Salt tectonics at passive margins: Geology versus models

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ABSTRACT

Salt tectonics at passive margins is currently interpreted as a gravity-driven process but according to two different types of models: i) pure spreading only driven by differential sedimentary loading and ii) dominant gliding primarily due to margin tilt (slope instability). A comparative analysis of pure spreading and pure spreading is made using simple mechanics as well as available laboratory experiments and numerical models that consider salt tectonic processes at the whole basin scale. To be effective, pure spreading driven by sedimentary loading requires large differential overburden thicknesses and therefore significant water depths, high sediment density, low frictional angles of the sediments (high fluid pore pressure) and a seaward free boundary of the salt basin (salt not covered by sediments). Dominant gliding does not require any specific condition to be effective apart from the dip on the upper surface of the salt. It can occur for margin tilt angles lower than 1° for basin widths in the range of 200–600 km and initial sedimentary cover thickness up to 1 km, even in the absence of abnormal fluid pressure. In pure spreading, salt resists and sediments drive whereas in dominant gliding both salt and sediments drive. In pure spreading, extension is located inside the prograding sedimentary wedge and contraction at the tip. Both extension and contraction migrate seaward with the sedimentary progradation. Migration of the deformation can create an extensional inversion of previously contractual structures. In pure spreading, extension is located updip and contraction downdip. Extension migrates downdip and contraction updip. Migration of the deformation leads to a contractual inversion of previously extensional structures (e.g. squeezed diapirs). Mechanical analysis and modelling, either analogue or numerical, and comparison with margin-scale examples, such as the south Atlantic margins or northern Gulf of Mexico, indicate that salt tectonics at passive margins is dominated by dominant gliding down the margin dip. On the contrary, salt tectonics driven only by differential sedimentary loading is a process difficult to reconcile with geological evidence.

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1. Introduction

Numerous passive margins display thick layers of evaporites, among which halite (salt) dominates, that are most often deposited during and/or immediately after continental rifting. They commonly display deformations over periods as long as 100 Ma or more, as recorded by syntectonic sedimentation in extension or contraction (Fig. 1). Most often, these deformations are still active. Such salt basins are extremely unstable because salt is a very weak material that is able to flow under very low differential stresses, even at surface temperature conditions. In many salt margins where no regional tectonic effects are recorded, the observed deformations are obviously gravity driven with two often-invoked origins of differential stresses, namely margin tilting and/or differential sedimentary loading. Therefore, passive margins are favourable natural laboratories to study the dynamics of salt tectonics, in the absence of crustal-scale tectonics.

When the only tools available to study salt-related structures were surface geology and drilling, attention was focused on salt diapirs and salt tectonics was mostly perceived as a deformation involving dominantly vertical displacements — i.e. salt upwelling from a source layer buried below a sedimentary cover. The exploration of passive margins became possible with the fast technological development of seismic imaging, thereby allowing the identification of “raft rafts” in the Angolan margin (Burollet, 1975). The nearly horizontal seaward displacements of these tectonic rafts on top of salt have controlled rollover-type sedimentary depocentres (Fig. 1). Today, this discovery appears to have been a breakthrough in the understanding of salt-related deformation processes. “Raft tectonics” is the most extreme form of
thin-skin extension, in which horizontal seaward displacements dominate (Duval et al., 1992). It is a gravity-driven process corresponding to the gliding of the salt—sediment package primarily stimulated by margin tilt, as modelled in laboratory experiments (Cobbold and Szatmari, 1991; Demercian et al., 1993; Gaullier et al., 1993; Mauduit et al., 1997; Fort et al., 2004a,b). Rafts form at early stages of extension by brittle faulting of thin sedimentary layers, generally less than 1 km thick, deposited on top of the salt (Lundin, 1992; Liro and Cohen, 1995; Spathopoulos, 1996; Rouby et al., 2002). Seismic data in the ultra-deep part of the margins has shown that updip extension was coeval with either thrusting and buckling of the sedimentary cover downdip or flowing of frontal salt nappes on top of the abyssal plain (Cobbold and Szatmari, 1991; Cobbold et al., 1995; Marton et al., 2000; Cramez and Jackson, 2000; Tari et al., 2001; Gottschalk et al., 2004; Brun and Fort, 2004; Fort et al., 2004a). Even if sedimentary loading participates in the gliding process, its role is not a major one as salt is able to flow downdip without any sedimentary loading.

The northern Gulf of Mexico (GoM) presents a dual situation (Worrall and Snelson, 1989). The northwest margin (Texas) shows a whole picture that is somewhat comparable to south Atlantic margins with an updip extension characterised by long and straight normal faults and a downdip contraction characterised by a well-developed fold belt (Peel et al., 1995; Hall, 2002), in conceptual models (Schuster, 1995; Rowan, 1995) as well as in experimental and numerical dynamic models (Talbot, 1992, 1993; Vendeville, 2005; Gaullier and Vendeville, 2005; Gradmann et al., 2009). The concept of salt tectonics dominantly driven by differential sedimentary loading, which was elaborated upon in the GoM, has been tentatively applied to other passive margins including those of western Africa (e.g. Fig. 10 in Hudec and Jackson, 2002; discussion in Hudec and Jackson, 2004), and Brazil (e.g. Szatmari et al., 1996; Fig. 17 in Ge et al., 1997).

Consequently, two different types of dynamic models are presently invoked to explain gravity-driven salt tectonics at passive margins: i) gravity spreading driven only by differential sedimentary loading and ii) gravity gliding primarily due to margin tilt. The difference between the two types of explanations or the choice between one over the other are not purely issues of academic interest because the two types of processes have significantly different patterns of displacement and different modes of relationships between deformation and sedimentation, with obvious consequences for hydrocarbon exploration.

Our aims in the present paper are first to compare the two types of models on the basis of simple mechanics and modelling results and second to discuss their applicability to salt tectonics at passive margins. We identify structures, or associations of structures, that can be considered as diagnostic of each of the two processes as well as those that are ubiquitous and therefore not significant for any of them. Finally, we discuss the reasons why the GoM is most often considered as being different.

2. Salt basins

2.1. Deposition of thick salt layers

Thick accumulation of evaporites requires specific conditions. Seawater evaporation progressively concentrates brines whose volume reduction is responsible for a drop of the water surface below the worldwide sea level. Consequently, the salt basin must be isolated from the ocean, at least temporarily. The evaporation of 1000 m of seawater does not give more than 15 m of evaporites and

Fig. 1. Composite regional-scale cross-section of the north Angolan margin showing the extensional and contractional deformation domains and the type of individual structures. See location of the two sub-lines 1 and 2 in the lower left insert (modified after Fort et al., 2004a). The two enlargements of the cross-section show the growth character of structures since 100 Ma, in extension to the right and contraction to the left.
consequently, the evaporation of at least 66,000 m of seawater is necessary to produce a 1000 m thick layer of evaporites. As at passive margins, evaporite layers sometimes reach several thousand metres and salt basins cannot result only from the single evaporation of one body of seawater, regardless of its depth. A quasi-constant seawater refill of the basin is needed. For a full discussion of depositional environments leading to several kilometre-thick accumulations of evaporites, see the recent syntheses by Rouchy and Blanc-Valleron (2006) and Warren (2006).

Among the modes of lithosphere necking that can give birth to passive margins, depth-dependent stretching (see review by Davis and Kusznir, 2004) leads to an extremely attenuated crust with thermal subsidence histories that favour shallow water deposition of evaporites under appropriate climatic conditions (Huismans and Beaumont, 2008).

2.2. Salt basin size and geometry

The mean width of salt basins in the margin dip direction is in the range of 200–400 km, as illustrated in the South Atlantic (see Fig. 1 in Davison, 2007). However, some salt basins could have initially been narrower — e.g. in Nova Scotia: possibly less than 100 km (Ings and Shimeld, 2006; Albertz et al., 2010). Other basins are obviously larger — e.g. Gulf of Mexico: between 500 and 800 km (Worrall and Snelson, 1989).

The initial basin geometry is often not well assessed. This is in particular due to the difficulty to clearly image the base salt, but also applies to narrow basins like in Nova Scotia and large basins like in the Gulf of Mexico. Some attempts to reconstruct the initial geometry of salt basins have led to bulk double-wedge shapes at the margin scale, with the salt pinching out landward and seaward and with a maximum salt thickness located toward the seaward basin termination — e.g. Angola: Marton et al., 2000; Brazil: Quirk et al., 2011. In detail, the basin geometry can be more complicated with the general double-wedge shape divided into several sub-basins separated by basement highs but which can remain connected — e.g. Angola: Hudic and Jackson, 2004 (see Fig. 2). Such basement highs are likely inherited from normal faulting and variations in the amount of crustal stretching related to rifting. However, some basement deformation can also occur after salt basin deposition — e.g. extensional in the North Sea (Nalpas and Brun, 1993; Stewart and Coward, 1995; Stewart et al., 1997).

