot{A}/\dot{\varepsilon}$) controls structural style of experimental diapirs during extension. (*Upper left*) High $\dot{A}/\dot{\varepsilon}$ (34 cm) results in deeply buried, low salt rollers bounded by lower parts of listric growth faults and overlain by rotated half grabens. (*Upper right*) Low $\dot{A}/\dot{\varepsilon}$ (4.5 cm) results in shallowly buried, tall stocks and walls and negligible faults. (*Bottom*) Medium $\dot{A}/\dot{\varepsilon}$ (6.4 cm) results in moderately tall buried diapiric walls whose triangular shape becomes increasingly asymmetric toward the right, where the overburden starts wedging out. S = finite stretch. (After Lin 1992, Vendeville & Jackson 1992a.)

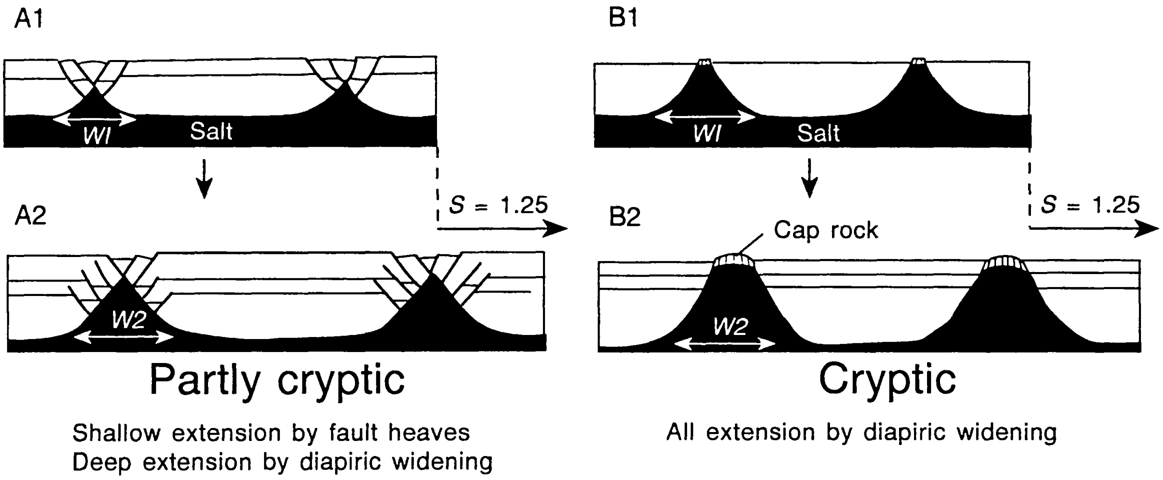


Figure 5 Partly cryptic extension (*left*) and cryptic extension (*right*) for same finite stretch *S*. W = width of diapiric base.

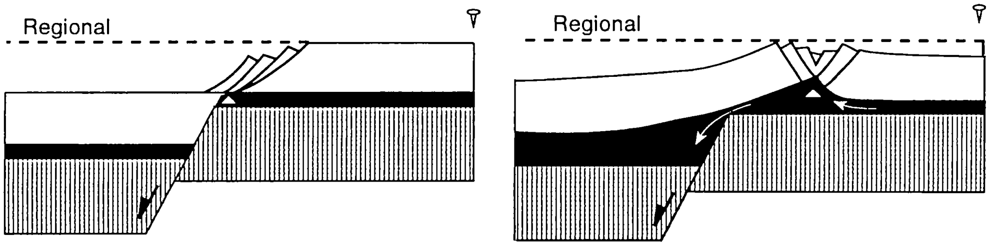
over indentations or divergent over salients, as on the Campos basin of Brazil (Cobbold & Szatmari 1991, Rouby et al 1993).

THICK-SKINNED EXTENSIONAL SALT TECTONICS

Thick-skinned extension stretches both basement and cover. Even trivial normal faults offsetting the base of salt have long been thought to initiate diapirs and thus to influence their location and shape. However, modeling indicates that the effect of basement faults can propagate upward through the salt into the overburden only if lateral salt flow is hindered by unusually high salt viscosity, very rapid regional extension, or extreme thinness of the salt (Figure 6*a*) (Vendeville et al 1993a).

