Numerical Modeling of Salt Tectonics on Passive Continental Margins: Preliminary Assessment of the Effects of Sediment Loading, Buoyancy, Margin Tilt, and Isostasy

Ings, Steven

Department of Earth Sciences

Dalhousie University

B3H 3J5 Halifax

Canada

email: <u>sings@dal.ca</u>

Beaumont, Christopher

Gemmer, Lykke*

Department of Oceanography

Dalhousie University

B3H 4J1 Halifax

Canada

*Current address:

School of Earth Sciences

The University of Leeds

LS2 9JT Leeds

United Kingdom

Abstract

Salt tectonics in passive continental margin settings is investigated using a 2D vertical crosssectional finite element numerical model of frictional-plastic sedimentary overburden overlying a linear viscous salt layer. We present preliminary results concerning the effects of sediment progradation over salt, buoyancy driven flow owing to density contrast between the sediment and salt, regional tilt of the salt layer, and local isostatic adjustment of the system. Sediment progradation causes a differential load on the underlying salt, which can cause the system to become unstable, leading to landward extension accommodated by seaward distal contraction. Slow progradation ($V_{sp} = 0.5 \text{ cm/yr}$) of slightly aggrading sediments gives a diachronous evolution comprising four main phases: 1) initiation of salt channel flow and the formation of mini-basins and associated diapirs; 2) onset of listric normal growth faulting and extension of the sedimentary overburden; 3) large-scale evacuation of the salt, formation of pre-rafts and rafts, and inversion of the mini-basins; 4) formation of a contractional allochthonous salt nappe that overthrusts the depositional limit of the salt.

Buoyancy effects are investigated using models with density contrasts between overburden and sediment of 0, 100 and 400 kg/m³. Although the lateral flow driven by differential loading dominates in all cases, the form of the mini-basins, the overall salt evacuation, and style of diapirism are sensitive to buoyancy forces, with the large density contrast producing the most developed diapiric and mini-basin structures.

A regional seaward tilt of 0.2° (of the type that may be produced by thermal contraction of the rifted margin) enhances and accelerates the overall seaward flow of the unstable slope leading to much earlier overthrusting of the distal depositional limit of the salt. The added downslope gravitational component also modifies the style of the mini-basins by enhancing the horizontal channel flow by comparison with the vertical buoyancy driven flow. This reduces the apparent efficiency of diapirism.

Local isostatic adjustment, owing to overburden and water loading, introduces a landward tilt of the system, thereby requiring salt to flow updip against gravity during evacuation. Isostasy also changes the overburden geometry and, therefore, modifies the stability and flow velocity of the extending overburden through the increased strength of the isostatically thickened proximal overburden, and through the modified differential pressure acting on the salt under these circumstances. The seaward flow of the unstable slope region is slower for the same overburden progradation velocity, more salt remains beneath the shelf as rollers and pillows during evacuation, counter-regional faults are more pronounced, and the allochthonous salt nappe progressively climbs above the isostatically adjusting sediments as it overthrusts.

Introduction

Evaporite rocks loosely termed 'salt' are much weaker than most other sediments, and this allows salt to become mobile and deform into the complex structures observed in many sedimentary basins. Salt tectonics occurs in many passive continental margin settings (*e.g.*, the Gulf of Mexico, numerous West African marginal basins, the margin offshore Brazil, and the margin offshore Nova Scotia), which have been studied through geological and geophysical observations (*e.g.*, Wu *et al.*, 1990; Demercian *et al.*, 1993; Ge *et al.*, 1997; Marton *et al.*, 2000; Rowan *et al.*, 2000; Cobbold *et al.*, 2001; Tari *et al.*, 2002; Shimeld, 2004) primarily owing to their potential for hydrocarbon exploration. They typically contain a seaward thinning sediment wedge, resulting from landward sediment influx. Moreover, they are characterized by a region of landward extension and a region of seaward contraction; which may be attributed to failure of the sedimentary overburden that accompanies the flow of the underlying salt under certain circumstances (*e.g.*, Worrall and Snelson, 1989; Koyi, 1996; Ge *et al.*, 1997; Rowan *et al.*, 2000; Vendeville, sub.).

Previous studies have discussed the mechanisms that drive salt tectonics, focusing on the buoyancy effects of the relatively low-density salt buried beneath denser sediments (*e.g.*, Talbot, 1992; Podladchikov *et al.*, 1993; Van Keken, 1993; Kaus and Podladchikov, 2001) as well as on differential loading of frictional sediments on top of the viscous salt (*e.g.*, Last, 1988; Vendeville and Jackson, 1992; Poliakov *et al.*, 1993; Jackson and Vendeville, 1994; Ge *et al.*, 1997; Gemmer *et al.*, 2004). For the latter case, two end-member styles of deformation may occur (Fig. 1). The differential load caused by the varying overburden thickness creates a pressure gradient in the viscous salt and causes a channel Poiseuille flow to form (Fig. 1A) (Turcotte and Schubert, 1982). With this end-member flow pattern, the overburden is stable and does not move horizontally. For larger differential overburden thicknesses, and

hence large pressure gradients, the overburden may become unstable and fail, initiating a Couette channel flow in the viscous layer. For the very unstable end-member (Fig. 1C), the system is dominated by the Couette flow, whereas for marginally unstable systems, the velocities may be characterized by a combination of Couette and Poiseuille flow (Fig. 1B).

Analog models have been used to investigate the mechanisms controlling salt mobilization beneath a brittle overburden (*e.g.*, Vendeville and Jackson, 1992; Koyi, 1996; Ge *et al.*, 1997; Cotton and Koyi, 2000; Vendeville, sub.). These studies successfully predicted complicated deformation structures similar to those observed in nature. In addition, recent studies show that analog models can simulate deformation during sediment progradation and aggradation over a viscous substratum (*e.g.*, Ge *et al.*, 1997; Cotton and Koyi, 2000; Vendeville, sub.), and therefore approximate equivalent depositional systems observed on many passive continental margins.