2.3. Initial sedimentary cover above salt

In all salt margins where the seaward salt pinch-out is clearly imaged, the most distal part of the salt basin appears to be covered by sediments soon after the end of evaporitic sedimentation, as exemplified in most Atlantic passive margins (e.g. West Africa: Marton et al., 2000; Tari et al., 2001; Brazil: Cobbold and Szatmari, 1991; Demercian et al., 1993; Davison, 2007) as well as in the Gulf of Mexico (e.g. Peel et al., 1995; Trudgill et al., 1999; Grando and McClay, 2004). This indicates that an initial sedimentary cover is generally deposited on top of the salt at the scale of the whole basin, soon after salt deposition.

In many salt margins, the initial position of the top salt surface can be deduced from the inferred location of landward and seaward salt pinch-outs, as illustrated in Figure 2a. As the top salt surface is initially close to horizontal, the straight line joining the two pinch-outs allows a direct and accurate calculation of the margin tilt (e.g. 1.17° in Fig. 2c). It also provides a good estimate of the initial salt basin geometry and volume, especially if the geometry of the base salt is known. In the particular case of the Angolan margin, it is interesting to note that the initial salt surface passes slightly above a central basement high, indicating that the salt basin at the margin scale was initially divided into two connected sub-basins and that both sub-basins have been tilted by the same angle (Fig. 2a). In addition, the comparison between the initial top salt surface and final salt–sediment interface help check the conservation (as in Fig. 2c) or non-conservation of the salt volume and the amount of

![Fig. 2](https://example.com/fig2.png)

**Fig. 2.** Regional section across the Kwanza Basin in South Angolan Margin. (a) Vertically exaggerated section. (b) Same section with no vertical exaggeration (a and b modified after Hudic and Jackson, 2004). (c) Simplified section showing the relative amounts of sediments and salt below and above initial top salt surface.)
salt transfer at the margin scale. Using a similar argument, the salt volume also seems to be conserved on the Brazilian margin (Davison et al., 2011).

3. Modelling of salt tectonics at passive margins

The geometrical and geological features of margin salt basins summarised in the previous section provide important constraints for the setting up of the initial and boundary conditions of salt tectonic models applicable to passive margins (Fig. 3).

3.1. Initial conditions

At passive margins, the salt basin width is commonly in the range of 200–400 km. As salt is most often pinched out landward and seaward, the salt thickness is not constant and can reach several thousand metres inside the basin. A shallow water environment of thick salt deposition implies a horizontal top salt surface, close to the global sea level.

In very specific salt deposition conditions — e.g. in the Mediterranean during the Messinian salt crisis, in continental rifts or in the Red Sea — diachronous salt deposition can occur in several small basins with different levels of top salt (see reviews in Rouchy and Blanc-Valleron, 2006; Warren, 2006). In such basins — e.g. Messinian — the local sea level can be well below the global sea level. In addition, small and separated salt basins with different levels of top salt are too small to result in salt tectonics processes like those observed at the scale of whole passive margins.

On the Brasilian margin, salt appears to have deformed prior to the deposition of the sedimentary cover indicating a margin tilt at an early stage of evolution (Fiduk, 2010; Davison et al., 2011; Quirk et al., 2011). More commonly, the salt basin is entirely covered by an initial sedimentary layer, often called a prekinematic layer, whose thickness is often around several hundred metres, but which can reach a kilometre or more locally. In such circumstances, the frontal (seaward) part of the salt basin cannot be easily modelled as a free boundary.

3.2. Boundary conditions

In the absence of tectonic displacements applied to the salt basin basement, two types of boundary conditions must be considered: i) margin tilt and ii) differential sedimentary loading.

Margin tilt is a consequence of either continuing rifting or post-rift thermal subsidence and depends on the margin width. At the end of subsidence, the tilt can reach values between 1 and 2° for margin widths of several hundred kilometres (e.g. Fig. 2). For narrower margins, it can reach values up to 3–4° (Mauduit et al., 1997).

Synkinematic sedimentation (sedimentary loading) cannot exceed water depth and therefore is directly dependent on margin tilt. This implies that the role of sedimentary loading is minimum at the onset of the process and increases with progressive tilting.

3.3. Two categories of models

Available numerical and physical models of salt tectonics at passive margins fall into two categories (Fig. 3b and c): i) pure gravity spreading driven only by sedimentary loading and ii) dominant gravity gliding primarily driven by margin tilt. Because dominant gravity gliding is also accompanied by sedimentation, margin tilt is the main difference between the two categories of models, in terms of boundary conditions. A single ductile layer (Brun and Merle, 1985) or a brittle-ductile two-layer system such as sediments lying on top of salt (Mauduit et al., 1997) that flows down

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Fig. 3. Setting of salt basins at lithosphere scale at the end of rifting (a) and the two categories of salt tectonic models: Pure spreading (b) and dominant gliding (c).
a slope under its own weight undergoes a combination of gliding and spreading. For simplicity, this is here referred to as “dominant gliding” instead of “gliding-spreading”.

Before any detailed analysis is presented, it must be noted that in pure spreading, salt starts to flow seaward as it is expelled from below the prograding sedimentary wedge whereas in gliding, the entire salt layer flows down-dip (Fig. 3b and c). In other words, salt resists in spreading whereas it drives in gliding. This major difference requires a separate treatment and justifies the making of a clear distinction between pure spreading and dominant gliding. See Appendix 1 for further details.

4. Pure spreading models: salt tectonics only driven by differential sedimentary loading

4.1. Laboratory experiments

As already mentioned, the concept of salt tectonics only driven by differential sedimentary loading was inspired by studies carried out in the Gulf of Mexico (Worrall and Snelson, 1989; Diegel et al., 1995; Peel et al., 1995; Rowan, 1995; Schuster, 1995) where the salt is very thick and sedimentation rates were rather high in the Neogene. Different types of laboratory experiments (Ge et al., 1997; McClay et al., 1998; Vendeville, 2005; Gaulier and Vendeville, 2005; Krézsek et al., 2007) or numerical modelling (Cohen and Hardy, 1996; Gemmer et al., 2004; Ings et al., 2004; Ings and Shimmeld, 2006; Gradmann et al., 2009) have been carried out to characterise the mechanical behaviour of salt tectonic systems driven by the progradation of a sedimentary wedge on top of a salt basin, either covered by or not covered by sediments from the outset. In all these models (Fig. 4), a horizontal layer of viscous material — i.e. salt — is submitted to the progressive loading of a prograding sedimentary wedge. Salt is expelled forward and the wedge spreads.

Figure 5 presents an experiment of this type. The initial model is 150 cm long and made of a 2 cm thick layer of silicone putty covered by a 1 cm thick layer of sand. Both layers are horizontal and have a constant thickness. The initial geometry of the model represents a margin salt basin prior to the arrival of a prograding sedimentary wedge. Sedimentation starts at the “landward” model end and is done at regular time intervals with a step by step progradation toward the “seaward” model end. At the end of experiment, the model is wetted and vertical sections are cut to examine the internal structures (Fig. 5a). Top views of the model surface (Fig. 5b) show the progressive evolution of deformation during prograding sedimentation. The general principles of small-scale
modelling and the physical properties of the materials used to model the salt tectonic systems are given in Appendix 2.

After 20 h, extension starts to develop inside the wedge with a contractional domain formed at its tip (Fig. 5b). The top views at 40 h and 70 h show the progressive frontward migration of both extension and contraction. The model section (Fig. 5a) shows that both extensional and contractional structures were progressively sealed by the newly deposited sediments and that extension and contraction propagated frontward with sedimentary progradation.

This experiment, as well as those previously done by other authors (McCray et al., 1998; Vendeville, 2005) (Fig. 6a and b), shows that extension develops coeval with contraction at the tip of the prograding wedge. The sedimentary wedge deposited at one model end exerts a differential loading on top of the initial sand-silicone two-layer system. At the tip of the sedimentary wedge, the initial sand cover is submitted to a horizontal stress \( \sigma_h \) that is proportional to the loading applied by the wedge on top of the silicone layer. With ongoing sedimentation, \( \sigma_h \) progressively increases until it becomes large enough to overcome the strength of the initial sedimentary cover. This triggers a zone of contraction at the wedge tip and a simultaneous wedge collapse backward. In other words, the sedimentary wedge spreads under its own weight if the sedimentary loading is large enough to overcome the strength of the sedimentary cover. This experiment, like those shown in Figure 6, illustrates the basic mechanisms by which sediments deposited on top of a horizontal salt basin could be able to drive the gravity spreading of a salt–sediment pile.

The experiments shown in Figures 5 and 6 illustrate that sedimentary progradation leads to a frontward (seaward) migration of contraction. During progradation, contractional structures are first sealed by newly deposited sediments and can be later reactivated (inverted) in extension (Fig. 6a and b).

4.2. Stresses at failure

Experiments (Figs. 5 and 6) show that the thickness of the sedimentary overburden \( \Delta h \) necessary to activate gravity spreading is several times thicker than the initial sedimentary cover \( h \) (Fig. 4). In order to apply this to natural systems, because sedimentation requires a certain water depth, the effects of water must be taken into account. This can be easily done in numerical models but not in laboratory experiments. As pointed out by Gemmer et al. (2005), the presence of water has several effects among which, in particular, is the increase of solid and fluid pressure in the sediments. In the following, we consider the minimum value of overburden \( \Delta h \) as a function of the thickness \( h \) of sediments deposited on top of salt that is required to reach system failure. To keep the calculations as simple as possible, we assume that salt behaves like an inviscid fluid — i.e. negligible salt viscosity. By ignoring the resisting effect of salt ductile flow, the effect of sedimentary loading is slightly overestimated.