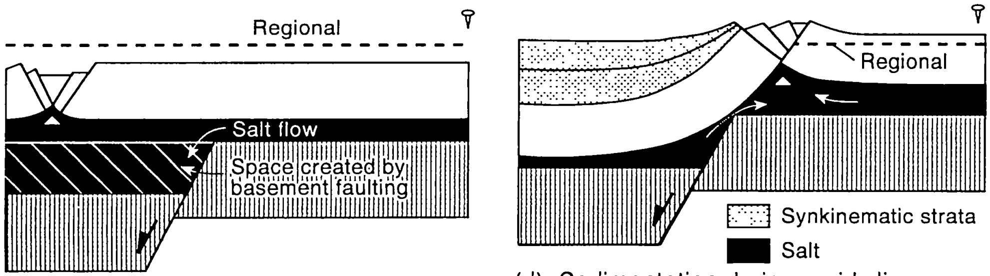
Where the basement fault slips at moderate rates or the salt is moderately thin (Figure 6*b*), the cover over the downthrown block sags, forming a monocline above the basement fault. Local stretching of the upper monoclinal hinge is accentuated by regional extension to form a graben above the upthrown basement block next to (not above) the basement fault. This thinning triggers reactive diapiric piercement into the faulted graben.

During slow extension, a thick (say >500 m) salt layer effectively decouples its brittle overburden from a faulting basement (Figure 6*c*). Salt flows from the upthrown block toward the downthrown block, allowing the overburden to subside uniformly across the basement fault. Extension widens the salt basin and stretches the cover. Faults and walls in the overburden form subparallel to the direction of regional extension, which may or may not be parallel to the basement faults. But as long as the salt



(a) Very rapid slip or very thin salt

(b) Moderate slip or thin salt



(c) Slow slip of thick salt

(d) Sedimentation during rapid slip and thin salt

Figure 6 Effect of salt thickness and slip rate of basement fault on structural style of cover. Regional datum delineates original top of overburden before deformation began. All rock units are prekinematic (originally uniform thickness before deformation began), except where labeled otherwise. Diapirs are initiated above white triangles. (After Vendeville et al 1993.)

layer is thick, the location, spacing, and throw of the overburden structures are effectively independent of all but the larger basement faults. Only after the source layer greatly thins does the basement footwall indent the overburden, forming new faults and folds directly above the basement fault.

Rather than being filled by salt, the space created by basement downthrow can be balanced by locally thickened synkinematic sediments situated above draped overburden (Figure 6*d*). The load of accumulating sediments can reverse flow in the salt, squeezing salt from above the downthrown block to above the upthrown block (Vendeville et al 1993a).

In summary, extension of the overburden directly causes diapirism, regardless of basement faulting, unless salt is thin or the slip rate is high. A basement fault affects diapirism only indirectly by causing or allowing the overburden to extend.

SUBSIDING DIAPIRS, SALT WELDS, AND FAULT WELDS

Regional extension eventually retards the rise of diapirs and even causes them to subside. Salt walls must widen between diverging blocks of overburden, thereby increasing demand on salt supply. The supply rate diminishes as the source layer thins due to extension and withdrawal into the diapir. Eventually salt import is too slow to supply salt for expanding walls. The diapir begins to sag. Sedimentary thicks fill either (*a*) half-graben rollovers above the diapir's subsiding flanks, leaving the crest unaffected (Figure 7*A*) or (*b*) a linear, or even circular, crestal graben above the sagging crest of the diapir, which inverts from a topographic bulge to a subsiding graben (Figure 7*B*). Adjoining horns of salt project into each fault bounding the graben. Salt dikes can intrude faults (Hale et al 1992), but these extensional horns are not injections; they are residual structures whose apices record the original diapiric crest.

Other products of source-layer depletion during regional extension include *turtle-structure anticlines* (Figure 7*A*), which form because each end of an overburden slab subsides as the diapir supporting it sags. This anticlinal bending of the slab forms a crestal fan of faults. Turtle-structure anticlines have been attributed solely to salt withdrawal from the periphery of a pillow as it evolves into a *rising* diapir (Trusheim 1960), but turtle structures also characterize a *subsiding* diapir during regional extension (Duval et al 1992, Vendeville & Jackson 1992b). Extensional turtle structures are subparallel and elongated, whereas classic withdrawal turtles are equant.

