Halokinetic modulation of sedimentary thickness and architecture: a numerical modelling approach

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ABSTRACT

Subsurface salt flow can deform overlying strata and influence contemporaneous sedimentary processes. Studying salt-sediment interactions is challenging in the subsurface due to poor imaging adjacent to salt, and in the field due to the dissolution of halite. Discrete Element Modelling provides an efficient and inexpensive tool to model stratigraphy and deformation around salt structures, which is advantageous over other modelling techniques as it realistically recreates brittle processes such as faulting. Six 2D experiments were run representing 4.6 Myr (46,000 timesteps) to determine the effect of salt growth on syn-kinematic stratigraphy. Halokinetic deformation of stratigraphic architecture was assessed by varying sediment input rates (slow, intermediate, and fast), and increasing and decreasing them through time. Results show the realistic formation and evolution of salt-related faults which define a zone of halokinetic influence, ~3 times the width of the initial diapir. Outside of this, pre- and syn-kinematic stratigraphy are undeformed. Within this zone, syn-kinematic strata are initially isolated into primary salt withdrawal basins, onlapping and thinning towards the salt-cored high. In most models, syn-kinematic strata eventually thin across and cover the diapir roof. Thinning rates are up to six times greater within 350 metres of the diapir, compared to further afield, and typically decrease upwards (with time) and laterally (with distance) from the diapir. Outputs are compared to a subsurface example from the Pierce field, UK North Sea, which highlights the importance of considering local fluctuations in diapir rise rate. These can create stratigraphic architectures that may erroneously be interpreted to represent 'apparent' increases/decreases in sedimentation rate. Exposed examples, such as the Bakio diapir, northern Spain, can be used to make inferences of the expected depositional facies, below model resolution. Our models aid prediction of sedimentary unit thickness and thinning rates, and can be used to test interpretations arising from incomplete or low-resolution subsurface and outcrop data when building geological models for subsurface energy.

INTRODUCTION

Deformation by salt-tectonics influences over 120 sedimentary basins globally (e.g. Hudec and Jackson 2007; Figure 1). These basins include some of the world's largest subsurface energy-producing provinces, such as the Gulf of Mexico (e.g. Booth et al. 2003; Hudec et al. 2013), the North Sea (e.g., Mannie et al. 2014; 2016; Charles and Ryzhikov 2015; Stricker et al. 2018; Figure 1A), offshore Angola (e.g. Oluboyo et al. 2014; Doughty-Jones et al. 2017; Howlett et al. 2020), offshore Brazil (e.g. Rodriguez et al. 2018; 2020; Pichel et al. 2019), and the Precaspian Basin (e.g. Duffy et al. 2017; Pichel and Jackson 2020). There is therefore a need to better understand the interactions of sedimentary processes and routing with topographic relief associated with salt tectonics to provide insight into sediment routing patterns around topography (e.g. Ribes et al. 2015; Cumberpatch et al. 2021a; Giles and Lawton 2002; Rowan et al. 2003; Pichel and Jackson, 2020). As a consequence, this can improve predictions of reservoir distribution and trap geometry and style for carbon storage (e.g. Maia da Costa et al. 2018; Roelofse et al. 2019), geothermal energy (e.g. Harms 2015; Daniilidis and Herber 2017; Andrews et al. 2020), and hydrocarbons (Figure 1; e.g. Hodgson et al. 1992; Jackson and Hudec 2017; Pichel and Jackson, 2020).

Constraining the dynamic evolution of the sediment-salt interface through time and space remains challenging. Despite advances in the quantity and quality of 3D seismic reflection data, the salt-sediment interface remains difficult to image due to poor velocity control, steep to overturned bedding and near-diapir deformation (Figure 1; Jones and Davison 2014). Resolution issues caused by variable lithological distributions, both within the salt and its overburden, further complicate seismic reflection-based, subsurface analysis (Davison et al. 2000; Jones and Davison 2014). This leads to uncertainty in prediction of facies and thicknesses via seismic methods (Figure 1; e.g. Berton and Vesely 2016; Hossain 2019). Therefore, halokinetically influenced depositional systems (or portions of such systems) benefit from calibration with outcrop analogues (e.g. Lerche and Petersen 1995; Jackson and Hudec 2017). Exposed examples are often limited in the rock record, largely due to dissolution of halite forming the core of the salt bodies (Jackson and Hudec 2017). Rare exposures provide sub-seismic scale facies information for shallow- (e.g. Giles and Lawton, 2002; Giles and Rowan 2012), deep- (e.g. Poprawski et al. 2014; 2016; Cumberpatch et al. 2021a) and non-marine stratigraphy (e.g. Banham and Mountney 2013a; b;14; Ribes et al. 2015; 2017). Detailed depositional facies models can be developed from subsurface (e.g. Madof et al. 2009; Rodriguez et al. 2020) and outcrop datasets (e.g. Banham and Mountney 2013a; b; 2014; Cumberpatch et al. 2021a) providing a useful framework that can be applied to salt-influenced basins globally. Many field examples are small in size compared to subsurface basins and therefore provide only small-scale details of, for example, sedimentary structures and stratal stacking patterns, rather than the larger, basin-scale tectonostratigraphic context of salt-sediment interactions provided by integrated subsurface datasets. While useful, each subsurface or field example represents a unique record of the ratio of salt rise and sedimentation rate. Also, subsurface and field examples provide only one snapshot in time (Figure 1). Physical models have an advantage in recreating the evolution of specific subsurface analogues (Dooley et al. 2013; 2015; 2020; Dooley and Hudec, 2017; Ferrer et al. 2017; Roma et al. 2018) and studying sedimentary gravity-currents flow distribution and evolution adjacent to salt topography (Gaullier and Vendeville, 2005; Sellier and Vendeville, 2010; Soutter et al. 2021).

A number of remaining questions can be addressed by taking a numerical modelling approach, that allows us to modify and isolate the key controls on salt-sediment interactions. These include: 1) How does salt topography influence depositional systems, and thus depositional facies, and how does this vary laterally and temporally? 2) How do unconformities, onlap contacts, and faults and fractures vary in salt-influenced settings? 3) How does sedimentation rate influence the width of the roof and basin salt-related deformation zones? and 4) How do stratigraphic thinning rates associated with salt growth vary with sedimentation rate and distance from the salt structure?

Most recent advances in numerical models use finite element models (FEM), which are based on continuum methods. Such studies have focused on the physical conditions required for the initiation and development of diapirism (Poliakov et al. 1993; Gemmer et al., 2004; 2005; Chemia et al. 2008; Fuchs et al. 2011; Fernandez and Kaus 2015; Nikolinakou et al. 2017; Hamilton-Wright et al. 2019; Peel et al. 2020), the stratigraphic architecture of subsiding minibasins (Sylvester et al. 2015; Wang et al. 2017; Fernandez et al. 2020), reconstructing evolutionary history of salt-affected stratigraphy (Ismail-Zadeh et al. 2001; 2004; Pichel et al. 2017; 2019), and salt-related stress (and strain) analysis (Luo et al. 2012; 2017; Nikolinakou et al. 2012; 2014a; b; 2018; Heidari et al. 2017). FEM often treats the overburden as a continuous frictional-plastic or viscous-plastic material, which prevents the development of realistic brittle deformation (e.g. fracturing and faulting) in overburden stratigraphy (Figure 1). Most FEM packages currently have limited capacity to generate faults during simulations, and therefore either have no faults or faults that are pre-defined at the start of the model simulation (e.g. Heirdari et al. 2016; Nikolinakou et al. 2014; 2018). FEM is advantageous for studying ductile deformation and salt flow dynamics (Albertz and Ings 2012).