At the future site of contractional failure (Fig. 4), at the wedge tip, the horizontal stress \( \sigma_h \) depends on the pressure gradient that results from the difference between the density of sediments \( \rho_s \) and water \( \rho_w \). Considering that the principal stresses are either close to vertical or horizontal, the horizontal stress at the wedge tip \( \sigma_h = \sigma_1 \), with \( \sigma_1 \geq \sigma_2 \geq \sigma_3 \), is:

\[
\sigma_h = \sigma_1 = (\rho_s - \rho_w)g\Delta h. \quad (1)
\]

In the sedimentary cover the vertical stress \( \sigma_v = \sigma_3 \) is:

\[
\sigma_v = \sigma_3 = \rho_wg\Delta h + \rho_sgh. \quad (2)
\]

The Mohr–Coulomb relation between the principal stresses \( \sigma_1 \) and \( \sigma_3 \) and the angle of friction \( \phi \) in compression being (for details, see Appendix 2 in Nalpas and Brun, 1993):

\[
\sigma_1 = K\sigma_3, \quad (3a)
\]

where

\[
K = (1 + \sin \phi)/(1 - \sin \phi), \quad (3b)
\]

the combination of equations (1)–(3) gives:

\[
\Delta h = K\rho_s h/|\rho_s - \rho_w(K + 1)|. \quad (4)
\]

Equation (4) shows that in absence of water \( \Delta h = Kh \). In laboratory experiments where the viscous layer representing the salt is entirely covered by a constant thickness layer of sand with a frictional angle \( \phi = 30^\circ \), \( \Delta h = 3h \). The experiments (Figs. 5 and 6) effectively show that failure occurred for an overburden that was always more than three times thicker (\( \Delta h \)) than the initial sand cover thickness (\( h \)).

In addition, for \( \Delta h > 0 \) in equation (4), the right hand term must also be positive:

\[
\rho_s - \rho_w(K + 1) > 0, \quad (5)
\]

or:

\[
K < (\rho_s/\rho_w) - 1. \quad (6)
\]
Condition (6) is verified only if the sediment density \( \rho_s \) is high and/or the frictional angle \( \phi \) is low. In agreement with this, available numerical models of salt tectonics driven only by sedimentary loading tend to use high sediment densities and low frictional angles (\( \phi_{\text{eff}} \)) — e.g. \( \rho_s = 2300 \text{ kg m}^{-3} \) and \( \phi_{\text{eff}} = 16.42^\circ \) in Gemmer et al. (2005) and \( \rho_s = 2500 \text{ kg m}^{-3} \) and \( \phi_{\text{eff}} = 5^\circ \) in Gradmann et al. (2009). Low values of the effective angle of friction \( \phi_{\text{eff}} \) assume high to very high fluid pressures. According to the models, \( \phi_{\text{eff}} \) is taken as a simple reduction of the angle of friction \( \phi \) — e.g. \( \phi_{\text{eff}} = 20^\circ \) in Ings et al. (2004) — or it is calculated using an arbitrary value of fluid pressure \( \lambda - \) e.g. \( \sin \phi_{\text{eff}} = (1 - \lambda) \sin \phi \) where \( \lambda = p_f/\sigma_w \)

\( p_f \) is the pore pressure ratio of Hubbert and Rubey (1959), in Gradmann et al. (2009). Using the recent compilation of sediment density made in the Gulf of Mexico (Fig. 7) (Hudec et al., 2009), it is possible to calculate that the mean density of a sedimentary column is around 2000 and 2220 kg m\(^{-3}\) at depths of 1000 and 4000 m, respectively. For \( \rho_s = 2200 \text{ kg m}^{-3} \), equations (3b) and (6) give \( \phi \geq 5.2^\circ \). Taking a slightly higher value of \( \phi = 6^\circ \), equation (4) gives \( \Delta h = 30.8 \text{ h} \). This indicates that extremely thick amounts of sediments are required to stimulate spreading above a horizontal salt basin that is entirely covered by sediments under the sea level, even if the frictional angle is very low. Moreover, as the shear traction at the sediment—salt interface is not taken into account in the above calculations — i.e. salt is considered as an inviscid fluid — the sedimentary loading \( \Delta h \) must be even larger than the calculated value. In agreement with this conclusion, it is noteworthy that the most recent numerical models consider an initial salt basin to either not be covered by sediments at all or to be covered by an extremely thin sedimentary cover (few 10 m thick). An interesting exception is found in Gradmann et al. (2009) where the salt is initially covered by 4.5 km of sediments but with an extremely high sediment density \( \rho_s = 2500 \text{ kg m}^{-3} \) and an extremely low coefficient of friction \( \phi_{\text{eff}} = 5^\circ \).

Before further considering geological arguments, the above results and comments show that salt tectonics driven only by differential sedimentary loading is a process that is rather difficult to activate. It requires very thick sedimentary overburden and/or anomalously high sediment densities and/or very strongly reduced sediment friction — i.e. high fluid pressure values.

4.3. The hypothesis of a seaward free boundary

As mentioned above, analogue experiments that consider a salt basin as being entirely covered by sediments (Figs. 5 and 6) confirm that in absence of water, the sedimentary overburden must i) be at least three times thicker than the initial cover — i.e. \( \Delta h = 3 \text{ h} \) — and ii) display a deformation pattern, contractional at the sedimentary

Fig. 7. Variation of sediment density with depth in the Gulf of Mexico. Modified after Hudec et al. (2009).

Fig. 8. Sections of pure spreading models simulating the progradation of a sedimentary wedge on top of a salt basin initially covered by an extremely thin sedimentary cover — i.e. free frontal boundary (modified after Vendeville, 2005). The series of sections shows the progressive development of rafts and diapirs inside the wedge and the progressive extrusion of a salt nappe at the front.
wedge tip and extensional inside the wedge, that migrates forward with sedimentary progradation.

Several experiments have been carried out to simulate the differential sedimentary loading of a salt basin only covered by an extremely thin layer of sediments, assuming implicitly that the seaward end of the salt basin is a free boundary (Fig. 9 in Rowan et al., 2004, Figs. 4–6 in Vendeville, 2005; Gaullier and Vendeville, 2005). In such experiments, the sedimentary wedge, as soon as it is deposited, starts to spread with the development of distributed extension (Fig. 8). The wedge sand layer is therefore cut into several rafts that progressively separate with ongoing sedimentation. Each new sand deposit leads to proportional amounts of raft subsidence and separation. Consequently, triangular shape diapirs progressively grow between the rafts as long as a source layer is available. At the model front, outside the sedimentary wedge, the layer of silicone putty undergoes a rather homogeneous shortening, as indicated by the crenulation of the thin sand cover. Its proportional thickening leads to a progressive frontal overflow, forming a ductile nappe (Fig. 8c and d). A similar result was also obtained numerically by reducing the sedimentation rate (see Model 4 “sediment starvation” in Gemmer et al., 2005). These models show that extensional diapirs can form inside a sedimentary wedge if i) the salt basin is not initially covered by sediments (also see the models in Figs. 4–6 in Vendeville, 2005) and ii) enough time separates each sedimentation interval (“sediment starvation”). Reciprocally, high and constant sedimentation rates forbid the development of diapirs within the sedimentary wedge, even when the salt basin is not initially covered by sediments (see models 1–3 in Gemmer et al., 2005). Finally, when a significant thickness of sediments is deposited on top of the salt layer, prior to the arrival of a prograding sedimentary wedge, extensional diapirs cannot form inside the wedge because extension, which is restricted by cover shortening at the wedge tip, never reaches amounts large enough to allow diapirism to take place, as exemplified in Figures 5 and 6.

Numerical models that take into account the presence of seawater also consider salt basins covered by only few 10s of metres of sediments prior to the deposition of the prograding sedimentary wedge (Gemmer et al., 2005) (Fig. 9). In such models, the frontal boundary can be considered as free, as stated by Ings and Shimeld (2006). Such an initial condition, which is considered necessary to allow the development of minibasins (Gemmer et al., 2005), is somewhat difficult to reconcile with the fact that in these models, the water depth is around 4500 m at the onset of sedimentation. How could a thick salt basin have reached a depth of around 4500 m without receiving more than few 10s of metres of sediments while in addition, keeping a horizontal position?
5. Dominant gliding models: salt tectonics primarily driven by margin tilt

5.1. Laboratory experiments

The concept of salt tectonics driven by dominant gliding down an inclined basal surface of salt–sediment two-layer systems was mostly inspired by studies carried out in the South Atlantic (Western Africa: Duval et al., 1992; Spathopoulos, 1996; Marton et al., 2000; Tari et al., 2001; Brazil: Cobbold and Szatmari, 1991; Demercian et al., 1993) and North Sea (Nalpas and Brun, 1993; Coward, 1995; Stewart and Coward, 1995). Different types of laboratory experiments (Cobbold and Szatmari, 1991; Gaullier et al., 1993; Mauduit et al., 1997, 1998; Mauduit and Brun, 1998; Fort et al., 2004a,b; Brun and Mauduit, 2008, 2009) or numerical modelling (Ings et al., 2004; Albertz et al., 2010; Quirk et al., 2011) have been carried out to characterise the structures produced by gliding above salt.