A comparatively smaller number of numerical model studies of passive margin salt tectonics have been published (*e.g.*, Last, 1988; Cohen and Hardy, 1996; Gemmer *et al.*, 2004). Last (1988) calculated the flow regime in a finite difference model of a sloping elasto-plastic layer overlying a viscous substratum. The numerical results for the end member where only the salt is mobile (like Fig. 1A) were in close agreement with the equivalent simplified analytical theory (Lehner, 1977) based on lubrication theory. The numerical model also demonstrated the Couette flow of the overburden (like Fig. 1C) leading to the prediction of deep extensional upslope basins and thrusting near the toe of the system. Cohen and Hardy (1996) used a numerical model to investigate salt withdrawal during differential loading by prograding continental margin sediments. The model was used to predict the large-scale stratigraphy for the end-member case where the sedimentary overburden only adjusts vertically during salt withdrawal (like Fig. 1A).

In our earlier research (Gemmer *et al.*, 2004) we expand on the results obtained by Last (1988) and Cohen and Hardy (1996). We use vertical section, plane-strain finite element models to study the stability of a uniform thickness linear viscous salt layer overlain by frictional-plastic sedimentary overburden that decreases in thickness from the proximal to distal limits of the salt, as shown in Figure 1.

The results from Gemmer et al. (2004) provide the basis for the work reported here; we therefore restate them briefly. The limiting stability of the overburden (i.e., the force balance at the transition to unstable overburden, the transition from Fig. 1A to 1B as a function of proximal and distal overburden thickness, the internal angle of friction of the overburden, and thickness of the salt layer were shown to be in good agreement with a modified form of Lehner's (2000) analytical theory for overburden failure caused by substratal creep. In addition, the initial velocity of the unstable overburden predicted by these numerical models is also in reasonable agreement with simplified analytical calculations based on thin-sheet approximations. Given the acceptable agreement between the numerical models and the analytical theory for the initial instability and the associated velocity distribution, Gemmer et al. (2004) also investigated the finite deformation and formation of salt-tectonic structures in a series of finite element experiments. The main focus was to investigate the sensitivity to the width of the salt layer and the role of the sedimentary overburden. The simplest model considered the non-physical case where the overburden has an initial thickness variation as described above, and shown in Figure 1. Such a model is non-physical because in a natural setting the overburden would become unstable during the sedimentation that is necessary to create this initial geometry, that is, before this configuration could be achieved. Other models considered the more natural progressive development of an unstable system as sediment progrades, with a specified finite velocity, across the model from the proximal (or landward) side of the system, forming a kinematically specified sedimentary prism that resembles a deltaic progradational wedge.

In this paper, we investigate the large deformation salt-tectonic structures that develop in similar plane-strain finite element model experiments. The results illustrate the potential of numerical models. The focus is on a partial assessment of the various processes that drive salt tectonics and the differences in the styles of deformation they produce. In particular, we examine the effect of sediment progradation, the roles of regional seaward tilt of the salt layer, and isostatic adjustment under the prograding sediment load. In these experiments we also use a kinematically specified model of sediment progradation. However, here we adopt a less aggressive, more aggradational than progradational sedimentation model, in which the rate of advance of the sedimentary profile is lower than those in Gemmer *et al.* (2004). This

approach allows the development of a richer range of salt tectonic structures. We also show results of a preliminary investigation into the role of the other important driving mechanism in salt tectonics, the buoyancy force that occurs because the salt has a lower density than that of the average overburden. This effect is shown in combination with the pressure driven instability for three cases; a normal density contrast, no density contrast, and a large density contrast between salt and overburden in which the salt is much less dense than the overburden. The results allow a partial assessment of the relative contributions of differential horizontal pressure and vertical buoyancy forces on the styles of salt tectonics within passive margin sedimentary prisms.

Numerical Model and Initial Conditions

Finite Element Model Formulation

The numerical model is a velocity-based plane-strain viscous-plastic finite element model (Fullsack, 1995; Willett, 1999) designed for large deformation fluid Stokes, or creeping, flows. The model solves the equilibrium force balance equations for incompressible flows in two dimensions

 $\nabla \cdot \sigma + \rho g = 0....(1)$ $\nabla \cdot v = 0...(2)$

where ρ is density, g is gravitational acceleration, v is the velocity vector with components v and u in the x and z directions, and σ is the stress tensor with components ($\sigma_{xx}, \sigma_{xz}, \sigma_{zz}$). The pressure is given by, $p = -(\sigma_{xx} + \sigma_{zz})/2$, and the deviatoric part of the stress tensor, σ' , relates stress to deformation by way of the rheological properties of the material.

We use linear viscosity to represent viscous material (*i.e.*, salt)

$$(\sigma'_{xx},\sigma'_{xz},\sigma'_{zz}) = 2\eta(\dot{\varepsilon}_{xx},\dot{\varepsilon}_{zz},\dot{\varepsilon}_{zz}) \tag{3}$$

where η is viscosity and the components of strain rate are

$$\dot{\varepsilon}_{xx} = \frac{\partial v}{\partial x} \quad ; \quad \dot{\varepsilon}_{xz} = \frac{1}{2} \left(\frac{\partial v}{\partial z} + \frac{\partial u}{\partial x} \right) \quad ; \quad \dot{\varepsilon}_{zz} = \frac{\partial u}{\partial z} \qquad (4)$$

Non-cohesive frictional-plasticity is modeled by the Drucker-Prager yield criterion, which is equivalent to the Mohr-Coulomb criterion for incompressible, plane-strain deformation.

$$(J_2')^{1/2} = p \cdot \sin \phi_{eff}$$
(5)

where J'_2 is the second invariant of the deviatoric stress and ϕ_{eff} is the effective internal angle of friction of the plastic material, defined in the next sub-section. If an effective viscosity, η_{eff} , is defined using second invariants of deviatoric stresses, J'_2 , and strain rates, I'_2 , as $\eta_{eff} = (J'_2)^{\frac{1}{2}}/2(I'_2)^{\frac{1}{2}}$, then the incompressible plastic flow becomes equivalent to a viscous flow. The numerical solution is determined iteratively using $\eta = \eta_{eff}$ (Willett, 1999).