Discrete Element Models (DEM), which essentially treat the contact between each element as a slip surface, are able to replicate spontaneous, realistic, localised fault nucleation and growth (Pichel et al. 2017; 2019; Cumberpatch et al. 2021b) and are therefore appropriate for studying the interactions between salt-related topography, sedimentation, and stratigraphic evolution. DEM provides a quick, efficient and inexpensive method to investigate system evolution through time (Allen and Tildesley 1987; Donzé et al. 1994; Finch et al. 2003; 2004). It is therefore possible to test a number of scenarios and collect structural growth and syn-kinematic sediment thinning rate data during deformation. This helps to improve recognition of the

109 processes involved with salt-sediment interactions. DEM studies have recently been adapted to salt
110 tectonics so that elements representing salt behave as viscous-plastic materials (Pichel et al. 2017; 2019).
111
112 Here, we use a DEM approach to understand how sedimentation rate effects stratal geometries in salt
113 basins experiencing diapirism (Figure 1). First, we generate a baseline model with no sediment fill, to
114 determine the effect of salt growth on pre-kinematic, diapir-capping stratigraphy. We define 'pre-kinematic
115 layers' as layers deposited prior to our simulation, they are discordant to the diapir and thus represent the
116 early diapiric syn-kinematic strata related to the phase of diapirism that is assumed to have emplaced the
117 diapir into our model (Jackson and Hudec 2017). We then vary sedimentation rate and patterns to study
118 how these control the stratigraphic record of halokinesis. The aims of this study are to: 1) investigate
119 variable syn-kinematic sedimentation rates adjacent to a dynamic salt diapir using a DEM; 2) quantify near-
120 diapir thinning rates and how this is controlled by varying sedimentation rates and patterns; and 3) compare
121 the results to subsurface and field analogues to test the validity of our approach and our model predictions.

DATA AND METHODS

123
124 The Discrete Element Model (DEM) applied here is a discontinuous numerical method, which derives
125 from the Particle Dynamics Method and Lattice Solid Model (Mora and Place 1993; 1994). DEMs have
126 been successfully used in physics and chemistry to study liquid and gas behaviours (Allen and Tildesley
127 1987; Hardy and Finch 2006). In geoscience, DEMs have been applied in two- and three-dimensions
128 (Longshaw et al. 2009) to forecast geological hazards, often associated with mining (Cil and Alshibli, 2012;
129 Thoeni et al. 2014; Lu et al. 2014; Zhao et al. 2016; Benseghier et al. 2020), to investigate deformation
130 caused by faulting and folding (Donzé et al. 1994; Finch 1998; Finch et al. 2003; 2004; Imber et al. 2004;
131 Hardy and Finch 2005; 2006; Schöpfer et al. 2006; Bellheine et al. 2009; Longshaw et al. 2012; Abe and
132 Urai 2012; Katz et al. 2014), and to study regional-scale salt tectonics (Pichel et al. 2017; 2019).

133
134 DEMs offer advantages over other numerical methods in that scale is not a restriction, complex re-meshing
135 is not required, and results are easily reproducible. DEMs are fundamentally discontinuous, and therefore
136 each element simulates the specific physical properties of a given rock. These conditions make it a fit-for-
137 purpose method to quantitatively study syn-kinematic deformation. DEMs do have limitations, including
138 the need for careful calibration of element parameters (Abe et al. 2011; Botter et al. 2014), and the limited
139 number of elements and the duration of simulations (Zhu et al. 2008).

140
141 DEM treats objects as assemblages of circular elements, connected by breakable elastic bonds through a
142 'repulsive-attractive' force. Elements remain bonded until their breaking separation (defined as the relative
143 strength of the assemblage) is exceeded (Donzé et al. 1994; Finch et al. 2004). Once these bonds break,
144 previously connected elements experience no further 'attractive' force, if these elements return to contact
145 with each other a 'repulsive' force acts between them, preventing the healing of bonds (Finch et al. 2003;
146 2004; Hardy and Finch 2006). Motion of elements is frictionless and cohesionless with elasto-plastic
147 behaviour (Finch et al. 2003; Hardy and Finch 2007). Forces are resolved in the x and y directions and
148 elements are subject to gravity (F_g) (Finch et al. 2003). The equations that define the inter-relationship of
149 all forces acting on the DEM are:

150
151 $F_x = F_{i,n} - V\dot{X}$
152 $F_y = F_{i,n} - V\dot{Y} + F_g$

153
154 Where $F_{i,n}$ corresponds to the total elastic force acting on an element, V represents the viscosity and $\dot{X}$ and
155 $\dot{Y}$ correspond to the velocity of the element in the x- and y- directions. A viscous term is added to counteract
156 the elastic behaviour within a closed system, making it ideal for studying quasi-steady state tectonic
157 processes (Finch et al. 2004; Pichel et al. 2017; 2019). For a comprehensive description of the equations
158 governing DEM, see Mora and Place (1994), Finch et al. (2003; 2004), and Hardy and Finch (2005, 2006).

159
160 Pichel et al. (2017, 2019) recently used a DEM to model salt-tectonics for the first time, studying regional-
161 scale compressional salt tectonics (Pichel et al. 2017), and the effect of base salt relief on salt flow and
162 overburden deformation styles (Pichel et al, 2019). Cumberpatch et al. (2021b) adapted these models to

focus on the modulation of stratigraphy by salt diapir growth. In these models, the elements representing salt were adjusted so they behave as a viscous-plastic material in order to represent rock-salt (Pichel et al. 2017; 2019). This requires inter-element interactions to be adjusted so they behave macroscopically as viscous-plastic materials and deform microscopically by dislocation creep, which is expected for dry rock salt (Spiers et al. 1990; Pichel et al. 2017; 2019). This does not completely reproduce salt, which typically deforms on a spectrum of mechanisms including diffusion and dislocation creep (Spiers et al. 1990; Jackson and Hudec 2017; Pichel et al. 2017), and usually contains traces of brines (Warren 1999; 2006), but is considered a satisfactory assumption for studying salt tectonics. Pichel et al. (2017) tested breaking separations (BS) using biaxial compression tests. Values representing tenths of a model unit (e.g. 0.05) develop defined fault segments, and produced responses typical of brittle materials; these values are therefore used to represent overburden sediment in this study. Values representing hundredths of a unit (e.g. 0.001), however, show a minor elastic component (e.g. $F_i,\alpha \approx 0$), representing ductile viscous-plastic materials that accumulate strain without significant stress variations. Consequently, a breaking separation of 0.001 for salt elements is used in this study (see also Pichel et al., 2017), and other physical (Spiers et al. 1990) and numerical (Li and Urai 2016) experiments of salt deformation. By using this value we ensure salt element motion is entirely controlled by the viscosity and gravity of the system (viscous-plastic behaviour). The scaled viscosity of the salt is 1.1×10^9 Pa s, which is lower than its real-world viscosity (10^{17}-10^{18} Pa s); Hudec and Jackson 2007; Jackson and Hudec 2017), but works as a reasonable approximation when compared with physical models (e.g. Vendeville et al. 1995; Dooley et al. 2009; 2012).

The modelled media in this study consists of a simulated 4.5 km by 1.5 km box, to ensure that the outer boundaries do not influence the structural evolution of the model centre. The media consist of an undeformable base and ~44,500 elements with varying radii (0.175-0.35 units, representing 5.25-9.75 metres); these are randomly distributed to reduce failure in preferential orientations within the matrix. A 150 m-thick salt layer is overlain by nine coloured 150 m thick pre-kinematic overburden layers. A salt density of 2.2 g cm^{-3} is used, to mimic a slightly impure halite composition, comparable to many global salt basins (e.g., Warren 1996; 2006; Hudec and Jackson 2007; Jackson and Hudec 2017; Grant et al. 2019) and previous models (Pichel et al. 2017; 2019; Cumberpatch et al. 2021b).

We do not investigate the initial stages of diapir evolution, which have been well-studied elsewhere (Trusheim 1960; Vendeville and Jackson 1992; Costa and Vendeville 2002; Hudec and Jackson 2007), and instead focus on late stage diapir growth. Therefore, we simplify a complicated three-dimensional process into a two-dimensional model, where we assume a linear salt wall or radially symmetric diapir was emplaced by an earlier phase of diapirism; such an assumption is used in other numerical (Pichel et al. 2017; 2019) and physical models (Davison et al. 1993; Dooley et al. 2009; 2012). This allows us to focus on the coupled deformation-sedimentation characterising the late stage of diapir growth, when a tall pre-existing diapir may be rejuvenated by compression (not modelled here) or rise due to buoyancy (active diapirism). During the experiment the diapir is assumed to grow by halokinetic active rise, driven by the pressure of the overburden on the salt source layer and to a lesser extent by the density difference between the salt and the overlying stratigraphy (Jackson and Hudec 2017). Such growth can happen in the absence of regional tectonics, although mild far-field compression or extension, which can enhance deformation rates, are likely in most natural settings (Jackson and Hudec 2017). Active rise is retarded by roof thickness and strength, and salt viscosity. Therefore, diapir height must be >66-75% of the surrounding overburden thickness for substantial halokinesis to occur and the roof thickness must be <750 m (Schultz-Ela et al. 1993). In adhering to this rule, we invoke an individual sinusoid, 750 m (base width) wide, 1050 m (70% of the 1500 m overburden) tall diapir and thus a 450 m roof (Figure 2). The diapir geometries used in our models are comparable to those observed in natural examples (Davison et al. 2000; Jackson and Hudec 2017). Overburden breaking separation (relative strength; BS) increases with depth linearly from 0.023 – 0.027, in agreement with increasing rock strength with depth. An overburden density of 2.4-2.6 g cm^{-3} is used in agreement with natural and modelled examples, and increases with depth (e.g., Dooley et al. 2009; 2012; Fuchs et al. 2011).