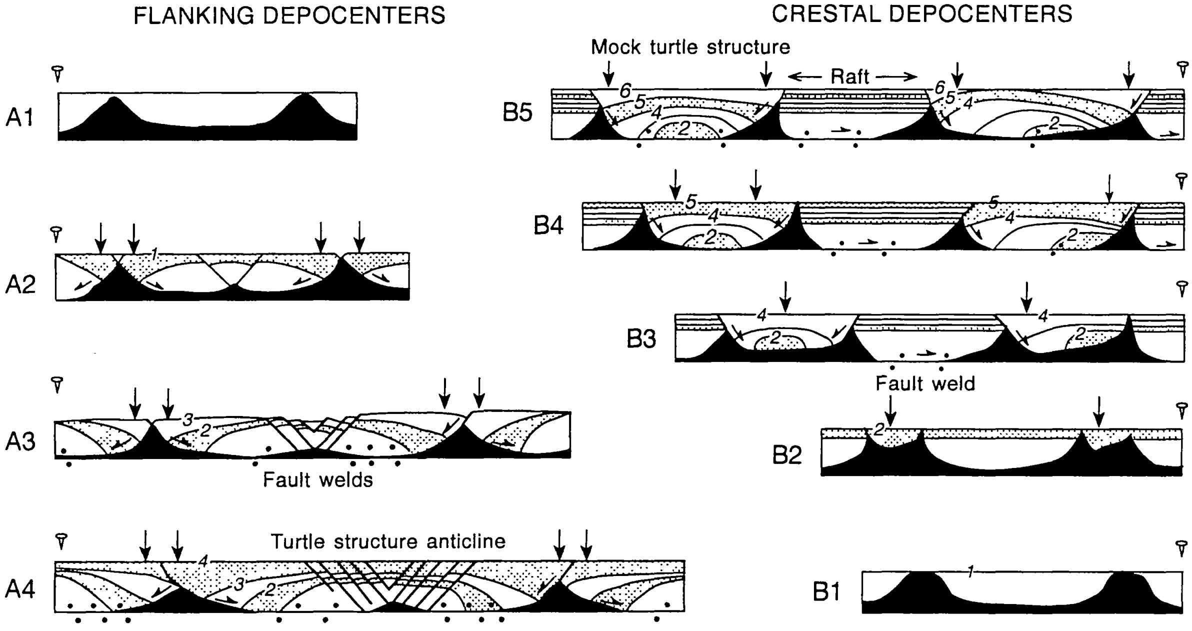


Figure 7 Formation of turtle structure anticline (A1–A4) and mock turtle structures and rafts (B1–B5) during extensional salt tectonics. Vertical arrows show depocenters extensionally created by subsiding diapirs.

During extreme thin-skinned extension, another type of turtle structure can form (Figure 7*B*). A crestal depocenter indents a subsiding diapir so much that the graben grounds onto the basement, segmenting the original diapir into two triangular relics separated by a deep, synclinal depocenter. If extension continues, the flanks of these diapiric relics subside, and the synclinal depocenter inverts into a crestally-faulted arch called a *mock-turtle anticline* (Vendeville & Jackson 1992b). A mock-turtle anticline can be distinguished from a turtle anticline because it initiates *above* a diapir rather than *between* diapirs; also, below a mock-turtle anticline a large stratigraphic section—representing the duration that the precursor diapir was at the surface—is missing.

The formation of mock-turtle anticlines is part of *raft tectonics*. Normal-fault blocks separate entirely into glide blocks called rafts (Figure 7*B*). The Kwanza basin of Angola is the archetypical area for raft tectonics (Duval et al 1992). There, transgressive carbonate overburden began to extend when only a few hundred meters thick, forming many small, tilted, Phase 1 rafts over relatively thick salt. These rafts were yoked together by sedimentation before rupturing into much larger nontilted Phase 2 rafts 10 to 20 km long, which slid concordantly over thinned salt. Younger sediments accumulated asymmetrically in strike-parallel depocenters created by widening grabens separating Phase 2 rafts. Apart from sea-floor spreading, lateral space for extension was created locally by downdip fold-and-thrust-belts and by transfer of salt to higher stratigraphic levels in the form of allochthonous sheets.