In the dominant gliding experiment shown in Figure 10, a thin sand layer is deposited on top of a double-wedge shaped layer of silicone putty with a length l, representing a salt basin (Fig. 11). The model is then inclined at an angle \(\alpha\) to simulate the tilting of a passive margin during post-rift thermal subsidence (Fig. 3c). As soon as it is inclined, the sand-silicone two-layer model starts to glide downdip. The model displays extension updip and contraction downdip, giving the typical succession of structural domains observed at many passive margins (e.g. Fig. 1). In the downdip direction, these domains are primarily: i) sealed tilted blocks, grabens, rollovers and diapirs in the updip extensional domain and ii) folds, thrusts and contractional diapirs (squeezed silicone-cored anticlines) in the downdip contractional domain. All these structures develop synchronously with sedimentation and therefore, they display a growth character in extension – e.g. rollovers – as well as in contraction – e.g. growth synclines. Contraction occurs due to a downdip constraint – e.g. edge of the silicone and/or decreasing slope – and starts at the downdip end of the salt basin, progressive shortening migrates in the updip direction, ultimately reaching the extensional domain and inverting in contraction structures that were previously extensional – e.g. squeezed diapirs (Fig. 10a). In addition, it can be noted from the top view taken at 20 h that the initial contraction zone does not directly initiate against the downdip salt pinch-out but at a small distance from it. Consequently, during progressive deformation, contraction also migrates downdip down to the salt pinch-out, forming a frontal thrust. More details on this type of experiments can be found in Brun and Fort (2004) and Fort et al. (2004a,b).

The formation of frontal salt nappes is also obtained in this type of experiment when the sedimentation rate in the downdip contractional domain remains moderate (Fort and Brun, 2011). In the model shown in Figure 12, the contractional domain starts to develop (Day 6) in the same way as in Figure 10. With ongoing shortening, the squeezing of salt-cored anticlines produces a series of small zones of salt extrusion whose progressive connexion leads to the complete failure of the contractional front and the formation of a large salt nappe (Day 10). The domain that was previously underformed, located on the updip side of the contractional domain, is then submitted to a backward extension that reaches the previous domain of the updip extension. This illustrates that when salt nappes form after a long period of frontal contraction, the salt that thickened in the downdip contractional domain and which is suddenly liberated by the frontal failure allows the whole downdip domain to spread seaward at a fast rate on top of the abyssal plain. In this process, salt is massively extruded in two ways: i) by the frontal salt nappe and ii) by diapirism in the backward zone of extension.

5.2. Stresses at failure

When it is inclined at an angle \(\alpha\), the whole salt–sediment two-layer system tends to glide downslope, creating a contractional state of stress in the vicinity of the downdip salt pinch-out. But gliding can start if the resistance opposed by the sedimentary cover is overcome, leading to folding and/or thrusting. Considering that the principal stresses are either close to vertical or horizontal, the maximum horizontal stress \(\sigma_1\), with \(\sigma_1 \geq \sigma_2 \geq \sigma_3\), exerted by the two-layer system at the downdip end of the salt basin is:

\[
\sigma_1 = \sigma_1 = (\rho_s + \rho_e)gl \sin \alpha. \tag{7}
\]

where \(\rho_e\) is the density of salt.

The vertical stress at the base of the sedimentary cover \(\sigma_v = \sigma_3\) is:

\[
\sigma_v = \sigma_3 = \rho_sg l \sin \alpha + \rho_sgh_s. \tag{8}
\]

With the Mohr–Coulomb relation between \(\sigma_1\) and \(\sigma_3\) in compression \(\sigma_1 = K\sigma_3\), the combination of equations (7) and (8) gives the critical value of the angle \(\alpha\) necessary to reach failure as a function of cover thickness \(h_s\), salt basin length \(l\) and densities \(\rho_s\), \(\rho_e\) and \(\rho_W\) of the materials involved:

\[
\alpha = \arcsin(h_s/l)(K\rho_s/(\rho_s + \rho_e - K\rho_W)). \tag{9}
\]

Figure 13 is a graphical illustration of equation (9) showing \(\alpha\) as a function of basin width \(l\) for different values of sedimentary cover thickness \(h_s\) using typical density values (\(\rho_s = 2000\ \text{kg} \cdot \text{m}^{-3}\), \(\rho_e = 2200\ \text{kg} \cdot \text{m}^{-3}\) and \(\rho_W = 1000\ \text{kg} \cdot \text{m}^{-3}\)). The two sets of curves correspond to a sedimentary cover both without (in red: \(\phi = 30^\circ\) – i.e. \(K = 3\)) and with (in blue: \(\phi_{\text{eff}} = 20^\circ\) – i.e. \(K = 2\)) fluid pressure, respectively. The diagram shows that gliding occurs for tilt angles lower than \(1^\circ\) for basin widths in the range of 200–600 km and sedimentary cover thicknesses up to 1 km, even in absence of fluid pressure. With a fluid pressure \(\lambda = 0.31\) giving \(\phi_{\text{eff}} = 20^\circ\) and \(K = 2\), even if a 2 km thick sedimentary cover is present, gliding occurs for tilt angles of \(1^\circ\) or less. In other words, gliding is a very efficient
process of salt tectonics that in most common geological cases can be easily stimulated by margin tilt angles smaller than 1°, even in the absence of abnormal fluid pressure.

5.3. The effect of sedimentary loading on gliding

Sedimentary loading is not a necessary condition of salt tectonics at tilted margins because the salt layer can even start to flow downdip if tilting occurs before the deposition of the sedimentary cover. However, as the sedimentary pile thickens the sedimentary pile above the salt layer, the vertical stress increases and tends to accelerate the rate of gliding, although only slightly, as illustrated in experiments by Mauduit et al. (1997) and Fort et al. (2004b). The rate of gliding primarily depends the differential height from updip edge to downdip edge of the basin and not on layer thickness (eq. (7)).

5.4. Dwarf and giant examples of dominant gravity gliding above salt

The Needles area of the Canyonlands National Park in Utah provides a 12 km wide natural model of gliding above a sequence of evaporites forming a décollement layer with a 1–2° dip, on top of which a stepped series of grabens has formed within a 460 m thick brittle layer of upper Palaeozoic sandstone and shale, in response to the incision of the Colorado River (McGill and Stromquist, 1979; Trudgill and Cartwright, 1994). In this small-scale example of dominant gravity gliding without frontal buttress, the sedimentary layer deforming on top of the salt décollement is entirely submitted to extension down to the Colorado River.

On planet Mars, over a distance of approximately 3000 km, the superficial cryosphere of the Thaumasia plateau displays a domain of frontal contraction and a domain of backward extension interpreted as a “mega-slide” on top of the mega-regolith made of ice and impure salts dipping approximately 1° due to arching above a mantle plume (Montgomery et al., 2008).

Interestingly, these two examples are one order of magnitude smaller (dwarf) and larger (giant) than the dominant gravity gliding systems observed at passive margins. A supplementary loading was not added during gliding in either cases.

5.5. Dominant gliding in the North Sea and above the Messinian salt in the Mediterranean

The Zechstein salt basin of the North Sea was submitted to regional-scale extension from the Triassic to the late Jurassic (Bishop et al., 1995). Basement normal faulting was responsible for a local salt basin tilt of 1–3° that induced dominant gliding of the salt-sedimentary cover with synsedimentary updip extension and
diapirism and downdip folding and thrusting (Nalpas and Brun, 1993; Stewart and Coward, 1995). In the passive margins considered in the present paper, the margin tilt occurs mostly after salt deposition, as a consequence of thermal subsidence following lithosphere stretching and thinning — i.e. post-rift salt basins. In the North Sea, the salt basin is pre-rift and, contrary to Atlantic-type passive margins, it is not tilted as a whole but is locally affected by the tilting of underlying basement blocks. Nevertheless, North Sea salt tectonics display comparable organisation in extensional and contractional domains controlled by dominant gliding above a basal slope.

In the Mediterranean, the deposition of Messinian salt occurred in two main stages: first between 5.75 and 5.60 Ma, marked by a sea level fall of a few hundred metres, and second, between 5.60 and 5.32 Ma, marked by a sea level fall possibly as much as 1500 m (Clauzon et al., 1996). Despite their young age and local regional tectonic effects, the Messinian evaporites and Pliocene–Pleistocene sedimentary cover display extensional—contractional salt tectonic systems that obviously result from dominant gliding above very low angle basal slopes (Cyprus Arc: Bridge et al., 2005; Levant Sea: Cartwright and Jackson, 2008; Gradmann et al., 2005; Nile Delta: Loncke et al., 2006). The origin of tilting is not formally identified but it could likely result from lithosphere flexure due to a reloading of the Mediterranean lithosphere by seawater and salt.