In nature, diapir growth periods are hugely variable in duration, ranging from 100,000s of years to 100s of millions of years (Jackson and Hudec, 2017). Our focus on late-stage of diapir growth and using seismic stratigraphic observations from natural examples (Oluboyo et 2014; Grimstad 2016), and run times of previous numerical models (Pichel et al. 2017; 2019), supports our experimental run times of 46,000

219 timesteps with a timestep equivalent to 100 years (4.6 Myr in total). We impose an upwards motion of 0.023
220 mm/year, based on North Sea diapirs (Davison et al. 2000) to all elements representative of salt to mimic
221 diapir growth rate (Figure 2); this aims to replicate the volumetric salt supply rate (Q) described by Peel et
222 al. (2020). The diapir grows for 2.2 Myr (22000 timesteps) to allow the model to equilibrate creating seabed
223 or surface topography, prior to the addition of sediment (Figure 3).

225 Sediment is added from 2.2 Myr with a constant density of 2.3 g cm $^{-3}$, in agreement with natural examples
226 of near seabed sediment (Tenzer and Gladkikh 2014; Rider and Kennedy 2018) and a breaking separation
227 of 0.023. Sediment is added in three 0.8 Myr (8000 time steps) stages (S1-S3, Table 1). Sedimentation rates
228 in nature are extremely variable (Sadler 1981). Here, the sedimentation rate was varied between 0.15-0.45
229 mm/year to match Cenozoic rates measured in the North Sea (de Haas et al. 1996) and the North Atlantic
230 (Whitman and Davies 1979).

232 We present results from six experiments: a baseline zero sedimentation model, and five models with variable
233 sedimentation rates and patterns (slow, intermediate and fast constant sedimentation, increasing and
234 decreasing). Increasing and decreasing sedimentation rates are used to replicate the local advance and retreat
235 of depositional sedimentary systems. Model set-up and parameters have been rigorously tested (Finch et al.
236 2003; 2004; Pichel et al. 2017; 2019) and are summarised in Supplementary Table 1.

238 **Model limitations**
239 The complex three-dimensional processes occurring in salt basins are simplified into a two-dimensional
240 model for this study. Therefore, we assume our models represent a cross-section through a three-
241 dimensional linear salt wall or radially symmetric diapir; this is likely to be an oversimplification based on
242 the complex, often asymmetric, geometries of salt structures (Hudec and Jackson 2007). Modelling in two-
243 dimensions also assumes that all processes (such as salt withdrawal and stratigraphic bed rotation) are equal
244 in all directions, which is an oversimplification (e.g. Ismail-Zadeh et al. 2004; Dutta et al. 2016; Jackson and
245 Hudec 2017; Mattson et al. 2020; Pichel and Jackson, 2020). However, this is a suitable assumption in
246 simple models that focus primarily on the role of sedimentation rate variability on the halokinetic
247 depositional record. In order to prevent circular reasoning, our model inputs do not attempt to recreate a
248 specific real world diapir, but rather a simplified universally applicable structure. The absence of more
249 complicated salt geometries (e.g. salt overhangs or welds) prevent direct comparison of the models to
250 specific settings with complicated three-dimensional salt structures (e.g. the Gulf of Mexico). However, our
251 approach allows us to generate more general, possibly transferrable insights that are applicable to global
252 salt basins. Finally, sedimentation rates are extremely variable and non-linear (Sadler 1981); thus when
253 comparing to certain analogues, 'slow' and 'fast' sedimentation (stated in Table 1) should be taken as relative
254 rates rather than absolute values. Sediments can be deposited above the diapir relief suggesting that our
255 models are most applicable to subaqueous settings, and are assumed to aggrade evenly, preventing direct
256 comparison to deltaic systems (clinoforms).

257 **RESULTS**

259 The model with intermediate sedimentation rate (M3) is first presented (Figure 3) to examine the
260 relationship between the rates of salt diapir rise sedimentation. Subsequent sections describe and compare
261 diapir growth, deformation and stratigraphic architectures across all models (Figure 4-8; Tables 2, 3).

263 **Summary of temporal relationship between halokinesis and sedimentation**

265 The initially horizontal basal salt layer thins adjacent to the diapir, and the diapir geometry changes from
266 initially triangular/sinusoidal to vertically elongated during evolution (Compare H with A in Figure 3); this
267 is consistent across all 6 models (Figure 3-6). Up to 45% thinning of the salt layer, ~450 m either side of
268 the diapir base, occurs in all final model outputs (Figure 3H). The growth of the diapir is accompanied by
269 withdrawal effects in the adjacent stratigraphy, which is indicated by thinning of the source layer and
270 faulting in the basal layers (layers 1-3, Figure 3; M3 on Figure 5B). Salt withdrawal and evidence for upwards
271 salt growth are shown by the basal part of the diapir narrowing between T=0 and T=2.2 Myr (0 to 22,000
272 time steps). Sediment is first introduced to all models after 2.2 Myr (22,000 timesteps; Figure 3B). Up to
273 this time, the diapir has risen such that 'pre-kinematic' layers 4 and 5 are folded during its rise and thin

274 dramatically towards the salt due to structural attenuation, accommodated by layer-parallel slip. Pre-
275 kinematic layers, originally overlying the diapir (layers 6-9, Figure 3), are passively folded as the diapir rises.
276 This results in 'post-depositional' layer thickening on the flanks and thinning over the crest (Figure 3B).
277 Above the diapir, a topographic high is generated with associated faulting in layers 6-9 observed in T=2.2
278 Myr (Figure 3B). The topographic high influences subsequent sedimentation (layers A-L) and is the focus
279 of our study of stratigraphic modulation (Figure 3C-H).
280
281 Sediment is added in Stage 1 (S1, T= 2.2 Myr, 22,000 time steps; Figure 3C), filling two salt withdrawal
282 basins either side of the diapir. For simplicity, following the methodology of Pichel et al. (2017), the first
283 syn-kinematic layer, Layer A, fills to a flat base level in all models, and subsequent layers are input by
284 assigned sedimentation rate (in the case of M3, 0.3mm/yr). Later in S1 (T= 3Myr, 30,000 time steps, Figure
285 3), the rate of deposition outpaces diapir rise. Layers extend across the salt rather than onlapping it (Figure
286 3C, D). Throughout diapir rise, pre-kinematic stratigraphy is rotated away from the diapir crest, thickening
287 in the adjacent depocentres, due to structural layer-parallel slip. The diapir stem narrows throughout
288 evolution (compare H and A, Figure 3). The thinning of pre-kinematic layers above the rising diapir
289 continues into S2, where pre-kinematic layer 5 approaches vertical at the salt-sediment interface (Figure 3E,
290 F). Upward movement of the diapir is associated with, and driven by, salt withdrawal underneath the basins,
291 and increased displacement of faults at the base of the model (layers 1-4; Figure 3; M3 in Figure 5B).
292
293 In the final stage (S3), the pre- and syn-kinematic stratigraphy thin above the diapir crest, and are further
294 rotated and thicken into the salt withdrawal basins (Figure 3G, H). Faulting is present above the diapir tip
295 propagating through the pre-kinematic and into the syn-kinematic sediment (M3 in Figure 5B). Throughout
296 diapir evolution, the dip of the pre-kinematic layers increases towards the structure, and thus the
297 overburden anticline steepens and narrows as sedimentation progresses (compare H with B in Figure 3).
298 The deformation in the pre-kinematic overburden, described here, is similar across all models (M1-6).
299
300 The fault furthest from the diapir is taken as the edge of the halokinetically-influenced part of the
301 succession, which is ~1150 m wide on either flank (from diapir centre to fault edge), resulting in an ~2300
302 m zone of diapiric-influence, in all models. Outside this zone, pre- and syn-kinematic strata appear
303 undeformed. Salt mobility has a limited influence on sediments at the extremities of M3, which is consistent
304 across all models, so in subsequent figures only the central 3000 m is shown (grey box, T = 4.6 Myr, Figure
305 3).
306
307 To permit comparison of stratigraphic variability across all models, subsequent figures (Figure 4-6) present
308 all models (M1-M6) at the end of the experiment (T=4.6 Myr). In the following section, we describe and
309 compare diapir growth and roof folding, and stratigraphic architecture. In each case, we first present and
310 discuss M1, the case where there is no sediment input for comparison with models in which sedimentation
311 rate is constant (M2-M4), increasing (M5) and decreasing (M6).
312
313 **Diapir growth and roof folding**
314
315 *Diapir Rise with No Sedimentation (M1)*
316
317 In M1, the diapir crest rises a total of 425 m, and the final width of the pre-kinematic folded roof, taken
318 from the greatest change in dip in pre-kinematic layer 9 is 961 m (Figure 4; Supplementary Figure 1). These
319 values represent the base-case to compare the effect of different sedimentation rates on the final geometry
320 of the salt structure and the pre- and syn-kinematic stratigraphic architectures (Table 2, 3; Figure 5, 6).
321
322 *Constant sedimentation rates (M2-M4)*
323
324 Under slow (M2), intermediate (M3) and fast (M4) constant sedimentation rates the diapir rises by 393 m,
325 363 m and 297 m, respectively (Figure 5). Diapir growth compared to M1 is reduced by 8%, 15% and 30%
326 for the different aggradation cases, respectively (Table 2). The width of the pre-kinematic folded roof
327 decreases from 961 m in M1 to 770 m (M2), 760 m (M3) and 734 m (M4), which accounts for 20-24%
328 reduction relative to M1. Where syn-kinematic stratigraphy is present across the model (i.e. not in M2), the
329 syn-kinematic folded roof is measured from the point within layer F where there is the greatest change in