Some of the structures of extensionally-induced subsidence have been interpreted as arising purely by dissolution. Subsurface dissolution cannot form large crestal grabens or flanking half grabens wherever the salt diapir is encased in impermeable shale, deeply buried, lacking cap rock, in an environment of rapid sedimentation, or where palinspastic restoration indicates abrupt, deep subsidence of a diapir crest after a much longer time at the surface (Duval et al 1992, Vendeville & Jackson 1992b).

Salt welds are surfaces or thin zones joining strata originally separated by thicker salt that is mostly or completely removed by lateral creep or dissolution (Jackson & Cramez 1989). Because salt can seal petroleum systems, the thickness of salt remaining in the weld affects the vertical migration of hydrocarbons. Welds commonly separate discordant or disconformable strata. Primary, secondary, and tertiary welds, respectively, join strata originally separated by gently dipping bedded salt, steep-sided salt diapirs, or gently dipping allochthonous salt sheets.

Salt welds are difficult to differentiate from fault welds, which have significant fault slip or shear along the smeared layer of salt. Both types of welding can create discordances and disconformities, and both create accommodation space (Figure 8). Most cross sections of salt structures can be ambiguously restored by assuming either salt welds and salt withdrawal or fault welds and extension (Hossack & McGuinness 1990, Diegel et al 1993). However, the type of welding can be distinguished by local structural criteria (Figure 8) or by knowing the total extension of the section or the original salt thickness.

CONTRACTIONAL SALT TECTONICS

Contractional salt tectonics is produced by convergent (including transpression) plate tectonics—often superposed on a divergent rift or continental margin—or is restricted to the narrow seaward margin of a divergent continental margin. Contractional salt tectonics conveniently divides into two types, cases in which: (*a*) thin salt merely lubricates decoupling of the overburden from its basement and fills anticlinal cores and (*b*) thicker salt forms diapirs before contraction.

Examples of thin-skinned contraction over thin salt are in Arctic Canada; the Franklin Mountains of northwest Canada; the Appalachians; and the deep-water fringes of largely divergent salt basins such as the Perdido and Mississippi Fan fold belts of the Gulf of Mexico, the Kwanza basin of Angola, the Campos basin of Brazil, and the southern Red Sea.

These contractional belts over thin salt are exemplified by the eastern Parry Islands fold belt of Arctic Canada (Harrison & Bally 1988, Harrison et al 1991), which contains two décollements: a lower evaporite and an upper shale. The sinuously arcuate fold belt comprises upright folds cored by thrust complexes. Anticlines were initiated by buckling of the competent carbonate-dominated strata between the weak salt and shale (Figure 9*a*). The salt then migrated laterally into the cores of the anticlines, forming pinched ridges called *salt welts*. The less lithified sequence above the strong carbonates contracted by compaction and uniform thickening. The carbonates then contracted by thrusting as the overlying strata buckled (Figure 9*b*). Shale was squeezed into the anticlinal cores to form sharp-crested shale welts stacked over the salt welts (Figure 9*c*). A second generation of thrusts subsequently ramped from the shale into the competent layers above (Figure 9*d*). Anticlinal cores also contain pop-up structures, thrust splays, and faults that zigzag up-section, reversing sense of slip. The folds and thrusts verge equally forward (to south) and backward (to north), a characteristic of décollement over salt (Davis & Engelder 1987).

The amount of shortening in diapiric provinces ranges from mild inversion of an extensional system (North Sea, Germany) to full fold-and-thrust belts in various Alpine chains of Europe, North Africa, southwest Asia, and elsewhere. Inversion in the North Sea and Germany is the foreland