The salt tectonic examples briefly described above (Canyonlands National Park, Utah; Thaumasia plateau, Planet Mars, North Sea; Messinian salt, Mediterranean) show that dominant gravity gliding that corresponds to a slope instability is i) not restricted to tilted passive margins and ii) can occur at various scales, from several tens to several thousands of kilometres.

5.6. Duration of margin tilt effectiveness

Once created, margin tilt remains and, as long as salt is present, dominant gliding can be effective. The effectiveness of gliding depends on the thickness of the salt. In a salt layer with thickness $h$ and viscosity $\mu$ undergoing a simple shear with a gliding velocity $v$, the shear strength is $\tau = \mu v / h$. Therefore, the salt layer strength increases as a direct function of the salt layer thinning. In other words, dominant gliding slows down progressively with salt thinning but the salt layer almost never disappears (Waltham, 1996; Fort and Brun, 2005). On the other hand, the system tends toward equilibrium at the margin scale as the system surface is approaching the horizontal and dominant gliding velocity approaches zero. This is likely the case for the section shown in Figure 2. However, it is interesting to note that active updip and downdip faulting and downdip folding affect the topography of the seabed, indicating that the system is still not in equilibrium 110 Ma after salt deposition.

6. Comparison between pure spreading and dominant gliding models

The two previous sections show that pure spreading and dominant gliding are two extremely different driving mechanisms of salt tectonics at passive margins, both in dynamic terms as well as from the point of view of their structural response. It is therefore useful to summarise here the most striking dynamic (Table 1) and structural (Table 2) features that help distinguishing between the two types of models.

6.1. Setting of models

Pure spreading driven only by differential sedimentary loading considers a salt margin that is not affected by tilting (Figs. 3b–9). Models show that even with a free frontal boundary — i.e. salt not covered by sediments — in such circumstances, dominant spreading requires significant water depths to allow the deposition of a several kilometre-thick sedimentary wedge, high sediment densities to increase loading for a given sediment thickness and significant reduction of sediment friction, assuming high fluid pore pressures, in order to decrease the strength of the sediments (Table 1).

Dominant gliding corresponds to a slope instability stimulated by margin tilt. It does not require any other particular condition in terms of water depth, sediment density or friction. Even in the absence of sediments that would increase the driving force, the entire salt basin flows downdip.

Sedimentation and deformation are strongly related in both model types, but in pure spreading, the sedimentation drives deformation whereas in dominant gliding, the sedimentation is trapped inside lows created by the deformation.

6.2. Structural response

The location of extension and contraction as well as their migration in time and space at the salt basin scale differ significantly (Table 2). In pure spreading, extension occurs within the prograding wedge and contraction occurs at the wedge tip; the latter only occurs if the salt basin is initially fully covered by sediments. Both extension and contraction migrate seaward with sedimentary progradation. Consequently, extension can lead to the inversion of previously contractual structures. In dominant gliding, extension occurs updip and contraction downdip. The contraction migrates updip, progressively reaching the domain of updip extension and leading to the contractional inversion of previously extensional structures — e.g. squeezed diapirs.

The formation of salt nappes at the basin front can occur progressively in both pure spreading (Fig. 8) and dominant gliding (Figs. 1 and 2) or can be sudden, but only in gliding (Fig. 12). Sudden frontal failure that gives birth to a salt nappe generally follows a long
period of frontal contraction and salt thickening (sometimes called “salt inflation”; e.g. Hall, 2002). Simultaneously, the downdip domain that was previously contractual is submitted to extension (Fig. 12).

7. Geological testing of models

7.1. Applicability of models to nature

A salt tectonic model, either experimental or numerical, is based on a particular concept (working hypothesis) of the physical processes involved. The model has its own coherence that links any combination of the input parameters to a specific structural response with two important implications. The model response to variations in the input parameters i) helps to understand the mechanical behaviour of the system defined by the working hypothesis and ii) can be compared to natural examples in order to test the applicability of the working hypothesis to nature. In other words, a model does not directly demonstrate that a given process exists in nature. It is only a tool to study the dynamics of a system based on a physical concept of a natural system.

There are two different concepts of salt tectonics leading to two different categories of models (Fig. 3, Table 1) with significantly different types of responses (Table 2). To apply any of these models to nature, it is therefore critical to examine the geological relevance of the initial and boundary conditions used for building each model type. In particular, they are: i) the initial salt basin size and geometry, ii) the margin tilt, iii) the presence or absence of an initial sedimentary cover on top of salt, iv) the water depth and v) sediment physical properties, particularly density and strength. As reviewed in Section 2, well and seismic data often provide the necessary information to set up the initial conditions (i, ii and iii). Water depth (iv) is more conjectural. As discussed in Section 2, the deposition of thick evaporites would normally require shallow water conditions. This is in agreement with the initial conditions for gliding but not with those for spreading. The activation of spreading by differential sedimentary loading requires an extremely thick sedimentary overburden that fixes a minimum value for water depth (approximately 4500 m in most numerical models). This is rather unrealistic as the salt basin should have reached a water depth of approximately 4500 m without margin tilt before the onset of sedimentation. The density of sediments (v) is most often available from well data. Sediment friction (v) is rather well known from laboratory measurements but the effective friction also depends on pore fluid pressure, something that can be extremely variable and generally poorly known at the basin scale.

7.2. Systems and sub-systems

Models show that a salt basin submitted to gravity-driven deformation in spreading or gliding is a dynamic system that can be divided into extensional and contractional sub-systems (Figs. 5, 10 and 12) that can in turn be divided into sub-sub-systems such as tilted blocks, rollovers, diapirs, etc. (Fig. 10). As detailed above, both the pure spreading and dominant gliding models display a number of similar growth-like structures such as rollovers, extensional and contractional diapirs, minibasins, folds, thrusts or salt nappes. These structures, either considered individually or grouped together, can provide a direct diagnostic of pure spreading versus dominant gliding. Other structures are rather ubiquitous and are not specific to any type of model. More generally, it is the organisation of the structure in space and their evolution in time that gives the whole basin scale that can be used to discern between pure spreading and dominant gliding, as illustrated by basin-scale experiments (Figs. 5, 10 and 12; Table 2): i) distribution in domains of extension and contraction, ii) migration of the extensional and contractual domains in space and time and iii) type of structural inversion to which the migration of extension and contraction might lead. Therefore, the identification of the driving mechanism of salt tectonics in a given basin — i.e. pure spreading vs. dominant gliding (Table 1) — requires a study of the relationships between

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<th>Model</th>
<th>Pure spreading</th>
<th>Dominant gliding</th>
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### Table 1
Comparison of pure spreading and dominant gliding models considering dynamic parameters, initial conditions and the place and role of salt and sediments in the process.

<table>
<thead>
<tr>
<th>Model</th>
<th>Pure spreading</th>
<th>Dominant gliding</th>
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### Table 2
Comparison of pure spreading and dominant gliding considering the patterns of deformation observed in analogue and numerical modelling.

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<th>Model</th>
<th>Pure spreading</th>
<th>Dominant gliding</th>
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sedimentation and deformation at the whole salt basin scale (Table 2). Whole basin cross-sections showing the evolution of deformation since salt deposition are essential tools for this purpose. Unfortunately, not enough well documented and updated cross-sections of this type are yet available for most passive margins.

8. Diagnostic structures of pure spreading vs. dominant gliding

8.1. Sealed tilted blocks

Many margins display wide series of small tilted blocks in the updip part of the margin, close to the salt pinch-out (Fig. 14a). These structures, which are well reproduced in gliding models (Fig. 14b), cannot result from differential sedimentary loading because the thickness of sediments on top of salt at their time of development is too small to be able to activate pure spreading. As block size is a direct function of sedimentary thickness (Brun and Fort, 2008), this deformation pattern clearly indicates that deformation started at very early stage of deformation and sedimentation. In addition, because all normal faults dip in the same sense as the basal sole, salt flow occurred with a strong component of downdip shear. The sealing of blocks by ongoing sedimentation demonstrates that the salt layer has thinned enough to allow stabilisation of the shelf. Consequently, sealed tilted blocks provide an excellent diagnostic feature of dominant gliding.

8.2. Early frontal contraction

Evidence for early contraction in the downdip part of the salt basin (Figs. 1 and 2) is the best evidence that demonstrates

![Fig. 14. Sealed tilted blocks in nature (Angolan margin) (a) and experiment (b). Modified after Brun and Fort (2008).](image1)

![Fig. 15. Structures that provide a diagnostic of pure spreading vs dominant gliding even considered individually. Sealed tilted blocks in the updip margin and early folds and thrusts in the downdip margin are excellent diagnostic of dominant gliding (a). Landward (updip) migration of contraction is diagnostic of dominant gliding (b). Seaward migration of contraction is diagnostic of pure spreading if the salt basin is initially covered of sediments (c).](image2)
dominant gliding (Fig. 15a), even when considered by itself. Laboratory experiments (Figs. 10 and 12) show that they are the first structures to develop at the basin scale, prior to updip extension.

8.3. Migration of contraction

In dominant gliding, the progressive downdip thickening of salt and sediments leads to the updip (backward) migration of contraction, toward the extensional updip domain. This results in the inversion of previously extensional structures – e.g. squeezed diapirs – (Figs. 10 and 15b).