330 dip (Supplementary Figure 1). Layer F is chosen as it is the first layer that is laterally extensive across all
331 models (M3-6). The syn-kinematic folded roof is 839 m and 890 m wide in M3 and M4, respectively. The
332 syn-kinematic roofs are therefore 110% and 120% greater than the width of the pre-kinematic folded roof
333 in the same models (Supplementary Figure 1).
334
335 *Variable Sedimentation Rates: Increasing (M5) and Decreasing (M6) sedimentation rates*
336
337 In M5, under increasing sedimentation rate, the diapir rises by 368 m, a reduction of 14% compared to M1.
338 Under decreasing sedimentation rate conditions (M6) the diapir rises by 346 m, a 19% reduction when
339 compared to the base case M1 (Table 2; Figure 6, 7). The pre-kinematic folded roof width is reduced to
340 ~750 m in both models, a decrease of 22% compared to M1 (Table 2; Figure 7). The syn-kinematic folded
341 roof is 723 m and 872 m in M5 and M6, respectively, representing a 4% reduction and 16% increase
342 compared to the pre-kinematic folded roof in the same model.
343
344 **Effect of sedimentation rate on stratigraphic architecture**
345
346 Here, we discuss the lateral extent, thinning rates, and termination styles of stratigraphy onto the
347 topographic high for M2-6. In all models, layer A fills to a fixed base level, and therefore does not represent
348 the sedimentation rate of S1, so the first layer described is layer B. Stratigraphic architecture and thinning
349 rates (Table 3) are discussed with reference to three points that remain fixed throughout all models: the
350 crest, flank and undeformed zone (see uninterpreted M2 in Figure 5).
351
352 *Constant sedimentation rates (M2-M4)*
353
354 Syn-kinematic layers (B-L) are not laterally extensive in M2; this model is defined by slow sedimentation
355 rates. In this case, deposition is restricted to primary salt withdrawal basins, with no sedimentation occurring
356 over the crest of the pre-kinematic anticline. All layers terminate adjacent to the diapir, with the uppermost
357 layers (I –L) onlapping the remnant topography created by the layers below (Figure 5). The entire
358 stratigraphic package thins by 34%, at a rate of 0.029%/m from the undeformed zone to salt flank, before
359 pinching-out towards the crest.
360
361 In M3, the model defined by intermediate sedimentation rates, the earliest syn-kinematic strata (Layers B
362 and C) are preserved only in the withdrawal basin, offset some distance from the diapir (Table 3). However,
363 in contrast to M2, layers D-L are laterally extensive across the model, extending across the diapir crest
364 (Table 3). The overall stratigraphic thinning for intermediate aggradation is 55% from undeformed to crest,
365 at a rate of 0.037%/m, with 35% of this occurring between the undeformed and flank, accounting for 19%
366 thinning at a rate of 0.017%/m. Thinning rates of 0.1%/m, totalling 35% stratigraphic thinning, are
367 observed from flank to crest. The thinning rate between the salt flank and the crest is ~ 6 times greater
368 than that between the undeformed section and the flank (Table 3).
369
370 In the fast sedimentation model, M4, all layers are extensive across the model except for layer B (Figure 5;
371 Table 3). The overall stratigraphic thinning for fast aggradation is 33% from undeformed to crest at a rate
372 of 0.022%/m. Thirty-five percent of this total thinning is between the undeformed section and the salt
373 flank, accounting for 12% stratigraphic thinning at a rate of 0.01%/m, and the other 65% occurs between
374 the salt flank and the crest with a thinning rate of 0.06% accounting for 21% stratigraphic thinning. The
375 thinning rate between the salt flank and the crest is ~6 times greater than that between the undeformed
376 section and the flank (Table 3).
377
378 *Variable Sedimentation Rates: Increasing (M5) and Decreasing (M6) sedimentation rates*
379
380 In M5 (increasing sedimentation rate), layers B-D represent slow sedimentation; these layers are restricted
381 to the salt withdrawal basin either side of the structure and thin-towards the salt influenced topography
382 before pinching out. Layers E-H were deposited under intermediate sedimentation rates. Layer E does not
383 extend across the model and thins towards topography between the undeformed section and the salt flank,
384 before pinching out towards the crest. Layers F-H are extensive across the model. Layers I-L were deposited
385 under the fast sedimentation rate, and are laterally extensive across the entire model (Figure 6; Table 3).

Overall thinning in the increasing sedimentation model accounts for 58% total stratigraphic thinning at an average rate of 0.039%/m; 67% of this thinning takes place between the salt flank and the crest, accounting for 39% stratigraphic thinning at a rate of 0.11 %/m. The remaining 33% of thinning occurs between the undeformed section and the salt flank, at a rate of 0.017%/m accounting for 19% stratigraphic thinning. The thinning rate between the salt flank and the crest is 6.6 times greater than that between the undeformed section and the salt flank (Table 3).

In M6 (decreasing sedimentation rate), layers B-D represent fast sedimentation. Layer B is isolated to either side of the diapir and thins between the undeformed section and the flank before terminating towards the diapir (Table 3). Layers C and D are laterally extensive. Layers E-H represent intermediate sedimentation, and are deposited across the entire model. Layers I-L are deposited under slow sedimentation and are extensive across the whole model, but thin markedly over the crest (Figure 6; Table 3). The overall stratigraphic thinning for the decreasing sedimentation rate model is 52% at a rate of 0.035%/m. Thirty-five percent of this thinning takes place between the undeformed section and the diapir flank, accounting for 19% stratigraphic thinning at a rate of 0.016%/m, and the remaining 65% thinning occurs between the flank and the crest, at a rate of 0.097%/m, accounting for 34% stratigraphic thinning. The thinning rate between the salt flank and the crest is ~6 times greater than that between the undeformed section and the salt flank (Table 3).