The model uses an Arbitrary Lagrangian-Eulerian (ALE) method (Fullsack, 1995) in which computations are made on a Eulerian grid that adapts to the evolving model domain. The material properties are tracked and updated using a set of Lagrangian nodal points. An initial Lagrangian mesh is defined, and a large number of injected meshless particles are also added to improve the accuracy of the tracking. This approach allows calculations to be made for large strains without the associated, and often numerically unacceptable, distortion of the finite element mesh. Moreover, it allows for material flux across the model surface, whereby sedimentation and erosion may be included. The model result figures and animations show the Lagrangian mesh. In many places, however, it is too distorted to be useful and has been deleted. In is useful to note here that, although we use the term 'faulting' throughout, the continuum finite element model produces finite width shear zones, not true faults.

Model Geometry and Properties

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The model setup is illustrated in Figure 2. A 1000 m thick viscous layer is overlain by a prograding frictional-plastic overburden wedge of laterally varying thickness. The total model thickness, h, varies from 5500 m to 1010 m following the half-Gaussian sediment profile function

where h_c (1000 m) is the thickness of the viscous layer and h_1 (4500 m) and h_2 (10 m) are the limiting thicknesses of the overburden. The width of the transition region (Fig. 2), l, is determined by the value of d = 200 km (l approximately equals 2d) and the transition starts at $x = x_0$. The total model length is 500 km. The viscous layer is 300 km long with 100 km wide buffer zones of overburden material at each end to allow for deformation beyond the limits of the salt.

The Eulerian and Lagrangian grids consist of 88 vertical by 800 horizontal elements. The grids are adjusted vertically to form a half-Gaussian-shaped topography (Eq. 6). The solution space has a stress-free top surface. No horizontal or vertical velocities are permitted at the vertical model boundaries or the base of the model. Viscous flow is therefore driven solely by the pressure gradient set up by the prograding differential load of the prograding sediments on the viscous substrate.

Sediment progradation is modeled by translating the sediment profile function (Eq. 6) from landward to seaward across the model at the progradation velocity V_{sp} . That is, $x_0(t) = x_0(0) + V_{sp}$. t, where t is time. As the prograding surface is translated across the model, new overburden material is added to fill the region between the current model surface and the prograding surface. This new material is then included in the model calculations. If, during this process, the current sediment or salt surface is above the kinematically defined prograding surface, material is neither added nor removed. The sedimentary material added during progradation has the same properties as the pre-existing sediments. Material properties are constant for all models in this paper, with the exception of models used to test the role of density-driven processes (sub-sections 'Model 2: No Density Contrast Between Salt and Overburden' and 'Model 3: Large Density Contrast Between Salt and Overburden'). The cohesionless frictional-plastic sedimentary overburden has an effective angle of friction $\phi_{eff} = 20^{\circ}$. The effective angle of friction is chosen to represent an average sediment strength that takes into account the effects of pore fluid pressure

$$p\sin\phi_{eff} = (p - p_f)\sin\phi \qquad (7)$$

where *p* is the solid pressure, p_f is the pore fluid pressure, ϕ is the internal angle of friction of the dry material, and ϕ_{eff} , is the effective internal angle of friction. A ϕ_{eff} of 20° represents conditions that are close to hydrostatic pore fluid pressures for the parameters used here. Salt is modeled as a linear viscous material with viscosity $\eta = 10^{18}$ Pa.s.

Model Results

In the animations and figures that follow, each 2 Myr of deposition is shown as a single color band. These thin bands are organized into thicker red, orange, yellow, green, and blue packages, each package representing 10 Myr of deposition. The 2 Myr bands are visible as shades within the colored packages. In this manner, the colored packages repeat every 50 Myr, beginning again with the brightest red. Salt is colored magenta. The material properties are the same as those in Gemmer *et al.*, (2004); however, the width of the salt layer and the sediment geometry and progradation rate differ.

Model 1: Reference Model - Effect of Sediment Progradation

Model 1 is designed to investigate the effect of uniform rate, relatively slow, $V_{sp} = 0.5$ cm/yr, progradation of uniform property, $\phi_{eff} = 20^{\circ}$, $\rho = 2300$ kg/m³ sediments over the 300 km wide, 1 km thick salt layer, which has a density of 2200 kg/m³ and a viscosity of 10^{18} Pa·s. The initial configuration has a 10 m thick sedimentary overburden at the distal end of the salt and an 85 m thick overburden at the proximal, or landward, end of the salt.

Salt deformation is initiated as soon as the sedimentary wedge starts to prograde (Fig. 3A) demonstrating that for thin, and therefore weak, overburden only a small pressure gradient is

required to cause seaward salt flow. The overburden is less than 150 m thick at the landward end of the salt when the salt begins to flow, and contractional deformation of the downdip part of the prograding sediments causes mini-basins to form. (Note that our definitions of selected salt-tectonic terms are provided in the appendix). It is clear that the differential loading by the sediment, and not buoyancy forces owing to the density contrast, initiates the instability because an equivalent model with no progradation remains stable.

In Phase 1 of the evolution, the warping of the overburden, the evacuation of the salt from beneath the downwarps, and its flow into the adjacent upwarps results in the formation of mini-basins. The salt evacuation creates accommodation space for the sediments and the system evolves diachronously forming a series of mini-basins (Anim. 1, 14.5 Myr). These local flows occur simultaneously with the regional salt channel flow, which has a Poiseuille character in the proximal salt and a Couette character in more distal regions (Fig. 3A, deformed tracking mesh).

Phase 2 of the evolution involves a combination of several processes. Extension and listric normal faulting are initiated in the overburden above the proximal salt (Anim. 1), as is clearly seen in the yellow and green strata of Figure 3B. This part of the system has evolved during overburden progradation, becoming part of the unstable region updip from the toe of the slope where the mini-basins form.

At more distal locations the combined regional and local flows lead to the development of triangular-shaped diapirs at the flanks of the mini-basins, some of which are fault bounded (Fig. 3B). Some of these mini-basin bounding diapirs deflate (Fig. 3B, example at 138 km, and Anim. 1) and new depocenters form directly above them.