In pure spreading, when salt is initially covered by sediments, both extension and contraction migrate seaward with the sedimentary progradation (Figs. 5 and 15c). During migration, the extensional domain progressively reaches domains that were previously contractional. This can result in extensional inversions of previously contractual structures (Fig. 6).

These contrasted migration patterns of extension and contraction and related structural inversions provide a good diagnostic of dominant gliding (Fig. 15b) versus pure spreading (Fig. 15c).

9. Diagnostic of pure spreading vs. dominant gliding provided by structure associations

9.1. Early rafting

Rafting of the thin sedimentary cover deposited directly on top of salt at the landward basin end indicates that extension started at an early stage of the salt basin evolution (Burollet, 1975; Duval et al., 1992). Because the sedimentary thickness is low, it could be tempting to directly conclude that differential sedimentary loading cannot be responsible for the deformation observed. However, rafting is obtained by differential sedimentary loading at the early stages of spreading experiments when the salt basin is not overlain by an initial sedimentary cover of significant thickness – i.e. frontal free boundary (cf. Section 4.3) – (Fig. 8). This shows that early rafting and simultaneous diapirism alone cannot be used to tell the difference between pure spreading and dominant gliding. Therefore, an essential additional test consists in examining if sediments of the same age are present or not at the seaward end of the salt basin (Fig. 17b). A lack of sedimentary cover thicker than a few tens of metres would favour the hypothesis of pure spreading driven only by differential sedimentary loading. The presence of a sedimentary cover, especially if it is affected by contractional deformation, would argue in favour of dominant gliding stimulated by margin tilt (Fig. 17a).

9.2. Diapirism

Diapirism results from the rise of salt in the footwalls of normal faults (Vendeville and Jackson, 1992; Quirk and Pilcher, 2011) where the sedimentation rate remains low (Fig. 16a and b). In the absence of sedimentation, salt can even reach the surface and flow horizontally on top of the bounding rafts.

In dominant gliding, diapirs generally form in the downdip part of the extensional domain where sufficient salt is present (Figs. 1, 10 and 12). According to the models (see Section 4), two situations must be considered in pure spreading. In the absence of an initial sedimentary cover, the prograding sedimentary wedge can spread and form rafts allowing the development of diapirs (Fig. 8). This explanation for diapirism requires that the frontal basin boundary is free and therefore, the same test as that used for rafting can also be used here (Fig. 17b). If an initial sedimentary cover overlies the whole salt basin, extensional diapirs cannot form inside the prograding wedge because of insufficient extension (Fig. 6). Therefore, diapirism requires the analysis of synchronous deformation in the downdip domain to discriminate between pure spreading and dominant gliding.

9.3. Frontal salt nappes

Many passive margins show salt nappes that formed at the seaward front of salt basins. According to models, they can develop in two ways: progressive or sudden.
In pure spreading models in which the salt basin is not initially covered by sediments, the salt nappe develops progressively as the amount of salt flowing out of the initial limits of the salt basin is a direct function of the amount of sediments deposited on top of the salt basin. This is well illustrated in Figure 8 (also see models 1–3 in Gemmer et al., 2005). In pure spreading models where salt is initially covered by sediments, salt nappes do not form easily (see the models in Figs. 8–13 in Gemmer et al., 2004).

In dominant gliding models, the development of frontal salt nappes is more generally sudden, as illustrated in Figure 12. During a long first stage, the downdip part of the salt basin undergoes contraction and salt inflation. When the frontal failure occurs, the thickened salt flows rapidly on top of the abyssal plain and carries rafts of the sedimentary cover. The whole downdip part of the basin that was previously contractional is submitted to extension.

The two models apparently produce similar salt nappes but in pure spreading, nappes develop progressively and only if the seaward part of the salt basin is not covered by sediments whereas in dominant gliding, salt nappes form suddenly and involve the sedimentary cover. These differences can be used to discriminate between pure spreading and dominant gliding.

10. Ubiquitous structures

10.1. Minibasins

In their review of salt tectonics in the Gulf of Mexico, Worrall and Snelson (1989) introduced the general term “minibasin” to identify small sedimentary basins only a few tens of kilometres wide that have deeply subsided into salt. The term was so appealing that it has been further used in most salt basins, in various extensional or contractual settings. However, as argued by Hudec et al. (2009), this explanation would hold true only if the average density of the basin fill exceeds that of salt, which in the Gulf of Mexico needs at least 2300 m of siliciclastic fill to ensure enough compaction. Hudec and his co-workers identified several alternatives to the density-driven subsidence of minibasins, including extension, contraction, topographic–bathymetric effects or sub-salt flow. In other words, minibasins do not represent a category of structures with a single mechanical origin and they are not characteristic of early salt tectonics driven by differential sedimentary loading.

10.2. Rollovers and stepped counter-regional systems

Contrary to diapirs, rollovers result from a permanent filling up of the surface depression created by extension between rafts (Mauduit and Brun, 1998) (Fig. 16c). They can be either synthetic or antithetic according to the seaward or landward dip of the main growth fault, respectively (Brun and Mauduit, 2008). Both synthetic and antithetic rollovers develop in pure spreading (Fig. 6; see also models 1–6 in Gemmer et al., 2005) as well as in dominant gliding models (Fig. 10).

Ge et al. (1997) have experimentally shown that rollover-like structures could result from differential sedimentary loading without extension. Figure 18 shows a rollover that developed by the frontal expulsion of salt without extensional faulting of the initial sedimentary cover. The initial sedimentary cover undergoes a migrating rolling-hinge flexure at the tip of the sedimentary progradation, accommodating forward salt expulsion. In this experimental model, the overburden forming the rollover is more than seven times thicker than the initial sedimentary cover and no sediment is deposited in front of it. Therefore, the applicability of this process to natural rollovers is extremely restrictive.
In the Gulf of Mexico, antithetic rollovers have been interpreted as structures resulting from the expulsion of salt by the differential sedimentary loading of a salt sheet by a prograding sedimentary wedge, called “stepped counter-regional systems” (Diegel et al., 1995; Rowan, 1995; Schuster, 1995; Hudec and Jackson, 2006) (Fig. 19). The footwall of the rollover is considered fixed in reference to the underlying basement and the concave-shape rollover base is interpreted as a salt weld instead of an antithetic normal fault. As stated by Schuster (1995) “Interpretation as a detachment fault would require northward extension and gliding of the overlying strata. However, there are no known toe structures updip to have accommodated such extension; furthermore, extension updip would violate the general rule of gravity tectonics”. This structure was therefore qualified as a “pseudo-extension” as neither of the two blocks has moved horizontally (Fig. 19a and b). In Figure 19a, the upward expelled salt is removed by salt dissolution. Figure 19b, displays a double sequence of expulsion and sub-horizontal salt flow giving a staircase-type salt weld. It is the application of the classical model of a listric fault separating two blocks (Fig. 20a) could be regarded as violating the general rule of gravity tectonics. However, in salt tectonics, the development of a rollover requires a three-block system: an updip block and a downdip block both moving downdip and that separate from each other on top of an extending decollement layer lying above a basement block (Brun and Mauduit, 2008). As shown in Figure 20b, an antithetic rollover can develop between two blocks moving downdip. Laboratory experiments have shown that a synthetic rollover can pass laterally to an antithetic one (see Fig. 9 in Brun and Mauduit, 2008), indicating that they developed synchronously between the two same rafts and that they both result from “true extension”. Consequently, the evidence for counter-regional or antithetic rollovers does not provide arguments either against extension or in favour of differential sedimentary loading.

In this regard, the spectacular antithetic rollover of the Cabo Frio structure in the Santos Basin of the Brazil margin (Fig. 21) (Demercian et al., 1993; Mohriak et al., 1995) is especially interesting. The Cabo Frio structure can be mapped along the strike for hundreds of kilometres but has been interpreted either as a result of differential sedimentary loading (Szatmari et al., 1996; Ge et al., 1997; Mohriak and Szatmari, 2001) or as an antithetic rollover related to downdip rafting (Mauduit and Brun, 1998). However, a recent detailed seismic study by Quirk et al. (2011) shows that the rollover development is controlled by the landward dipping Cabo Frio fault and that remnants of the Albian rafts are present at the rollover base. Quirk and his co-workers conclude that: “it is an extensional feature caused by salt draining downdip (seaward) as the passive margin tilted with thermal subsidence”.

11. What kind of exception is the Gulf of Mexico?

As stated in the introduction, for a long time now, the GoM has been a natural laboratory where many concepts of salt tectonics have been elaborated upon. The northern GoM displays a rather complicated situation with a massive extrusion of allochthonous salt bodies during the Cenozoic, as much as 8 km thick in some places, on top of which Neogene minibasins are deposited, sometimes very deeply. The presence of minibasins and stepped counter-regional systems (see Section 9), among other structures, have strengthened the impression that salt tectonics was mainly driven by differential sedimentary loading. For more than 30 years, salt tectonics in the GoM has therefore been considered to be different from what is observed and/or described in other margins.
Interestingly, Diegel et al. (1995, p. 145) concluded their review of the tectono-stratigraphic framework by asking the following question: “Are Gulf Coast style allochthonous salt structures more common, but unrecognized, or are the scale and complexity of salt-related structures in the Gulf Coast unique?”. It is beyond the scope of the present paper to enter into the details of the GoM salt tectonics. However, in the framework of this present paper, which is devoted to salt tectonic models and their applicability to passive margins, it seems necessary to pay some attention to this question that still remains: Is the GoM a unique case in salt tectonics? Or, said differently, what kind of exception is the GoM?