SUMMARY OF HALOKINETIC INFLUENCE

Diapir growth and roof folding

In our model, we invoke a constant upward movement, or growth, of the salt diapir. Therefore, our models show how different sedimentation rates can dampen the late-stage growth of diapirs. The greatest upward movement of the diapir is observed in the model with no sedimentation (M1) (Table 2; Figure 7), because in this case there is no roof to resist the upward flux of salt. M1 is taken as the base case. Upward movement (growth) of the diapir is reduced with the addition of sediment (Table 2; Figure 5, 6), which in essence increases the roof thickness towards the limit where diapir growth can occur (i.e. diapir height approaches <66% of overburden: Schultz-Ela et al. 1993). The amount of growth reduces, compared to the base case (M1), with increasing sedimentation rate (being limited to 70% in M4), showing that sedimentation rate is a key control on burial of salt topography. This observation is in agreement with existing models (Giles and Lawton 2002; Hudec and Jackson 2007; Fuchs et al. 2011; Jackson and Hudec 2017; Peel et al. 2020). We also note that the anticline defining the *pre-kinematic* folded roof is widest in M1 and decreases with increasing sedimentation rate (Table 2; Figure 7, 9). However, the anticline within the *syn-kinematic* folded roof is widest in M4, appearing to increase with increasing stratigraphic thickness (i.e. sedimentation rate). In models with added sedimentation (M2-M6), the pre-kinematic roof anticline is 20-24% narrower than in M1 (no sedimentation). The lack of variability between different sedimentation conditions may imply that sedimentation rate has only a minor control on pre-kinematic anticline width, with other controls such as salt supply, salt viscosity and regional tectonics (not modelled) being more important (Koyi 1998; Fuchs et al. 2011). Syn-kinematic folded roof thicknesses are more variable across the models, because they are controlled by the sedimentation rate. Increasing syn-kinematic folded roof width with sedimentation rate supports fold wavelengths being larger for thicker overburdens (Davison et al. 2000b; Bonini 2003; Hudec and Jackson 2011; Duffy et al. 2018).

Fault distribution and deformation zone extent

Fault distribution and withdrawal basin extent are interpreted using the nearest neighbour outputs (Figure 4; Supplementary Figure 2). Nearest neighbour outputs highlight the amount of displacement that has occurred during 4.6 Myr, relative to an element's initial neighbours; this is used to highlight discontinuities as a proxy for faults. Fault distributions are broadly similar across all models (Figures 4-6) and are summarised here. Numerous predominantly extensional faults, with variable dip directions, are identified in the pre-kinematic layers in all experiments. These faults have displacements of metres to 100's of metres, with the greatest throws being observed between pre-kinematic layers 1-7 (Figures 4-6). Steep structures appear to develop over the crest of the growing diapir, but are difficult to decipher in terms of slip style (i.e. normal or reverse) due to the relatively small number of displaced neighbouring elements. These crestal structures extend into the syn-kinematic strata (layers A-G) overlying the overburden anticline (M3-M6);

441 these discontinuities are largest and extend furthest into the syn-kinematic overburden under greater
442 sedimentation rates (compare M4 with M3; Figure 5). In all experiments, pre-kinematic layers 8 and 9 are
443 dominated by small-scale faults that are localised to those layers.
444
445 The edge of the salt withdrawal basin is taken as the distance of the furthest faults from the diapir; outside
446 this zone the strata are not deformed (Figure 4-6). The salt withdrawal basin, and associated deformation
447 zone, is ~2300 m wide in all models, accounting for three times the initial maximum width of the diapir
448 (Figure 4-6). The similar extent of the withdrawal basin across all experiments suggests that syn-kinematic
449 sedimentation only has a minor control on deformation of pre-kinematic layers and structural
450 configuration.
451
Stratigraphic architecture variability with sedimentation rate
453
454 Here, the variability in thinning rate is compared between different models with syn-kinematic
455 sedimentation. Layer A is excluded from descriptions in all models, and thus from our comparisons, as it
456 fills to a linear, instantaneous base level and therefore does not always represent the sedimentation rate of
457 S1. Therefore 11 layers (B-L) are described and compared in Table 3.
458
459 Under slow sedimentation rate (M2) all 11 layers onlap topography and are not laterally extensive. Under
460 intermediate sedimentation rate (M3) two layers onlap and nine are laterally extensive, rising to 10 laterally
461 extensive and one onlapping in the fast (M4) and decreasing (M6) sedimentation rate cases. Under
462 increasing sedimentation rate (M5), the initial four layers (including three which are deposited under slow
463 sedimentation) onlap topography and the remaining 7 are laterally extensive (Figure 5). As expected, layers
464 are more laterally extensive under higher sedimentation suggesting that the effects of halokinetic
465 modulation decrease more rapidly upwards under higher sedimentation rates (Peel et al. 2014; Sylvester et
466 al. 2015; Cumberpatch et al. 2021b). The final stratigraphic thickness is greatest under fast sedimentation
467 rate (M4, 570 m), least under slow sedimentation rate (M2, 228 m), and at a similar intermediate level for
468 intermediate (M3, 401 m), increasing (M5, 441 m) and decreasing (M6, 382 m) sedimentation rates, logically
469 showing that net sediment volume is the most important control on sediment thickness (Figure 7, 8, 9).
470
471 Overall stratigraphic thinning, from the undeformed to the flank, is greatest under slow sedimentation rate
472 (M2, 34% thinning) and least under fast sedimentation rate (M4, 12% thinning) (Table 3). This shows that
473 stratigraphic thinning rates decrease with increasing sedimentation, suggesting that diapir modulation
474 (thickness and dip variability) decreases quicker under higher sedimentation rate (Pratson and Ryan 1994;
475 Koyi 1998; Fuchs et al. 2011).
476
477 In all models, the thinning rate is ~ 6 times greater between the salt flank and the crest than it is between
478 the undeformed section and the salt flank (Table 3; Figure 9). This rate is higher (6.6 times) under increasing
479 sedimentation (M5), which suggests that salt structures have greater influence in models where
480 sedimentation rate is initially slow (e.g., S1 in M5; Giles and Lawton 2002; Fuchs et al. 2011; Giles and
481 Rowan 2012; Jackson and Hudec 2017).
482
483 Whilst the primary mechanism modulating the stratigraphic architecture is stratigraphic thinning,
484 stratigraphic thickening into basins driven by salt withdrawal at depth must not be disregarded. Localised
485 thickening into salt withdrawal basins is observed in all models. Such thickening is observed in all models,
486 accounting for 13% of the stratigraphic thickening in M5 (Cumberpatch et al. 2021b). This highlights how
487 the presence of a growing diapir can be associated with localised additional accommodation (as well as a
488 reduction of accommodation) due to the at-depth evacuation of salt from the source layer to feed the
489 growing diapir, as is evidenced by subtle thickening into the basin (Figure 5; 6). Accommodation reduction
490 over the crest of the diapir is driven by diapir growth, which is recorded by stratigraphic thinning.
491 Accommodation increases in salt withdrawal basins are accounted for by stratigraphic thickening and salt
492 migration at depth.
493
494 Through time, thickening and thinning are eventually reduced as the halokinetic modulation on stratigraphy
495 is minimised with the burial of the salt-cored topographic high and its flanking depocentres. In all
496 experiments with layers that extend across the entire model, the thinning rate and bedding orientation

497 change up-stratigraphy (Table 3, Figure 8). In M3-M5, a decrease in thinning rate up-stratigraphy is
498 observed. Bedding orientations are variable but generally decrease (flatten) upwards in M2-5. However,
499 under decreasing sedimentation (M6) an overall increase in thinning rate up-stratigraphy is observed (Figure
500 6; Table 3), in addition to a slight increase in bedding dip between layers J, K and L. This suggests that
501 initially thicker layers are less deformed, but that as diapir growth continues and sedimentation decreases,
502 thin layers are still influenced by topography associated with rising salt (Giles and Lawton 2002; Hudec and
503 Jackson 2007; Giles and Rowan 2012; Sylvester et al. 2015; Soutter et al. 2019).

505 Overall, halokinetic modification reduces with increasing sedimentation rate as halokinetic bathymetry is
506 buried. Typically, such alteration decreases up stratigraphy, and laterally outwards from the diapir in
507 agreement with outcrop and subsurface analogues globally (Figure 9, 10, 11; e.g. Pratson and Ryan 1994;
508 Giles and Lawton 2002; Mayall et al. 2010; Giles and Rowan 2012; Kernen et al. 2012; 2018; Banham and
509 Mountney 2013a; 2013b; 2014; Oluboyo et al. 2014; Poprawski et al. 2014; 2016; Ribes et al. 2015; Doughty-
510 Jones et al. 2017; Wu et al. 2020; Rodriguez et al. 2020; Cumberpatch et al. 2021a).