Phase 3 (Anim. 1, 40.5-60.5 Myr) is characterized by the wholesale extension of the system beneath the slope, and the seaward translation of pre-rafts (defined in appendix) of older overburden that become progressively separated by widening, mainly asymmetric, fault-bounded grabens to form rafts (Fig. 3C; defined in appendix). During the formation of the large fault-bounded depocenters, the strata in the mini-basins are deformed to create a turtle structure (Fig. 3B and 3C). This 'inversion' results from a

combination of the rollover style synkinematic sedimentation on the hangingwall of the listric normal growth faults and the regional evacuation of the salt.

Sedimentation and salt withdrawal in Phase 3 leads to deflation of the diapirs that bound the minibasins and salt is expelled to form more diapirs and mini-basins at the base of the prograding slope. In this manner, diapirs, mini-basins, and extensional grabens form diachronously with their age decreasing seaward. The progressive evacuation of salt, from landward to seaward, leaves extensional basins superimposed on the older mini-basins in areas where the majority of the salt has been evacuated (Anim. 1).

This process of diachronous mini-basin formation would continue, but at 40.5 Myr the flowing salt reaches its distal deposition limit, which acts as a buttress to its continued seaward flow and forces thickening of the evacuated salt (Anim. 1). At 60.5 Myr, Phase 4 of the evolution begins when the salt has thickened sufficiently that it overthrusts younger sediments beyond its depositional limit and forms a large salt nappe (Anim. 1 and Fig. 3D; refer to appendix for definition). The mini-basins located adjacent to the depositional limit of the salt and within the salt nappe (Fig. 3D) are highly allochthonous and were formed earlier at the base of the slope and were advected seaward with the salt. No new distal mini-basins form after 30.5 Myr in this model because the specified prograding depositional profile progressively falls below the surface of the main body of evacuated salt as the salt thickens at the base of the slope (Anim. 1, 50.5 Myr). This means that the prograding sediment wedge cannot create new mini-basins by topographically-induced flow. Even the existing distal mini-basins evolve progressively more slowly as they are advected both seaward and upward as the salt surface rises (Anim. 1, 70.5-100.5 Myr and Fig. 3D).

Natural systems may behave in this way, but it is equally likely that sediments will continue to prograde over the distal salt and that new mini-basins will form. The interesting model result is the clear linkage between the development of mini-basins and the topographic forcing by the prograding sediment wedge.

Model 2: No Density Contrast Between Salt and Overburden

The recent paradigm shift in the interpretation of salt tectonics has lead to a strong emphasis on the differential pressure that exists between thick proximal sediments and thin distal sediments overlying the salt as the main driving force (*e.g.*, Jackson and Vendeville, 1994). While this interpretation is correct, the effect of buoyancy forces acting in concert with the pressure driven flows requires further investigation. Two model experiments are used to investigate the effect of buoyancy forces. Models 1, 2, and 3 are the same except that in Model 2 (this section) the salt has the same density as the overburden (2300 kg/m³) and in Model 3 the density contrast is larger than that found in natural systems.

The initial response to sediment loading by the prograding overburden is similar in both Models 1 and 2. The loading drives a topographically-induced flow and a mini-basin is created (Anim. 2, 8 Myr). Subsequently in Model 2, a single wide basin develops (Anim. 2, 20 Myr), in contrast to the set of minibasins separated by salt diapirs in Model 1. There is some tendency for diapirism of the salt as it is evacuated, but the main effect is the development of pillows and rollers (defined in the appendix) as the salt is segmented during Phase 2 of the evolution, when listric normal faulting and extension of the overburden become important. For the parameter values chosen for these models it is clear that even a small density contrast, 100 kg/m³, between the salt and overburden has a significant impact on the development of mini-basins. When the buoyancy forces act together with the channel flow, there is significant diapirism, as seen in Model 1, but large-scale salt evacuation dominates when the salt is not buoyant, as in Model 2. The difference between Models 1 and 2 becomes more subtle later in their evolution. The general distribution of extensional structures, the formation of pre-rafts and rafts and the intervening depocenters are similar, as is the overall rate of salt expulsion (Fig. 4A and 4B; Anim. 1 and Anim. 2). The salt begins to overthrust its depositional limit at about the same time, 60 Myr, and the large overthrust salt nappe develops in a similar manner (Fig. 4A and 4B). Hence, the differential sediment load controls the overall pattern of proximal extension, distal contraction, and seaward salt flow. However, the buoyancy force owing to the density contrast is responsible for the remaining distal mini-basins and salt

diapirs in Model 1. There is also a distinct difference between the final structure of the advected and disaggregated wide, single early mini-basin (Anim. 2, 24 Myr) of Model 2 (Fig. 4A, 280-370 km, orange and yellow) and the equivalent separated set of mini-basins in Model 1 (Fig. 4B, 270-360 km, orange and yellow). The most striking difference between the models is the inclined smooth interface between the salt and overburden in Model 2 (Fig. 4A, 250-450 km).

Model 3: Large Density Contrast Between Salt and Overburden

It is often useful to investigate systems for limiting, and sometimes unnatural, values of the control parameters to determine end-member responses. In Model 3 the salt has a density of 1900 kg/m³, and other parameter values are the same as those of Models 1 and 2. The density contrast of 400 kg/m³ between overburden and salt is 2-4 times larger than typical situations in sedimentary basins. The model results clearly demonstrate the increasing importance of the buoyancy force under these extreme circumstances.