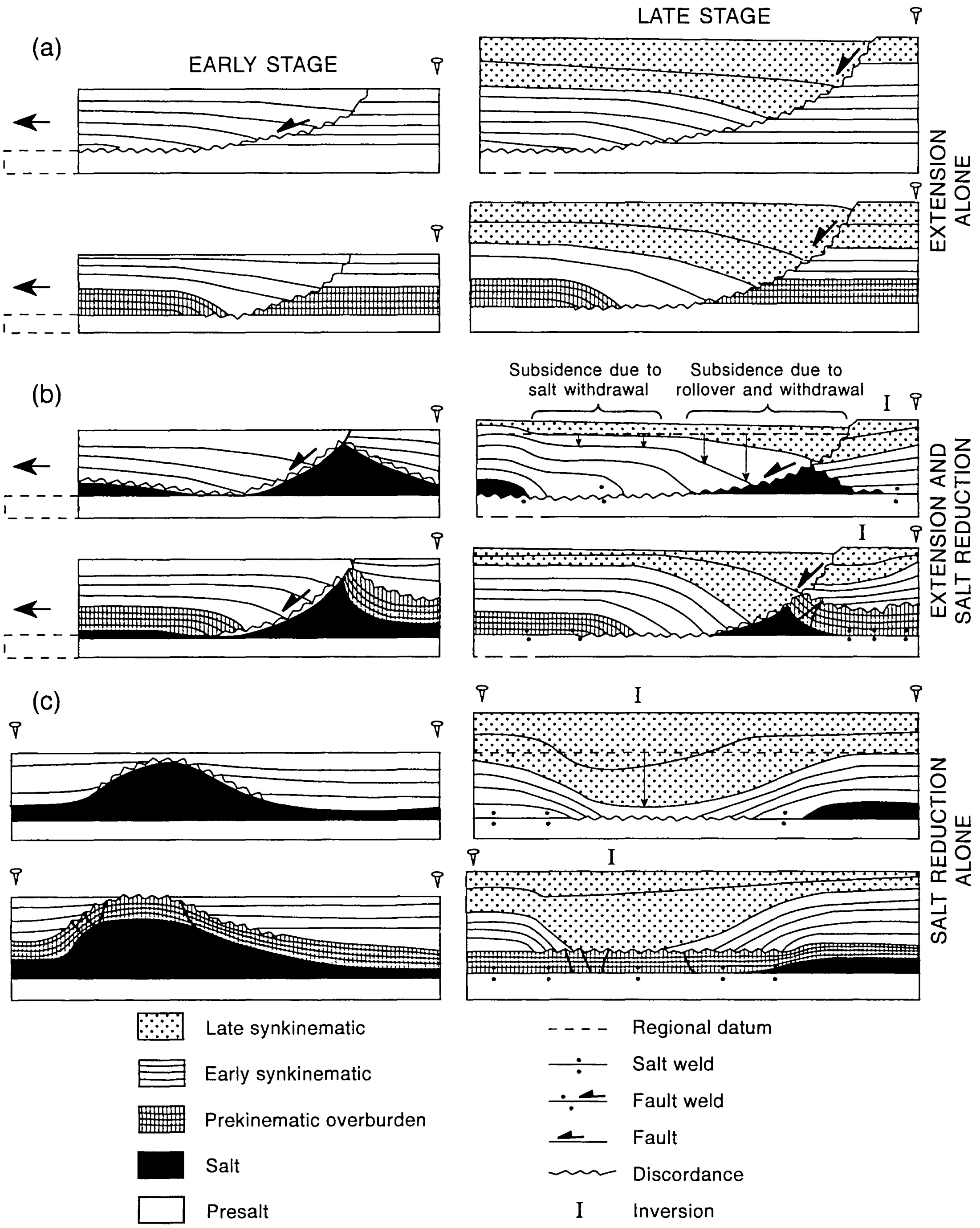


Figure 8 Diagnostic features of (*a*) growth faults, (*b*) fault welds, and (*c*) salt welds. Upper section of each pair shows only synkinematic strata, whereas lower section shows basal prekinematic unit of overburden. Vertical arrows show subsidence below regional datum (*dashed line*).