11.1. Salt flow at regional scale

We recently mapped the regional direction of salt flow (Fort and Brun, 2011) using two types of data (Fig. 22): i) shelf break migration and ii) seafloor deformation in the slope domain. For the purpose of the present paper, we will summarise below the major conclusions. The full data analysis can be found in Fort and Brun (2011).

The shelf break has migrated over approximately 400 km since the Cretaceous (Galloway et al., 2000). In a salt basin, a shelf can form and stabilise only if the salt layer underlying the sediments

Fig. 22. Kinematics of salt flow in the northern Gulf of Mexico as recorded by the shelf migration (up) and seafloor deformation in the slope (down) (modified after Fort and Brun, 2011). Successive positions of the shelf break are identified by their age in Ma. Thick lines represent transfer zones that principally developed from Miocene to present. In the slope domain (Bathymetric data from NOAA), orange arrows indicate the mean salt flow direction deduced in four domains from the statistical analysis of the senses of shear in the deformation zones separating the minibasins and yellow bars indicate the direction of stretching in zones of co-axial deformation.

Fig. 23. Line drawings of regional seismic sections showing the major structures in the slope area in the northern margin of the Gulf of Mexico (modified after Fort and Brun, 2011). (a) Keathley Canyon salt nappe. (b) Section of Walker Ridge showing the four-layer structure.
stops flowing (Fig. 14a) or is flowing at a very low rate. Therefore, at a given moment of basin evolution, a shelf break contour represents the boundary between the seaward part of the basin still deforming above the flowing salt and the landward part where salt has stopped flowing and where previous structures are sealed by sediments being deposited. If the rate of the salt flow varies on the seaward side of the shelf break, and in particular when sedimentary progradation is oblique to the salt flow direction, transfer zones develop parallel to the salt flow during shelf break migration. In the northern GoM, where sedimentary progradation is dominantly directed to the south (Galloway et al., 2000), transfer zones trend in a mean NE–SW direction indicating that salt has been flowing toward the SW since the Cretaceous (Fig. 22). In other words, there is a mean obliquity of around 55° between the directions of sediment progradation and salt flow in the northern GoM. Figure 22 shows that most transfer faults are well constrained by offsets of some tens of kilometres, and even reaching 80 km at few places.

In the slope area, the analysis of seafloor structures over a 450 × 250 km area has been carried out using Seabeam-type digital bathymetric data (NOAA). These structures mostly correspond to shear zones on top of salt ridges located between minibasins. They indicate relative displacements between minibasins. A detailed mapping of senses of shear with a resolution of approximately 100 m has been processed statistically, providing a mean direction of allochthonous salt flow for four domains covering the whole studied area (orange arrows in Fig. 22). They all indicate a SW directed salt flow in the slope domain bound by the shelf break to the north and the salt basin border to the southeast. In addition, local directions of stretching have been directly obtained from zones of co-axial deformation that also trend in a NE–SW direction (yellow bars in Fig. 22).

These different types of data (Fig. 22) indicate that salt flow was dominantly directed toward the SW since the Cretaceous.

### 11.2. Frontal salt nappe and backward salt extrusion

Recent seismic imaging in the slope area helped to understand the extrusion processes of allochthonous salt. The two nearly parallel sections trending NE–SW, shown in Figure 22, parallel to the mean direction of salt flow at the regional scale, illustrate the four-layer structure of the salt basin in the slope area. The overall structure is strikingly “flat” and, from base to top, consists of: (1) the middle Jurassic autochthonous Louann salt that pinches out to the southwest of Keathley Canyon, (2) a comprehensive pile of “lower” sediments, from the Jurassic–Cretaceous to the lower Miocene, which have locally been folded during sedimentation and later stretched, (3) a layer of allochthonous salt fed by diapirs cutting through the Cretaceous–Lower Miocene sediments in the Walker Ridge or a salt nappe extruded from a ramp, and (4) a layer of Late Miocene to Quaternary sediments forming minibasins.

The allochthonous salt nappe was mostly fed by a zone of salt extrusion above a frontal ramp, now located at the nappe rear. The nappe salt base rests on top of the Lower Miocene–Pliocene sediments at very low angle, over more than 100 km, indicating that salt was flowing very quickly during sedimentation.

In Walker Ridge, at the back of the salt nappe, the allochthonous salt layer was fed by a series of salt walls located between large Cretaceous to Early Miocene rafts that sunk down to the base of the autochthonous Louann salt. Salt was vertically extruded from a series of salt walls and then spread horizontally, forming convergent salt sheets on top of rafts. Synchronous sedimentation is cast in downward-tapering bodies whose apex corresponds to the sutures between convergent flowing sheets (5 in Fig. 23b). The flat attitude of the allochthonous salt base envelope, over distances larger than 100 km, indicates that the extruding salt in Walker Ridge has been able to flow along a near-horizontal surface. This implies that the salt flow was very fast and/or the sedimentation rate very low. Because the rate of sediment supply in the Early–Middle Miocene is in the medium range of values observed in the Gulf of Mexico (Galloway, 2001), it is more likely that salt flowed at a rather fast rate as observed for the Keathley Canyon salt nappe. Sedimentary units have been deposited on top of allochthonous salt since the Late Miocene but are dominantly of Pliocene–Pleistocene age and form minibasins with variable maximum depths.

For more than approximately 400 km and over 100 Ma, salt has flowed in a SW direction at a high angle to the mainly south-directed sediment supply in the area. The sudden onset of shelf migration in the Middle Miocene can be correlated with a frontal ramp failure in Keathley Canyon (Fig. 24), without an obvious temporal relationship with a particular sedimentary event (Galloway, 2001). The ramp failure to liberate salt flow at the basin front increased the rate of stretching and the thinning of the salt updip, triggering shelf migration. This sequence of events can be directly compared with the experiment in Figure 12 as well as the relationship between the formation of the frontal nappe (Fig. 23a) and backward extension and salt extrusion (Fig. 23b). The lack of an obvious temporal relationship with a particular sedimentary event and a direction of salt flow that is oblique to the direction of sediment supply excludes the idea that sedimentary loading could have been the main driving force of salt tectonics in the GoM, as generally advocated. The permanence of the SW directed salt flow, irrespective of the sedimentation rate, suggests that the salt tectonics in the northern GoM corresponds to dominant gliding controlled by a SW dipping margin. On the above basis, our answer to Diegel et al.'s (1995) question would be the following: If salt tectonics in the GoM can be considered as an exception, it is, before any other aspect, related to allochthonous salt and the dynamics of its extrusion from the mother Louann salt and not to differential sedimentary loading.

### 11.3. A methodological remark

Since several decades, salt tectonics in the GoM has been considered to be driven by differential sedimentary loading. This was suggested not only by high values of sedimentation rates, in particular in the Neogene, but also by some spectacular and intriguing local structures like deep minibasins or stepped counter-regional systems. But, as discussed above, minibasins do not result from subsidence under the weight of sediments (Section 9.2) and
“stepped counter-regional systems” do not provide arguments either against extension or in favour of differential sedimentary loading (Section 9.3). Therefore, it is noteworthy that in a region where local-scale structures are often complicated in 3D, a kinematic analysis at regional scale provides a rather consistent direction of salt flow with two major outcomes. First, this kinematic signal consistency at large scale (over 400 km) is such that it must be controlled by a physical cause at the same scale—i.e. dominant gliding stimulated by margin tilt. Second, from a practical point of view: i) the regional-scale direction of salt flow provides a dynamic reference for the interpretation of 3D structural complexities at local scales and ii) regional-scale cross-sections and section balancing that are essential tools to understand the tectonic history of the salt basin must be constructed parallel to the regional salt flow.

12. Conclusions

Two types of gravity-driven processes are invoked to explain salt tectonics at passive margins: i) pure spreading driven only by differential sedimentary loading and ii) dominant gliding primarily due to margin tilt (slope instability). Their comparative analysis using simple mechanics and available analogue and numerical models leads to the following conclusions.

(1) Dynamics:
- Pure spreading driven by sedimentary loading requires the following to be effective: i) a thick sedimentary overburden and therefore significant water depths, iii) high sediment density, iv) low frictional angles of the sediments (high fluid pore pressure), v) a seaward free boundary of the salt basin (salt not covered by sediments).
- Dominant gliding has no specific requirement to be effective provided the strength of the overburden can be overcome. It can occur for margin tilt angles lower than 1° for basin widths in the range of 200–600 km and for sedimentary cover thickness up to 1 km, even in the absence of abnormal fluid pressure.

(2) Role of salt and sediments
- Salt resists and sediments drive in pure spreading whereas both salt and sediments drive in dominant gliding.
- Sedimentation generates deformation in pure spreading and fills the bathymetric lows created by deformation dominant gliding.