During Phase 1, deep initially triangular shaped, widely spaced mini-basins form rapidly following the progradation of the overburden wedge onto the salt. The Phase 1 finite deformation shown by the deformed tracking mesh (Fig. 5A) demonstrates that the system evolves by a combination of channel flows driven by the differential overburden loading and what we interpret to be local finite Rayleigh-Taylor (R-T) instabilities in which the dense overburden sinks into the viscous salt and the salt rises diapirically. The diapiric Rayleigh-Taylor-like flows are asymmetric because they occur in combination with the seaward directed channel flow, giving a more upward diapiric flow on the downstream side of the mini-basins than on the upstream side. That the local flows are not just vertical inversions also shows that the salt layer extends and contracts (pinches and swells) to accommodate the diapirism as shown by the deformation of the Lagrangian tracking mesh (Fig. 5A). An aspect of this process is the enhanced Poiseuille flows immediately downstream of the mini-basins in comparison with the larger scale Couette character of the channel flow. An in-depth analysis is needed to understand the reason for the apparent orderly seaward decrease in the size and spacing of the mini-basins (Fig. 5A). During Phase 2 (Anim. 3, 20-60 Myr), listric normal growth faulting and shearing against the diapirs dominates the unstable and extending overburden and the salt is evacuated and shortened to form a series of massive salt walls (Fig. 5B). Large-scale symmetric and asymmetric mock-turtle (refer to appendix) anticlinal structures (Vendeville and Jackson, 1992) form in the depocenters between the salt walls. The early synformal mini-basins/grabens become nearly grounded on basement, the flanks subside further by growth faulting against the diapirs, and the mini-basins are inverted to form the cores of the arched mock-turtle anticlines. At this stage (Anim. 3, 40 Myr) growth is confined to the boundaries of the salt walls. However, later (Anim. 3, 80 Myr) the mock turtles are themselves disrupted by normal faults.

The evolved form of Model 3 (Anim. 3, 96 Myr) contrasts with Model 2. In particular, there are diapirs in the section below the slope region and the interface between the evacuated salt and the overburden is subvertical. and not shallowly dipping across a 100 km wide zone as in Model 2. This difference in geometry clearly reflects the role of the vertical buoyancy force which was absent in Model 2.

Model 4: Regional Seaward Tilt of the Model

Some passive continental margins, for example the Angolan Margin (*e.g.*, Duval *et al.*, 1992; Marton *et al.*, 2000), are interpreted to have undergone significant seaward tilt (*i.e.*, seaward end down) subsequent to the time of salt deposition. Thermal contraction and isostatic adjustment of the extended and rifted lithosphere, which would be expected to increase seaward, is one mechanism that would produce both syn-rift and post-rift tilting. The effects of seaward margin tilt are investigated in Model 4 (Fig. 6, Anim. 4) which is the same as Model 1, with the addition of an initial 0.2° seaward tilt applied to the entire model. This tilt equates to 1.75 km of subsidence of the distal end of the model, at *x* = 500 km, with respect to the proximal end. This model, which instantaneously tilts the salt layer, is equivalent to a natural example in which the tilt occurs sufficiently rapidly that it is finished before there is either sediment progradation or the system becomes unstable. More complex examples in which the tilt and salt tectonics evolve simultaneously can also be investigated. Tilting adds a downslope component of gravitational forcing to the model which is superimposed on the instability that develops during sediment progradation. The main question is whether the addition of such a small gravitational component modifies the model evolution by comparison with Model 1. The effect is significant. During Phase 1, the system forms a wide mini-basin, flanked by more distal discrete mini-basins (Fig. 6A, Anim. 4), a response that is intermediate to that of Models 1 and 2. The addition of the downslope gravitational forcing enhances the instability of the system, in comparison with Model 1, and the Couette flow and salt evacuation are therefore more efficient. Consequently, in relative terms, buoyancy is less efficient which leads to the lower level of diapirism, more like Model 2 where the buoyancy force is completely suppressed. The same effect manifests in Phase 2 and the result is less salt trapped beneath the proximal basin than in Model 1 (Fig. 6B and 6C). The overall faster Couette flow in Model 4 can be seen from the overthrusting of the distal depositional limit of the salt, which occurs at 28 Myr by comparison with approximately 60.5 Myr in Model 1 (Anim. 1). This effect can be seen in Figure 6 where panels B and C are shown at 50 Myr for Models 1 and 4, respectively.

Model 5: Local Isostatic Compensation of Sediment Load

Model 4 clearly shows that margin tilt can have a significant effect on salt structure development, even when the downslope component of gravitational forcing is small. In this section, we investigate the effect of local isostatic compensation of the prograding overburden, which introduces a tilt into Model 5 that has the opposite sense to that in Model 4. This model experiment specifically addresses the effects of the thicker overburden, which results from isostatic adjustment, and the landward tilt owing to compensation on the model evolution. Model 5 is the same as Model 1, with the addition of the isostatic compensation. As material is added to the model, each column of Eulerian finite elements is adjusted vertically in every timestep to account for the current total weight of material in the column. The total weight includes the existing material (taking into account the difference in density between salt and overburden), new overburden added this timestep, and the weight of the overlying water column between

the seabed and sea level, assuming the sea level is at a reference height z = 5500 m, which is the maximum elevation of the prograding overburden wedge (Anim. 5 and Fig. 7). Here, the water loading only influences the isostatic balance. Elsewhere, Gemmer *et al.* (in prep.) investigate the effect of water loading and pore fluid pressure on the deformation of the system. The compensating mantle density is 3200 kg/m³. The profile of the prograding overburden remains the same as that in Model 1, therefore, the prograding wedge is much thicker owing to the additional accommodation space created by isostatic compensation (Anim. 5).

The Phase 1 evolution of Model 5 is similar to Model 1, with the early creation of a series of mini-basins (Fig. 7A, 100-140 km) and intervening diapers. In the next phase, listric normal growth faulting starts on the basin boundaries (Fig. 7B). Later in Phase 2 the subsidence caused by the isostatic adjustment reduces the seaward Couette flow by comparison with Model 1 (Anim. 5, 48 Myr). The evacuation of the salt is less efficient than in Model 1 and significant amounts of salt remain trapped beneath the proximal overburden (Fig. 7C and 7D and Anim. 5).

Another aspect of Phase 2 is the style of the listric growth faults. At approximately 40 Myr there is no strongly preferred polarity of the basin-bounding faults (Anim. 5). The faulting has modified the early diapirs into large amplitude rollers but these have a generally symmetrical triangular shape. Between 40 Myr and 60 Myr, the extending system develops several significant counter-regional faults positioned on the proximal flanks of the salt rollers (Anim. 5 and Fig. 7C).