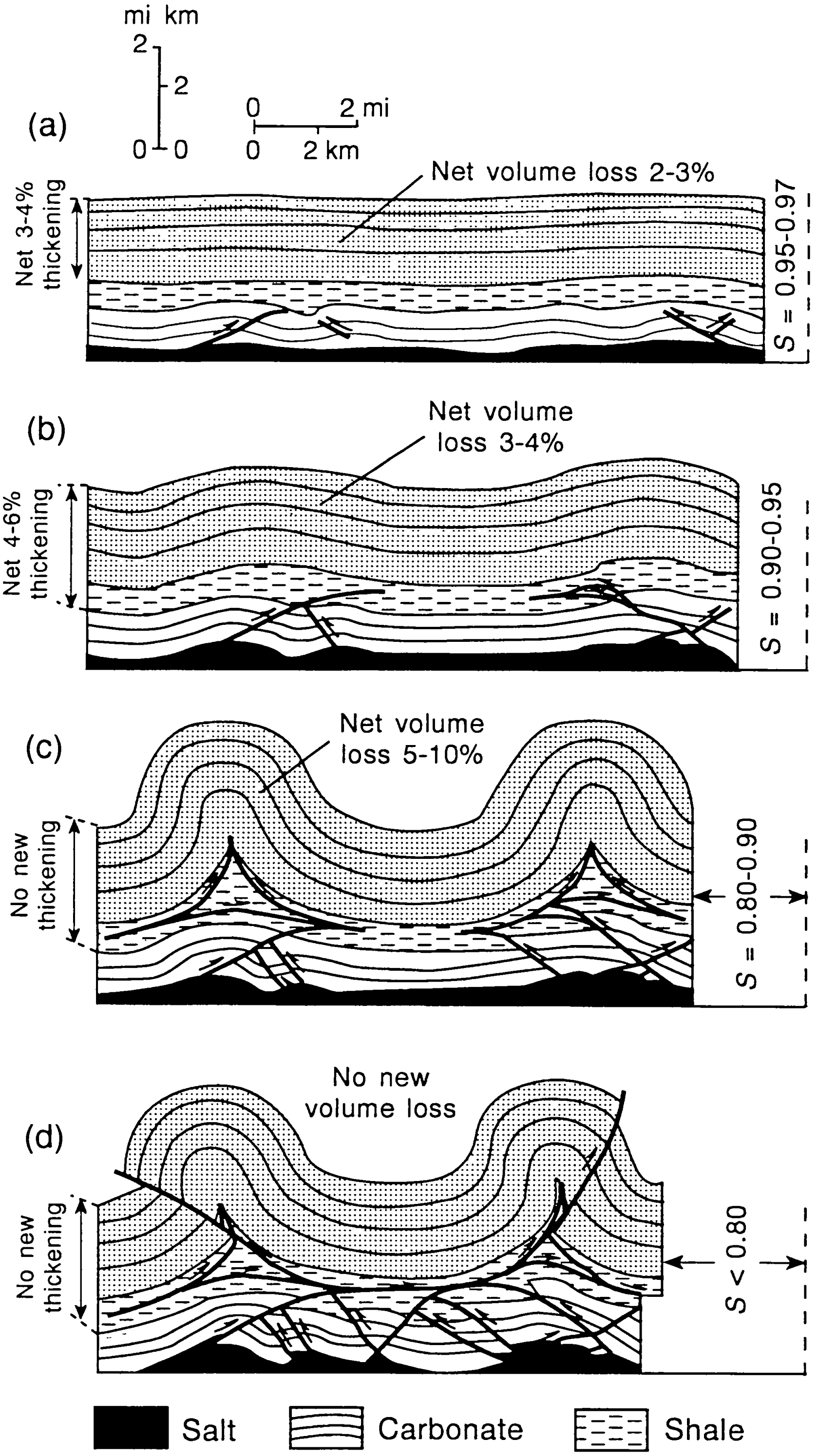


Figure 9 Evolution of salt and shale welts (pinched cores of anticlines) and associated thrusts during contractional salt tectonics in the eastern Parry Islands fold belt, Arctic Canada (after Harrison & Bally 1988, Harrison et al 1991).

expression of severe Alpine shortening to the south. Many concordant salt structures in this region (e.g. Silver Pit region in the UK North Sea) are still widely attributed merely to buoyant upwelling of Zechstein salt by gravity-driven halokinesis. Yet these swells are overlain by domes or anticlines of overburden, whose imposing thickness and uniformity exclude a purely halokinetic origin by active doming. Rather, these folds must have formed by regional contraction during inversion. Salt merely collected in the lower-pressure anticlinal cores.

Salt lubrication promotes the lateral propagation of fold-and-thrust belts. Kastens (1991) attributed accelerated advance (<10 km Ma^{-1} increasing to 12–22) of the Mediterranean Ridge accretionary prism to the Neogene deposition of Messinian salt across the ridge.