512 **DISCUSSION**

514 **Comparison to natural examples**

516 A key challenge for numerical models is ground-truthing against natural prototypes (Oreskes et al. 1994;
517 Burgess 2012). Here, the key findings and predictions from the DEM (Figure 9) are compared to published
518 subsurface (Figure 10,11) and field analogues (Figure 12) to understand their applicability and limitations.

520 *Comparison to subsurface: stratigraphy around the Pierce diapirs, Eastern Central Graben, UK North Sea*
521 Jurassic-Pleistocene syn-kinematic stratigraphy around the north and south Pierce diapirs, Eastern Central
522 Graben, UK North Sea (Figure 10) shows evidence for halokinetically driven changes in bed thickness and
523 dip (Figure 11C, 10E; Birch and Haynes 2003). Pierce's tectonostratigraphic history spans ~200 Myr and
524 is summarised as Jurassic reactive-active diapirism, followed-by Cretaceous-Cenozoic passive diapirism,
525 and contraction-driven active diapirism during Alpine compression (Scott et al. 2010). Despite a longer-
526 lived and more complex evolution, the Pierce diapirs show geometrical similarities with several of our
527 DEMs (Figure 9, 10). In the Pierce example, stratigraphy is near horizontal ~2 km away from both diapirs,
528 but is upturned adjacent to them, comparable with model results (Figure 5, 6, 9, 10). The generation of
529 brittle deformation throughout the Cenozoic stratigraphy over the crest of both north and south Pierce
530 (Figure 11; Carruthers et al. 2013) corresponds to high zones of relative displacement across the crest of
531 the diapir, in M3-M6, which extend into the syn-kinematic stratigraphy (Figure 5, 6, Supplementary Material
532 1). Similar crestal deformation has been demonstrated in physical models (e.g., Davison et al. 1993).
533 Different model outputs may be applicable to different parts of stratigraphy due to the changing ratio of
534 diapir rise rate and sedimentation rate. For example, stratigraphic architectures comparable to M2, M3 and
535 M5 are observed in different stratigraphic packages around the Pierce diapirs (Figure 11).

537 The unique observation from M2 (slow sedimentation) is non-extensive layers that thin towards and
538 eventually onlap and pinchout against the long-lived, salt-related topographic high (Figure 9A). M2 results
539 are analogous to the Top Cretaceous (lime green) to Mid Eocene (red) stratigraphy around the Pierce diapirs
540 (Figure 11C), which is not laterally extensive across the diapirs (Davison et al. 2000a; Birch and Haynes
541 2003; Scott et al. 2010; Carruthers et al. 2013). The Top Cretaceous – Mid Eocene package, equivalent to
542 S1-S3 in the models, thins significantly towards both diapirs, before onlapping the flanks (Figure 9A; 10A;
543 10C). The amount and rate of thinning reduce through time, from 51% at 0.044%/m in the Paleocene (Top
544 Cretaceous – Top Lista interpretations) to 21.3% at 0.019%/m between in the Eocene (Tay and Eocene
545 interpretations (Supplementary Table 2) in agreement with model observations showing thinning rates
546 reducing upwards. These values are similar to the overall thinning of the slow sedimentation model (34%,
547 at 0.029%/m; Table 3). Despite this apparent similarity in thinning rate values, regional sediment volumes
548 are high throughout the Palaeocene (10,000 km³/my; Lui and Galloway 1997). Specific lithostratigraphic
549 units such as the Forties Sandstone are associated with ~200 m of sandstone deposited in c. 1 million years
550 (Kilhams et al. 2014; Eldrett et al. 2015). Sedimentation rates for the UK Palaeocene stratigraphy are
551 uncertain, and spatially and temporarily variable around the Pierce diapirs, with rates ranging from 0.085 -

0.4 mm/yr (Liu and Galloway 1997; Kilhams et al. 2014; Eldrett et al. 2015). The lower values are in broad agreement with our slow sedimentation input values. The upper end, however, indicates intermediate – fast sedimentation rates in our models, which have probably been modified by very high diapir rise rates to form an overall geometry typical of slow sedimentation rates (Carruthers et al. 2013). This is further evidenced by the steep upturning of stratigraphy adjacent to the diapirs (Figure 11C; Giles and Lawton 2002; Hudec and Jackson 2007). In our model, we isolate and vary sedimentation rate, whilst in reality the dynamic ratio of sedimentation rate and diapir rise rate control stratigraphic architecture. It is suggested that during the Paleocene – Eocene sedimentation rates were high, but diapir rise rates were higher, likely driven by sediment loading (Carruthers et al. 2013) creating an overall effect of a 'low sedimentation rate' akin to our M2.

In M3 (intermediate sedimentation rates), layers are initially non-extensive (S1), displaying early onlap and thinning (26.4% at 0.023 %/m; Table 3; Figure 9B; 10B) similar to Eocene (orange) to Oligocene (peach) stratigraphy adjacent to the north Pierce diapir (Figure 11), which thin by 16% at 0.014 %/m. Subsequent modelled layers (S2) are extensive but thin towards the diapir high (60% at 0.04 %/m; Table 3), analogous to the Oligocene (peach) to Mid-Miocene (light peach) across north Pierce (65% at 0.044 %/m; Supplementary Table 2; Figure 11C; Carruthers et al. 2013). In our model simulations, 76% of this thinning occurs between the undeformed and the flank location, similar to subsurface observations of 85% thinning around north Pierce. After this initial modulation (S1, S2), subsequent M3 stratigraphy (S3) records a reduction in thinning rates and halokinetic influence on stratigraphy upwards (reduction from 60% to 23% total thinning: Table 3; Figure 9B; 10B). This is also observed, albeit at a less extreme rate, between the Middle Miocene and Pliocene interval around the Pierce diapirs (reduction from 65% to 49% total thinning; Supplementary Table 2; Figure 11). Much of the excess thinning in the Pierce example, compared to the model, occurs between the undeformed and flank position (28% of thinning, compared to 6% in M3; Table 3, 4). This is likely due to Cenozoic compressional forces driving diapir growth and upturn of stratigraphy (Birch and Haynes 2003; Scott et al. 2010; Carruthers et al. 2013), resulting in a less gradual reduction in halokinetic alteration upwards with respect to M3.

Based on this comparison of the lateral extent of layers, it is possible to infer that these examples represent intermediate sedimentation rates, relative to diapir rise rate. Sedimentation volumes for the Eocene to Oligocene are ~4000 km^3/Myr, lower than for the Top Cretaceous-Eocene. This suggests that overall diapir rise rate had been slower during the Eocene to Oligocene, with respect to the Top Cretaceous-Eocene time, giving a stratigraphic architecture typical of lower diapir rise rates relative to sedimentation rates (Carruthers et al. 2013).

In M5 (increasing sedimentation rate), S1 and early S2 are initially isolated either side of the diapir in salt-withdrawal basins and onlap salt-cored topography (Table 3). Late S2 stratigraphy is extensive across the model, thinning towards the topographic high, with modulation decreasing up-stratigraphy (S3, Figure 9D). When combining the Cretaceous to Eocene (M2 analogue, Figure 11) and Eocene to Mid-Miocene (M3 analogue) stratigraphy adjacent to the Pierce diapirs, we observe an overall increasing upwards sedimentation trend, relative to diapir rise rate (Den Hartog Jager et al. 1993; Jennette et al. 2000; Birch and Haynes 2003; Kilhams et al. 2014). The Cretaceous to Eocene was deposited when sedimentation rate < diapir rise rate, leading to isolation of salt withdrawal basins and onlapping of the diapir flanks (S1, M5, Figure 11D; Carruthers et al. 2013). The Eocene – Mid Miocene (S2 equivalents) were likely deposited when sedimentation rates were higher relative to diapir rise rates, or when salt supply was equal to sedimentation rate (S2, M5, Figure 11D; Carruthers et al. 2013). Subsequent layers (Mid-Miocene – Pleistocene) are broadly extensive across the diapirs, similar to S3 in M5. The sedimentation volume for this interval is fairly low (~2000 km^3/Myr; Liu and Galloway 1997), so this observed reduction in halokinetic influence upwards is likely driven by a slowing or cessation of diapir growth rather than being driven purely by high sedimentation rates (Carruthers et al. 2013). Pliocene to Pleistocene stratigraphy is extensive across north Pierce (S2 and S3, M5), but only the Pleistocene is present extensively across south Pierce, due to differential diapir growth histories and cessations (Scott et al. 2010; Carruthers et al. 2013). The upwards reduction in sedimentary thinning observed in the increasing sedimentation model (61% in S2 to 22% in S3) is similar to that observed between the Eocene to Mid-Miocene (88%) and the Mid-Miocene to Pleistocene (18%) in the subsurface example, highlighting the potential applicability of the results from our models.