In Phase 3, growth faulting in the proximal basin becomes more symmetrical with graben development augmenting the counter-regional faults. This process evolves diachronously in the seaward direction and appears to be a precursor to stabilization of the overburden. The pattern of counter-regional faulting, symmetric graben formation and eventual stability correlate with decreasing asymmetry and decreasing local taper of the system suggesting that asymmetric faulting is favored when the system has a higher taper. The evolution of the system is accompanied by the progressively seaward migrating diachronous creation of mini-basins and diapirs, the deflation of the diapirs, and formation of counter-

regional faults. These processes are repeated until the salt flow reaches and eventually overthrusts its depositional limit in phase four (Anim. 5, 80 Myr).

Discussion

The numerical models described in the previous section display a wide range of behavior and structures related to salt mobilization and continued deformation; structures that are comparable to those shown in published natural examples and analog models (*e.g.*, Peel *et al.*, 1995; Koyi, 1996; Ge *et al.*, 1997; Tari *et al.*, 2002; Vendeville, sub.). On the basin or regional scale, the dominant process driving deformation is differential loading. However, the role of buoyancy, owing to the density contrast, is evidently important for the development of smaller scale features, such as mini-basins and diapirs.

The development of various structures and deformation patterns seen in the models can be categorized into four phases. In Phase 1, warping of prograding overburden results in the formation of depocenters in the downwarps, and a concentration of salt in the upwarps. As sedimentation continues, salt is progressively evacuated from beneath the depocenters forming mini-basins flanked by diapirs. In Phase 2, the mini-basins begin to be dissected by listric normal faults as the overburden is extended in the region updip from the slope. This process leads into Phase 3 of the evolution, where the overburden experiences increasing amounts of extension and is rafted into blocks separated by grabens. In Phase 4, the final phase, the seaward salt flow reaches the distal buttress that is its depositional limit. At this point, seaward movement of the salt is inhibited and it begins to thicken substantially, eventually overthrusting younger sediments forming a large allochthonous salt nappe. The Phase 1 development of mini-basins and diapirs is concentrated at the toe of the prograding overburden slope, and propagates seaward as the sediment profile progrades. The phase-to-phase evolution of the system therefore occurs diachronously, such that more distal regions of the model may be experiencing Phase 1 deformation, while proximal regions have begun to develop Phase 2 or 3 extensional structures.

Each model progresses through the developmental phases somewhat differently, owing to the differences in model parameters. In some models, one or more phases may be suppressed. The evolution of Model 1 is driven by the prograding differential load and small buoyancy forces resulting from the 100 kg/m³ density contrast between the sediments and salt. In Model 2, there is no density contrast, and deformation is driven solely by differential loading alone. The regional salt flow pattern and rate in Model 2 are comparable to Model 1, which develops Phase 2 and 3 structures similar to those in Model 1 (Fig. 4). However, the early warping in Phase 1 of Model 1 is suppressed in Model 2, and the deformation is dominated by horizontal seaward salt flow, as opposed to the local vertical flows, and the Model 2 system evolves a smooth salt-sediment interface that contains only small perturbations (Fig. 4A). This is notably different from Model 1, which develops mini-basins and diapirs early on (Fig. 3A), and eventually large salt walls as Phase 4 begins (Fig. 3D). Based on this finding, we infer that the development of diapirs and mini-basins in Model 1 is due in part to buoyancy forces.

The nature of the instability that creates the mini-basins needs to be investigated more carefully. One end-member process is that described by Fletcher *et al.* (1995) in which a mini-basin forms by the topographically-induced flow induced by an overlying sediment wedge that has its surface slope maintained by deposition and erosion during deformation. The mini-basins in Model 1 have similar geometries to the Fletcher *et al.* (1995, Fig. 24 and 25) results. However, the development of these features may be more complicated because the current models have a frictional-plastic overburden (rather than viscous). As a result, there is a finite density contrast between the salt and overburden, and the overburden progrades during development of the mini-basins. Under these circumstances, the type of instabilities in a frictional material overlying a viscous half-space (Triantafyllidis and Leroy, 1994) may also be important. The length scales of the basins, 5-10 km, are too large for folding instabilities of the thin overburden and the process may best be described conceptually as warping of the overburden. The competition between seaward and upward salt flow is also important, and in cases where the upward component is enhanced, diapirs and mini-basins are favored. In Model 2, the upward component is completely suppressed, and therefore it does not produce diapirs or mini-basins.

Model 3 displays the opposite extreme. In this model, the density contrast is larger (400 kg/m³) than in Model 1, and the upward component of salt flow, driven by the larger buoyancy effect, dominates the system. Some aspects of the effect of buoyancy forces on lubrication squeeze flows have been addressed by Lehner (2000) who emphasizes that the early structural evolution of the top salt is sensitive to the combined effects of differential loading and buoyancy, as well as the geometry of the salt bottom and, "...consequently must have a decisive influence on the entire salt tectonic history of a basin". Buoyancy adds nonlinearity to the Reynold's equation that describes simplified lubrication flows (Lehner 2000, equation 11), and this nonlinearity may drive the instability that creates perturbation of the top salt that subsequently grows as a finite Rayleigh-Taylor instability.