Most salt diapirs in fold-and-thrust belts were apparently initiated in extensional systems long before contraction began (Jackson & Vendeville 1994). For example, both the Spanish Prebetic and the French Maritime Alps were rifted several times. Palinspastic removal of the overprint of Alpine folding reveals these rift-phase structures (de Ruig 1992, Dardeau & Graciansky 1990). Salt diapirism began during rifting soon after the salt accumulated. The diapirs rose reactively into tears of the less-dense overburden. During ensuing crustal shortening, the preexisting diapirs were distorted, truncated, and further injected into the contracting overburden, and new salt-cored folds formed. Diapirs in these Alpine contractional belts show no consistent relationship to anticlines or synclines.

Physical modeling shows that whether the salt is thin or thick, merely buckling the overburden cannot produce diapirs (Vendeville 1991). Salt could diapirically break out from anticlinal cores if (*a*) the anticlinal crest was thinned by erosion or local extension and (*b*) the anticlinal core of salt was confined and pressurized as adjoining synclines grounded onto the basement and blocked lateral escape of salt. Thrusting is generally more effective than buckling in forming diapirs.

ALLOCHTHONOUS SALT SHEETS

One of the most important discoveries in salt tectonics has been that irregular masses of salt hundreds of square kilometers in area are actually allochthonous sheets rather than rooted, autochthonous massifs. Salt allochthons characterize divergent continental margins such as the Gulf of Mexico and are also known from offshore Brazil, West Africa, and the Red Sea. Instead of extending down several kilometers to the source layer, these salt bodies overlie younger strata. This realization opened the possibility of prodigious subsalt hydrocarbon reservoirs, particularly in the Gulf of Mexico, where a vast complex of salt allochthons increases in size and degree of coalescence westward (Liro 1992, Wu 1993). Most salt sheets there have partly or completely coalesced into composite diapirs called canopies (Jackson et al 1990). The leading edge of this complex climbed 8 km vertically and up to 80 km laterally to form the Sigsbee Scarp at the base of the continental slope.

Thin allochthonous sheets gravitationally spread over or through weak, unconsolidated, mud-rich, low-density sediments within a few 100 meters of the sediment surface (Nelson 1991, Fletcher et al 1993, McGuinness & Hossack 1993). The observed shallow emplacement levels indicate that the mechanical control is overburden strength rather than overburden density; were density the controlling factor, sheets would spread at a level of neutral buoyancy 1 to 1.5 km deep.

Stratigraphic markers record the timing and direction of allochthonous salt emplacement (Figure 10). Sedimentary or structural truncation of underlying strata against the lower contact of a salt tongue forms a basal cutoff. Each cutoff marks the former leading edge of the advancing salt tongue. *Salt ramps* and *flats* are the steeply inclined and gently inclined segments of the stairstep base of a salt tongue, which cut up the stratigraphic section in the direction of emplacement. Ramps form where the ratio of aggradation to salt spreading is high. Salt spreads and flats form when sedimentation is very slow. The basal cutoffs of salt sheets suggest that their injection rate is up to 100 times faster than the coeval aggradation rate (Jackson & Talbot 1991). In the Gulf of Mexico, salt sheets appear to spread after each episode of downdip contraction associated with updip extension (Peel et al 1993).

An allochthonous sheet can act as a source layer for a second cycle of salt tectonics. During progradation, the rear (landward) margin of the sheet gradually segments. Intrasalt minibasins evolve from slope basins above tabular salt allochthons to shelf basins, which are bounded by arcuate growth faults formed largely by salt withdrawal downslope rather than by extension (Diegel et al 1993). The sheet is segmented into discrete salt structures separated by salt welds and fault welds, together constituting a *roho system* (Schuster 1993). Initially the allochthon acts as a deflating cushion that accommodates sediment deposition in intrasalt basins. After much time, loading, or extension, the original sheet transforms into thin but broad welds, which act as major subhorizontal detachments for deep growth faults whose slip creates space for further sedimentation (Diegel et al 1993, Peel et al 1993). Second-cycle salt structures rise from these deformed sheets (Worrall & Snelson 1989, Hossack & McGuinness 1990, Seni 1992).