In nature, halokinetic sequence architecture is controlled by the dynamic ratio between sedimentation rate and diapir rise rate (e.g. Giles and Rowan 2012; Pichel and Jackson, 2020), such that an 'apparent' increase in sedimentation (reducing halokinetic influence upwards) could represent a slowing of diapir growth due to regional tectonic quiescence or depletion of the salt source layer. Our observations and comparison to DEMs with variable sedimentation rates, are consistent with diapir rise to sedimentation rate ratios derived from halokinetic sequence studies (Carruthers et al. 2013). The role halokinesis plays in shaping the stratigraphic architecture of the north and south Pierce diapirs is reduced from the Oligocene and Pliocene respectively, due to source layer depletion, resulting in halokinetic bathymetry being gradually buried (Figure 11; Birch and Haynes 2003). Integration of, and comparison between DEM and subsurface data, demonstrates the importance of understanding local (salt layer variations) and regional (tectonic and sedimentation rate) controls when disentangling salt basin evolution.

Comparison to outcrop: stratigraphy around the Bakio diapir, northern Spain

The model simulations presented here document stratigraphic architectures and structural deformation, but do not invoke a specific sedimentary environment. Different compositions (e.g., carbonate or siliciclastic) and sedimentary environments (e.g., fluvial or deep-water) will respond differently to salt influence (Adams and Kenter 2012). Below, we compare the observations of M5 (increasing sedimentation) to a well-exposed halokinetically-influenced deep-water succession, and describe how integrating DEM results with outcropping examples can help reduce uncertainty in subsurface energy reservoir prediction.

Exposed Aptian-Cenomanian strata adjacent to the Bakio Diapir, northern Spain document an increase in sedimentation rate associated with an increase in erosion of the hinterland (Figure 12; García-Mondéjar 1996; Martín-Chivelet et al. 2002; Puelles et al. 2014). The resulting progradational pattern manifests in the deep-water succession as an upward change from thin-bedded low-density turbidites, deposited in the lobe fringe, to thick-bedded high-density turbidites deposited in a lobe axis (Cumberpatch et al. 2021a). Pre- (the Urgonian limestone: Figure 12) and syn- kinematic depositional elements are deformed closest to the diapir, and deformation intensity decreases away from the diapir, being minimal outside the halokinetically influenced sequence (~700 m wide either side at Bakio); this observation is consistent with the predictions of our model (Figure 6, 12). Outside the zone of halokinetic deformation, in terms of their architecture and facies distributions, strata look similar to those deposited in an unconfined, salt-free setting (e.g., Prélat et al. 2009: Figure 12D).

The Aptian-Albian is initially isolated either side of the diapir, with individual beds thinning and pinching out towards the salt-cored topographic high (S1, Figure 12) (Cumberpatch et al. 2021a). Under increasing sedimentation rates, the Albian-Cenomanian stratigraphy is laterally extensive and stratigraphy show a reduction in the numbers of mass transport deposits (MTD) upwards. The stratigraphy also show a reduction in halokinetic deformation and thinning rate upwards (S2), eventually healing remnant diapiric topography (S3). In the syn-kinematic stratigraphy, thick-bedded sandstones deposited in channels and lobes dominate topographic lows, where sedimentary flows were around salt-cored topographic relief (Figure 12D; Mayall et al. 2010; Sylvester et al. 2012; Doughty-Jones et al. 2017; 2019; Rodriguez et al. 2020). Towards the highs, the lower-density part of the flows may run-up topography depositing thinly-bedded muddier sandstones towards the pinchout (Figure 12E; Kneller and McCaffrey 1999; Soutter et al. 2019). Crestal deposition is limited due to elevation, until sedimentation period 3 (S3) where halokinetic bathymetry is healed.

Implications for subsurface energy

Despite advances in extent and resolution of 3D seismic reflection imaging, the salt-sediment interface remains difficult to image, yet determining its position and precise geometry is crucial in helping to appraise stratigraphic-structural traps for hydrocarbon, carbon storage, and geothermal prospects globally (Jones and Davison 2014; Warren 2016). Utilising outcrop analogues (Figure 12) can help provide sub-seismic scale depositional facies information, helping reduce uncertainty in reservoir quality and distribution. Numerical modelling results do not represent specific analogue conditions nor a 'snapshot' in time, and they can therefore help to quickly identify generic depositional architectures, deformation patterns and sediment thickness relationships as a function of several forcing parameters, such as variations in sedimentation rates.

664
665 Using stratigraphic architectures from our DEM and sedimentological data from field examples, we can
666 improve predictions of the likely architecture of syn-kinematic stratigraphy and sedimentology around salt
667 structures, which are poorly-imaged in seismic reflection data. Understanding this requires a multi-scalar
668 and multi-technique approach. For example, models provide details about gross thickness changes and
669 geometry, whereas field analogues enable inferences about reservoir quality and net-to-gross.

670

671 *Depositional reservoir quality*
672 Regardless of the amount of sedimentation, our models show that stratigraphy thins as it approaches the
673 diapir, suggesting a reduction in the amount of total reservoir close to the structure (Figure 8; Jackson and
674 Hudec 2017). Siliciclastic depositional environments show a thinning of sandstone towards the topographic
675 high and an overall concentration of high reservoir quality units at the base of topography (Figure 12A; D),
676 such that a salt-related combined structural-stratigraphic trapping mechanism becomes unlikely (Figure 13;
677 Kane et al. 2012; Stricker et al. 2018). Muddier (lower reservoir quality) and thinner (lower net-to-gross)
678 units are expected closer to the diapir (Figures 8; 10E; 11; Banham and Mountney 2013; Cumberpatch et
679 al. 2021a). These units are more likely to be over-pressured due to upward rotation, creating a large pressure
680 head, with the topseal rocks unable to hold back a significant hydrocarbon column (Figure 13; Nikolinakou
681 et al. 2014a; b; 2018; Heidari et al. 2017; 2019). In carbonate environments, shallow platform or reef facies
682 with excellent reservoir potential may be present over salt highs (e.g. Riding 2002; Burgess et al. 2013;
683 Teixell et al. 2017). As salt growth continues, fractures are generated in the overburden, which could form
684 significant secondary porosity within the carbonate reservoirs, increasing quality and producibility (He et
685 al. 2014; Howarth and Alves 2016; Saura et al. 2016).

686

687 Supplementing subsurface data with modelled stratal architectures and depositional facies observations
688 from exhumed halokinetically influenced settings globally (e.g., Banham and Mountney 2013a; 2013b; 2014;
689 Poprawski et al. 2014;2016; Counts and Amos 2016; Ribes et al. 2015; 2017; Counts et al. 2019;
690 Cumberpatch et al. 2021a) is recommended as a useful workflow for building reservoir models for salt
691 basins with limited data. Observations from multiple-scales and types of models can be combined to further
692 reduce the uncertainty associated with reservoir quality prediction, for example recent finite element
693 modelling (FEM) has shown porosity is lower than expected near the vertical parts of salt structures and
694 higher than expected at the base of diapirs, due to mean principal stress variations (Nikolinakou et al. 2014).

695

696 *Halokinetic zonation*
697 The model results show that a deformation zone exists either side of the diapir in all experiments (Figure
698 4, 5, 6). Outside of this zone, the syn- and pre-kinematic stratigraphy are undeformed (Figure 13). The
699 extent of this salt withdrawal basin is 1150 m on either side of the diapir (2300 m in total). Therefore, the
700 total zone of halokinetic influence in all models is approximately three times the original diapirs' maximum
701 width (750 m), with a deformed zone of 1.5 diapir widths either side of the structure. The width of the
702 deformation zones is comparable across all models (Figure 7), and therefore it is shown that sedimentation
703 rate is unlikely to have a significant control on the width of the zone of halokinetic influence (Giles and
704 Rowan 2012; Hearon et al. 2014). Other factors such as salt supply, salt viscosity, and style and magnitude
705 of regional tectonics (which are not modelled) will in nature, influence the width of the halokinetically
706 deformed zone (Koyi 1998; Fuchs et al. 2011). The model can be further divided into zones based on onlap
707 geometry and thinning rates, which highlight the 'trade off' between reservoir thickness and stratigraphic
708 trap potential in subsurface plays (Figure 13). In the flank locations, bedding dips and thinning rates are
709 shown to be greater under slower sedimentation rates compared to higher sedimentation rates (Figure 8;
710 Table 3), which is important when predicting hydrocarbon column height. Significant overpressures on
711 reservoirs below can be created by fast sedimentation rates (Figure 7; 8; Peeters et al. 2018).