The buoyancy force also enhances the Phase 3 development in Model 3 and large mini-basins and salt walls form (Fig. 5). Although the density contrast is greater than that in nature, the structures in Model 3 look similar to those seen in many natural examples. Equivalent models with an even larger density contrast develop a complete range of salt structures, including coalesced canopies. That the diapirism appears more realistic for density contrasts that are larger than those that occur naturally suggests that other model parameters need adjusting. The efficiency of buoyancy-driven diapirism depends on the relative rates of vertical and horizontal flow. Consequently, a model system that is less unstable horizontally (*i.e.*, has a lower Couette velocity) will evolve more slowly and therefore has more time for vertical diapirism driven by lower, more natural, density contrasts, to develop. Gemmer et al. (2004) show that the Couette velocity depends on the geometry and effective internal angle of friction of the overburden, the thickness and viscosity of the salt layer and the transition width of the sediment progradational profile. Any combination of, a reduction in the differential thickness, h_1 - h_2 , of the overburden; an increase in ϕ_{eff} ; or an increase in l or the salt viscosity, will reduce the Couette velocity and may correspondingly increase the relative efficiency of diapirism. However, the buoyancy-driven diapirism also depends on the strength of the overburden which is a function of ϕ_{eff} and it is likely that an increase in ϕ_{eff} alone will further suppress diapirism. Clearly, it is necessary to explore the model

behaviors in relation to variations in model parameters. This can be done most effectively if the search is done for variations in the values of the parameter groups that control a non-dimensional version of the model.

Model 4 is tilted seaward by 0.2° , which enhances seaward salt flow owing to the additional downhill gravitational component. In this model, which has the same density contrast as Model 1, diapirs and mini-basins comparable to those in Model 1 do not form; instead the system is dominated by seaward salt flow. This flow is also faster and overthrusting occurs ~30 Myr earlier in Model 4 than in Model 1 (Fig. 6). This has potentially important implications for regions that are thought to have experienced early thermal related subsidence, *e.g.*, the Angolan margin, and it may be an important mechanism that initiates early salt movement, even in the absence of differential loading.

Model 5 includes local isostatic adjustment. As sediments prograde, the model subsides creating more accommodation space and a thicker sediment wedge than in the other models. The subsidence also causes the salt layer to become tilted landward. It appears that this landward tilt of the salt layer has a larger effect on the flow regime than the increased differential pressure in the salt caused by the thicker overburden wedge resulting in a slower seaward Couette flow than in Model 1. However, there are complicating factors that need to be analyzed, including the effect of the thicker overburden at the proximal limit of the salt on the stability of the system and the velocity of the Couette flow when the system is unstable (*cf.*, Gemmer *et al.*, 2004). The slower seaward flow allows the buoyancy to be more effective and therefore the Phase 1 evolution of diapirs and mini-basins are more pronounced in Model 5 than in Model 1 (Fig. 7, part 1).

Conclusions

We have presented preliminary 2D finite element numerical models of sediment progradation above a linear viscous salt layer. The models were used to investigate how sediment progradation, buoyancy, margin tilt, and isostatic subsidence affect salt tectonic deformation in passive continental margin settings.

Sediment progradation causes a differential load on the salt and can induce an unstable seaward flow resulting in a deformational evolution that may be categorized into four main phases: 1) formation of mini-basins and associated flanking diapirs; 2) onset of listric normal growth faulting and extension of the overburden; 3) large scale salt evacuation, formation of pre-rafts and rafts, and inversion of the minibasins, and; 4) formation of a contractional allochthonous salt nappe that overthrusts the depositional limit of the salt.

Decreased salt density increases vertical buoyancy forces, thereby enhancing local upward salt flow and diapirism. In contrast, models with no density contrast show a near complete suppression of diapirism and localized mini-basin formation, indicating that these structures are sensitive to buoyancy effects. Seaward tilt of the model has a similar effect to a reduction of the density contrast, because, when tilted, seaward salt flow is enhanced by the gravitational effect of flowing downhill, thereby reducing the relative importance of the upward buoyancy-related flow component. Local isostatic adjustment causes subsidence and increased accommodation and sedimentation. This subsidence causes the salt layer to be tilted landward, and therefore salt is forced to flow upward. This decreases the seaward component of the flow, and in contrast to seaward tilt, causes the relative importance of the upward buoyancy-related flow component to be enhanced, leading to the formation of more pronounced diapirs and mini-basins.

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Appendix

Definitions of various salt tectonics terminology, as used in this paper:

<u>Mini-basin</u>: A basin in which most of the accommodation space is created by salt withdrawal from beneath the basin, and not by extensional graben formation. Sedimentation in these basins is contemporaneous with the salt withdrawal.

<u>Mock-turtle structure</u>: An anticlinal structure resulting from the inversion of stratigraphy in a mini-basin formed directly overlying a deflating diapir (modified after Vendeville and Jackson, 1992).

<u>Nappe</u>: A large-scale salt sheet formed when salt overthrusts its distal depositional limit. Salt nappes may carry overlying sediments and advected mini-basins formed further landward.

<u>Pillow</u>: A salt structure in conformable contact with overlying sediments in which the vertical dimension is much smaller than the horizontal dimension, resulting in a broad feature with shallowly sloping sides (modified after Jackson and Talbot, 1994).

<u>Pre-raft</u>: A fault block of laterally translating overburden that is bounded by listric normal faults. These blocks have not yet separated and still overlie their original footwall (modified after Jackson and Talbot, 1994).

<u>Raft:</u> A fault block of laterally translating overburden that has separated from adjacent fault blocks to the degree that it no longer rests in its original footwall. These packages form during extension of pre-existing mini-basins and are bounded by grabens (modified after Jackson and Talbot, 1994).

<u>Roller</u>: Triangular salt structures in which one flank is gently dipping and in conformable contact with the overburden, and a more steeply dipping flank whose overburden contact forms a normal fault (modified after Jackson and Talbot, 1994).

<u>Turtle structure</u>: An anticlinal structure resulting from the inversion of stratigraphy in a mini-basin as the flanking diapirs deflate (modified after Jackson and Talbot, 1994).

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Figure 1: Illustration of flow regimes associated with differential loading of frictional sediment overburden (gold) acting on viscous salt (magenta). A) Poiseuille channel flow of salt, stable sediment overburden; B) marginally unstable overburden giving combination of Poiseuille and Couette salt channel flows; C) very unstable overburden such that the salt channel is dominated by Couette flow. Panels B and C show regions of proximal extensional failure and distal contractional failure of the overburden.