(3) Deformation
- In pure spreading, extension is located inside and contraction at the tip of the prograding sedimentary wedge. Both extension and contraction migrate seaward with the sedimentary progradation. Migration of the deformation can create an extensional inversion of previously contractional structures
- In pure spreading, extension is located updip and contraction downdip. Extension migrates downdip and contraction updip. Migration of the deformation leads to a contractual inversion of previously extensional structures (e.g. squeezed diapirs)

To apply one of these two types of models to nature it is critical to previously examine the geological relevance of the initial and boundary conditions used for each type of model:

(4) Deformation, initial salt basin size and geometry, margin tilt and presence or absence of an initial sedimentary cover on top of salt can most often be checked using well and seismic data.

Reminder: the presence of an initial sedimentary cover on top of salt is a serious obstacle to pure spreading.

(5) Water depth. The deposition of thick evaporite layers would normally require shallow water conditions. This is in agreement with initial conditions for dominant gliding but not with those for pure spreading. The activation of spreading by differential sedimentary loading requires an extreme differential thickness of the overburden that fixes a minimum value for water depth (about 4500 m in most numerical models). This implies that the salt basin has reached a water depth of around 4500 m without tilt, before the onset of sedimentation. From a geological point of view, this is rather unrealistic.

(6) Sediment physical properties. Density is most often available from well data. Sediment friction is rather well known from laboratory measurements but the effective friction also depends on pore fluid pressure that can be extremely variable and generally poorly known at basin scale.

An analysis of the organisation of structures in space and their evolution in time at the scale of the entire basin is, in general, necessary to identify pure spreading from dominant gliding. However, few structures either considered individually or grouped together can provide a direct diagnostic of pure spreading versus dominant gliding.

(7) Diagnostic structures
- Good diagnostic structures of dominant gliding include sealed tilted blocks in the updip extensional domain and early folding and/or thrusting in the downdip contractional domain.
- Migration of the contraction is updip (landward) in dominant gliding and seaward in pure spreading.
- Early rafting associated to extensional diapirism could indicate pure spreading only if the seaward part of the salt basin is free of sediments. It indicates dominant gliding if it is synchronous with downdip contraction.
- Frontal salt nappes form progressively in pure spreading, as a direct function of the amount of sediments deposited on top of the salt basin, but only if the seaward part of the salt basin is free of sediments. They form suddenly, after a long period of downdip contraction, in dominant gliding.

(8) Ubiquitous structures
- Minibasins do not result from subsidence under the weight of sediments and do not represent a category of structures with a single mechanical origin. They cannot be used to identify pure spreading from dominant gliding.
- Both synthetic and antithetic rollovers develop in pure spreading as well as in dominant gliding. Antithetic rollovers that are so-called “stepped counter-regional systems” and qualify as a “pseudo-extension” do not provide arguments either against extension or in favour of differential sedimentary loading.

Acknowledgements

We are extremely grateful to Statoil for support over several years and for authorisation to publish this work. We benefited from a considerable freedom of action, large access to data and positive interaction with numerous people in Bergen, Houston and Oslo. We want to address our special thanks to Nicholas Ashton, Eric Blanc, Torbjørn Fristad, Rob Hunsdale, Bard Krokan, Tommy Mogensen, Egebjerg, Bjorn Rasmussen and Christian Zwach, for their confidence and help at various stages of this work. At Géosciences Rennes, we want to thank Frédéric Gueydan for stimulating...
does not deform. In dominant gliding and pure spreading the rock body is deformable.

Appendix 1. Displacements and strain in gravity-driven deformation

Salt passive margins display a broad spectrum of gravity-driven deformations even in the absence of extensional or contractional crustal-scale tectonics. In salt tectonics studies it has become rather common to consider that all types of deformations result from gravity spreading of layered salt–sediment systems. Consequently, the term “gravity spreading” progressively became understood as an equivalent (synonymous) of “gravity-driven deformation”. Beyond any semantic consideration, this shortcoming, which is partly true, is responsible for the obvious misunderstanding of the processes involved and their consequences in terms of strain and displacement.

Gravity-driven deformation can result either from the spreading under its own weight or the gliding down a slope of a deformable body. This is well known in glaciology where the gliding of glaciers down a slope (Ch. 6 in Paterson, 1972) is opposed, in terms of displacement and strain, to the spreading of ice caps above a horizontal rigid base (Ch. 9 in Paterson, 1972). In geology, similar approaches have been developed to establish the mechanics of nappes and thrusts driven by gravity (Ch. 7 in Price and Cosgrove, 1990; Merle, 1994). These situations are distinguished: sliding, dominant gliding and pure spreading. In sliding (Fig. 25a), a rock slab moves downslope along a décollement layer without significant internal deformation (e.g. Fig. 7 in Hubbert and Rubey, 1959). In dominant gliding (Fig. 25b), a slab moves downslope but deforms internally with a strong slab-parallel shear component. In pure spreading (Fig. 25c), the rock body flows under its own weight above a base that can be either horizontal or dipping opposite to the displacement (Ch. 9 in Ramberg, 1981). The term “dominant gliding” is used here instead of “gliding” because as the slab is deformable it necessarily undergoes a certain amount of spreading. This has been exemplified in laboratory experiments in which a single viscous layer (Brun and Merle, 1985) or a brittle—ductile two-layer system such as sediments lying on top of salt (Mauduit et al., 1997). In salt tectonics, gravity sliding can be omitted as salt—sediment systems undergoing gravity-driven deformation always display internal strain. In dominant gliding the maximum displacement can be large as it primarily depends on the slab length. In spreading, the maximum displacement is limited by the volume of material that must displace for the initially inclined surface to come to horizontal (Fig. 25c). Figure 26 illustrates the three types of strain patterns that characterise a viscous body undergoing true gliding (theoretical model) or a combination of gliding and spreading (dominant gliding), above an inclined base, and pure spreading, above a horizontal base. True gliding displays strong slab-parallel shearing whose intensity decreases upward (Fig. 26a). Pure spreading displays surface-parallel co-axial stretching in the upper part and base-parallel shear in the lower part whose intensity increases forward (Fig. 26c). In terms of strain pattern, a combination of gliding and spreading (dominant gliding) (Fig. 26b) is closer to true gliding (Fig. 26a) than to pure spreading (Fig. 26c). In summary, a major difference exists between pure spreading above a horizontal base and dominant gliding above an inclined base in terms of strain and displacement. Therefore, it is important to define two categories of salt tectonic models: i) pure spreading only driven by differential sedimentary loading and ii) dominant gliding primarily driven by margin tilt.

Schultz-Ela (2001) attempted to combines boundary conditions (base dipping or not) and rheology (slab deforming or not) in order to compare spreading and gliding: “The concepts of gravity gliding and gravity spreading are typically defined by the general rigidity and the orientation of an underlying detachment”. In using the term gliding instead of sliding, Schulz-Ela introduced a potential misunderstanding of the basic concepts concerned. The term “gravity sliding”, as used here in Figure 25a, was introduced and defined long ago by Schardt (1898) (quoted by Hubbert and Rubey, 1959). The sliding of a rigid block above an inclined décollement layer is obviously not applicable to salt tectonics. In addition, as pure spreading and dominant gliding have significantly different strain patterns (Fig. 26) a difference is required for the analysis of salt tectonics. Whether or not an inclined base is present is a decisive factor.

The term “salt drainage”, which has recently emerged (Davison et al., 2011; Quirk et al., 2011), describes the early flow of a salt basin non-covered by sediments. This corresponds to the dominant gliding of a viscous slab described above (Figs. 25b and 26b). A full analysis of strain variations in space and time can be found in Brun and Merle (1985).

Appendix 2. Analogue modelling of salt tectonics systems

For a small-scale model to be representative of a natural example (a prototype), a dynamic similarity in terms of distribution of stresses, rheologies, and densities between the model and the prototype is required (Hubbert, 1937; Ramberg, 1981). Following the above principle, it can be shown (Brun, 1999) that in sand-silicone models, the dynamic similarity is respected when the ratios of stresses and lengths are nearly equal. To represent the sediments, we use Fontainebleau sand with a density \( \rho \) of about 1400 kg/m\(^3\) and an angle of friction \( \phi \) in the range 30–31°, but without significant cohesion \( c \). To represent the salt, we use a silicone putty (transparent gum SGM36, Rhone Poulenc, France) with a Newtonian viscosity \( \mu = 10^4 \) Pa s and a density \( \rho = 1000 \) kg/m\(^3\). The density contrast between silicone putty and sand (\( \Delta \rho_s = 1.4 \)) is slightly higher than might be expected between salt and sediments in nature (\( \Delta \rho_s = 1.05–1.18 \) according to Weijermars et al., 1993). However, as pointed out by Weijermars et al. (1993) and Vendeville and Jackson (1992), this disparity is acceptable because the density contrast between salt and sediments is not the primary factor responsible for the rise of diapirs.
Fig. 26. Strain pattern (deformed grid) left and principal stretch $\lambda_1$ (right) with $\lambda_1 \geq \lambda_2 \geq \lambda_3$ in a viscous slab undergoing gravity-driven deformation: (a) pure gliding, (b) dominant gliding (gliding-spreading) and (c) pure spreading. Modified after Brun and Merle (1985).

References