As the sheet segments, the weight of overlying sediments squeezes salt toward the shallow leading edge of the allochthon. The salt sheet must

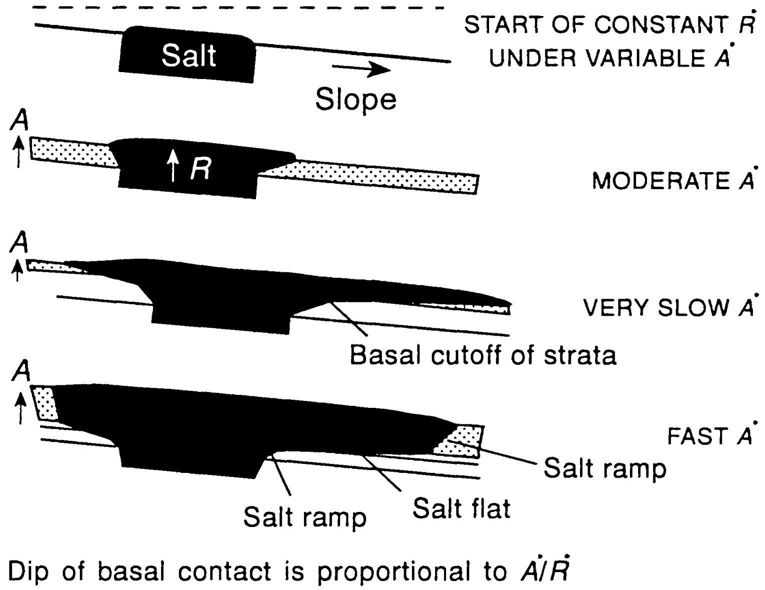


Figure 10 Asymmetric spreading of a passive salt sheet over or just below the sedimentary surface. Highly variable aggradation rate produces salt ramps, salt flats, and basal cutoffs along the base of the salt allochthon. A = aggradation increment, $\dot{A}$ = aggradation rate, R = salt rise increment, $\dot{R}$ = salt rise rate.

continually climb stratigraphic section and break through the continually accumulating sedimentary veneer. Physical modeling (Jackson & Vendeville 1993) suggests that salt is expelled seaward from beneath a prograding wedge into the core of a box fold fringing the wedge. Fore- and back-grabens form along the hinges of the box fold. Pressurized by the prograding overburden, a diapiric wall emerges through the structurally-thinned back-graben. Extrusions from the fore-graben flow down and invert the forelimb then spread radially to coalesce as a lobate canopy. The roof of the box fold detaches by arcuate slumps and ruptures into rafts, which are overthrust on the back of the spreading salt over the inverted limb of the isocline, locally stacking and repeating stratigraphy. Well intersections of repeated stratigraphy and structural reconstructions prove that the sedimentary roof of a spreading salt sheet is carried as an overthrust over equivalent strata below the salt sheet (McGuinness & Hossack 1993). The stretching carapace protects the underlying salt from dissolution, as does the residual crust of insoluble gypsum originally disseminated in salt.

The allochthons are fed by landward-dipping stalks, which flatten into salt welds by contraction and by withdrawal into the salt sheet. Once welded, these huge landward-dipping discontinuities, known as stepped counter-regional systems (Schuster 1993), resemble listric normal growth faults; but little slip has occurred along them, and the adjoining rollover anticlines are produced by salt withdrawal rather than by extension (Jackson & Vendeville 1993, Schuster 1993).

The cycle of sheet emplacement, inflation, segmentation, and climb can be repeated several times, creating several tiers of allochthonous salt. Three or four such tiers have been recognized in the northern Gulf of Mexico (Diegel & Schuster 1990, Peel et al 1993, Wu 1993).

CONCLUSION

Salt tectonics represents the interplay between salt, acting as a weak but highly pressurized fluid and its overburden, acting as a strong, brittle roof, which faults in response to the tangential forces of plate tectonics, to its own weight, and to salt pressure. Varied responses allow salt to smear as a lubricant, to rise as massive diapirs, or to spread as vast glaciers. Although some astonishing concepts have been revealed in the past few years, research on salt tectonics proceeds apace, and new concepts are rapidly overhauled by even newer ones. Several topics are still murky, but some are likely to clarify soon: 1. establishing whether extension above and below salt layers can be cryptically coupled, 2. quantifying the roles of extension versus salt withdrawl in creating accommodation space, 3. distinguishing between synclines produced by salt withdrawal (bending folds) and by regional contraction (buckling folds), 4. determining the role of contraction in the emplacement of salt sheets, 5. establishing the degree of exposure and protection from dissolution of submarine salt glaciers, and 6. examining suture processes where diapirs coalesce into canopies.

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