Figure 2: Basic model geometry and properties. Salt layer (magenta), pre-existing sediment overburden (gold), prograding sediments (10 Myr depositional intervals, orange through red cycles, each comprising five variably shaded bands of 2 Myr duration). All overburden sediments have the same frictional-plastic properties. Top surface is the prograding sediment profile function (Eq. 6) which advances seaward (to the right) at velocity V_{sp} . This kinematic sedimentation model produces an aggradational stratigraphy during progradation over the salt. See text for description of other parameters.

Figure 3: Model 1 results. A) Phase 1 formation of mini-basins (100-160 km) during early sediment progradation. Note Poiseuille and Couette styles of flow in salt shown by deformed Lagrangian tracking grid. B) Phase 2 extension and normal growth faulting, deflation of proximal diapirs (138 km, see also panel C) and development of distal mini-basins and triangular shaped diapirs. C) Phase 3 continued extension of slope region, formation of pre-rafts and rafts of older sediments, continued 'inversion' of mini-basins, and evacuation of salt. D) Phase 4 formation of large overthrust salt tongue carrying existing mini-basins. The phases evolve diachronously from proximal to distal regions.

Figure 4: Comparison of Model 2 (no density contrast between salt and overburden) and Model 1 (100 kg/m³ density contrast) results during Phase 4 showing the influence of the lower salt density on diapirism and salt evacuation in Model 1.

Figure 5: Model 3 results showing effect of large density contrast (400 kg/m³) between overburden and salt. A) Enhanced development of mini-basins and diapirs during Phase 1. B) Development of mock-turtle structures and massive salt walls during Phase 2.

Figure 6: Model 4 results showing effect of initial basinward tilt of model. A) Phase 1 development of wide

mini-basin by comparison with discrete local mini-basins in Model 1 (Fig. 3A). B) and C) comparison of Model 1 and Model 4 at 50 Myr showing enhanced seaward flow caused by seaward tilt in Model 4.

Figure 7: Model 5 results showing effect of isostatic compensation. A) Phase 1 development of mini-basins and associated diapirs. B) Phase 2 onset of extensional normal faulting during sediment progradation. C) Phase 2 landward tilting of salt and development of counter-regional normal growth faults. D) Phase 3 diachronous evolution to graben faulting augmenting counter-regional faults. E) Phase 4 formation of salt tongue that progressively climbs above the isostatically adjusting sediments during overthrusting.

Animation 1: Model 1 results showing effect of progradation at $V_{sp} = 0.5$ cm/yr of frictional-plastic sediments over salt. Color coding for all animations: salt (magenta), pre-existing sediments (gold), prograding sediments (10 Myr depositional intervals, orange through red cycles, each comprising five variably shaded bands of 2 Myr duration). All overburden sediments have the same effective internal angle of friction, 20°, and density 2300 kg/m³. Salt is linearly viscous with viscosity 10¹⁸ Pa.s, density 2200 kg/m³, and is initially 1 km thick. 0-28.5 Myr: Phase 1 formation of mini-basins (100-180 km) during early sediment progradation. Note Poiseuille and Couette styles of flow in salt shown by deformed Lagrangian tracking grid. 28.5-40.5 Myr: Phase 2 extension and normal growth faulting, deflation of proximal diapirs (138 km), and development of distal mini-basins and triangular shaped diapirs. 40.5-60.5 Myr: Phase 3 continued extension of slope region, formation of pre-rafts and rafts of older sediments, continued 'inversion' of mini-basins, and evacuation of salt. 60.5-100.5 Myr: Phase 4 formation of large overthrust salt tongue carrying existing mini-basins. The phases evolve diachronously from proximal to distal regions.

Animation 2: Model 2 has the same properties as Model 1 except that the salt has the same density as the sediments, 2300 kg/m³. 0-28 Myr: Phase 1 evolution of wide continuous mini-basin and Couette flow of salt shown by deformed Lagrangian marker grid. 28-40 Myr: Phase 2 extension and normal growth faulting, local 'inversion' within the wide mini-basin. 40-60 Myr: Phase 3 wholesale extension of the overburden and formation of rafts, efficient evacuation of salt to form dipping interface between salt and overburden and no significant diapirs. 60-100 Myr: Phase 4 formation of allochthonous overthrust salt tongue.

Animation 3: Model 3 has the same properties as Model 1 except that the salt is much less dense than the sediments, 1900 kg/m³. 0-28 Myr: Phase 1 enhanced development of mini-basins and diapirs and onset of normal growth faulting. 28-40 Myr: Phase 2 development of mock-turtle structures and massive salt walls during extension of the overburden. 40-64 Myr: Phase 3 continued extension of slope region and formation of salt diapirs. 64-100 Myr: Phase 4 formation of allochthonous overthrust salt tongue, counter-regional faults and proximal graben structures.

Animation 4: Model 4 is the same as Model 1 except that there is an initial basinward tilt of 0.2° of the entire model. 0-28 Myr: Phase 1 development of wide mini-basin and adjacent discrete mini-basins, onset of extensional growth faulting and local inversions in the interior of the single wide mini-basin. 28-40 Myr: Phases 2 and 3 extension and normal growth faulting, formation of distal mini-basins, and flow of salt beyond distal depositional limit. 40-82 Myr: Phase 4 enhanced evacuation and flow of the salt forming allochthonous overthrust tongue, seaward advection of mini-basins that were formed at the base of the slope in Phase 3, continued wholesale extension of the slope region with widespread rafting of older sedimentary packages.

Animation 5: Model 5 is the same as Model 1 except the system is locally isostatically compensated for the weight of the salt and sediments and assumed water loading above the sediments to the level of the prograding shelf (5500 m). The mantle density used in the isostatic calculation is 3200 kg/m³. 0-20 Myr: Phase 1 development of mini-basins and associated diapirs. 20-32 Myr: Phase 2 onset of extensional normal fauting during sediment progradation. 32-60 Myr: Phase 2 landward tilting of salt and development of counter-regional normal growth faults. 60-80 Myr: Phase 3 diachronous evolution to graben, faulting augmenting counter-regional faults. 80-108 Myr: Phase 4 evacuation and thickening of distal salt and formation of salt tongue that progressively climbs above the isostatically adjusting sediments during overthrusting.