712

713 *Fault distribution*
714 Radial faults associated with salt diapirs are shown to cause compartmentalisation of reservoirs (Birch and
715 Haynes 2003; Scott et al. 2010; Charles and Ryzhikov 2015; Peeters et al. 2018; Coleman et al. 2018). Our
716 DEM replicates localised fault growth, evolution and propagation because the contacts between elements
717 are treated as potential displacement surfaces. Our models all document a similar fault pattern, suggesting
718 that faulting, and therefore fault compartmentalisation is not heavily influenced by sedimentation rate. As
719 well as seismically resolvable faults, outcrop and borehole data indicate brittle deformation is significant in

13

salt basins (e.g. Koestler and Ehrmann 1991; Cumberpatch et al. 2021a). Our DEM replicates this brittle deformation (Figures 4-6); as in nature, extreme thinning, and termination of layers is in part accommodated by small-scale displacements. Understanding sub-seismic scale fault distribution is important for predicting reservoir compartmentalization and seal integrity in the subsurface. Faults, when sealing, could act as lateral permeability barriers, especially if the faults and surrounding reservoir rocks become cemented with salt and salt-related breccia (Van Bergen and de Leeuw 2001; Li et al. 2017). DEM is therefore advantageous in predicting potential traps, conduits and baffles, due to its replication of diapir-related brittle deformation.

Future work
The DEM presented here is useful for predicting regional trends and studying generic interactions of salt structures and stratigraphy, but further work is required to recreate specific complicated salt geometries. Other suggestions for further development of this model are to incorporate reactivation and cessation of halokinesis, studying the impact on stratigraphic architectures and the development of halokinetic unconformities. In nature, salt diapirs are rarely isolated structures, and therefore subsequent studies will focus on the interaction of multiple salt structures, with different growth histories on different sedimentary successions. Many of the limitations of the model described above, are due to its two-dimensional nature. The ultimate aim of future work is developing a three-dimensional DEM to better understand the four-dimensional variability in halokinetically-influenced stratigraphy and associated subsurface energy reservoirs.

CONCLUSIONS

1) Discrete element modelling (DEM) can form an integral part of the workflow when studying salt-sediment interactions. Here, a DEM is used to study the variability in stratigraphic architecture and deformation patterns, around a growing salt structure under different sedimentation rates.

2) The models generate realistic salt-related faults. In all models, structural deformation and extent of halokinetic influence are similar, and syn-kinematic strata, at least initially, are isolated to either side of the diapir, thinning and onlapping towards the high.

3) Under slow sedimentation rate (M2) deposition is restricted to salt withdrawal basins either side of the diapir throughout evolution, while in M3-M6 sedimentation eventually occurs over the diapir crest, often associated with significant lateral thickness variability. Diapir growth is most inhibited under fast aggradation (M4) and the halokinetic influence on stratigraphy reduces quickly with time.

4) Thinning of syn-kinematic stratigraphy from the undeformed section to the diapir flanks, is greatest under slow aggradation (M2, 34%), and least under fast aggradation (M4, 12%). In all models, thinning is about six times greater between the salt flank and crest, compared to the undeformed section and the salt flank, indicating more intense deformation close to the diapir.

5) Thinning rate decreases through time (up stratigraphy), showing a reduction of halokinetic modulation with increased sediment thickness, as halokinetic bathymetry is 'healed'. This is true for all models except for decreasing sedimentation (M6), which experiences a slight increase up stratigraphy.

6) Our simplified two-dimensional models provide useful analogues for salt-influenced basins with complicated four-dimensional evolutions. Natural examples record the interplay between sedimentation rate and diapir rise rate, whilst models isolate and vary sedimentation rate. Comparison to the Pierce field diapirs, North Sea, shows how different models can be applicable to different parts of stratigraphy and suggest how interpreters could infer likely sedimentation rates and conditions from subsurface stratigraphic geometries.

7) Facies, and thus reservoir distribution, around salt diapirs will differ from unconfined settings due to halokinetic modulation, both vertically and laterally. A deep-water analogue from stratigraphy

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774 adjacent to the Bakio diapir, shows that halokinetically influenced facies (e.g., the salt flank) contain
775 thin beds, sandstone pinchouts and increased mass transport deposits in comparison with the
776 allogenic deposits (e.g., the undeformed zone), which are difficult to decipher from deep-water
777 strata in non-salt influenced settings.

779 8) Integrating DEM with subsurface and outcrop data helps to reduce reservoir and trap uncertainty
780   in subsurface energy exploration and development.

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1368 **FIGURE CAPTIONS**

1369
1370 **Figure 1:** Seismic reflection cross-section from the Pierce diapirs, Eastern Central Graben, UK North Sea
1371 (located in Figure 10) highlighting some of the key questions and uncertainties of subsurface interpretation
1372 in salt-influenced basins. Mapping top salt, visualizing pinch-outs and onlap geometries in areas of lower-
1373 quality data adjacent to the salt (1), confidently mapping top salt (2) and studying the lateral and vertical
1374 extent of syn-kinematic deformation and sedimentation (3), are challenging. Seismic data courtesy of PGS
1375 (MegaSurvey Plus 3D seismic data).

1376
1377 **Figure 2:** Initial set up of the DEM (T=0) and key parameters. Initial geometry and rise rates taken from
1378 North Pierce (Davison et al. 2000). See text for discussion and Table 1 and Supplementary Table 1 for
1379 further details.

1380
1381 **Figure 3**: Model 3, the intermediate aggradation experiment with a sedimentation rate of 0.3mm/year.
1382 Diapir growth rate is continuous and constant throughout (0.023 mm/year). Outputs begin at the start of
1383 the experiment during a stage of no sedimentation (T=0 Myr, circle S0 (sedimentation stage 0)). At T=2.2
1384 Myr (B) the diapir has generated surface topography and sedimentation begins to be added (S0/S1).
1385 Subsequent outputs are generated at 0.4 Myr intervals across the three equal stages of sedimentation (open
1386 circles S1-S3) from 2.2 to 4.6 Myr. Active sediment input is constant in stages 1-3 and the boundaries are
1387 highlighted by the grey circle (e.g. T=3 Myr represents the end of sedimentation 1 and the beginning of
1388 sedimentation event 2). The final output is at 4.6 Myr (H), following the final sedimentation from S3. Due
1389 to minimal deformation in the outer section of all models, for clarity, subsequent figures will focus on only
1390 the central 3000 m around the diapir (shown by the grey outline, T=4.6 Myr). Subsequent figures are shown
1391 at T=4.6 (H).

1392
1393 **Figure 4:** M1, the model with no sedimentary fill, after 4.6 Myr. Displayed outputs represent the central
1394 3000 m of the original output model (located in Figure 3). A) Model output, uninterpreted, pre-kinematic
1395 layers are numbered for clarity, and ease of subsequent discussion. B) Static image of element displacement
1396 relative to initial neighbours for M1 after 4.6 Myr (provided for M1-M6 in Supplementary Figure 2). Cold
1397 colours represent elements that are in contact with their original neighbour and hot colours indicate high
1398 displacement. This is used to show discontinuities and is a proxy for fault location. C) (A) overlain with (B),
1399 interpreted with locations of discontinuities, highlighting the methodology used to interpret structures for
1400 all models. Note how the majority of the internal salt has remained in connection with its original
1401 neighbouring element and the radial faults are associated with salt withdrawal. The maximum displacement
1402 shows the relative movement of the salt, the neighbouring overburden, and the high mobility of layers in
1403 close proximity to the diapir. Note the lack of deformation outside of the salt withdrawal basin. D)
1404 Interpreted static DEM for 4.6 Myr, faults are taken from discontinuities in (C), and overlain onto (A).

1405
1406 **Figure 5:** Uninterpreted and interpreted static images of the DEM for M2-M4 (constant sedimentation
1407 rate models) after 4.6 Myr. A) Uninterpreted models, sedimentary layers are coloured and assigned letters
1408 for easier discussion. Crest, flank and undeformed labels for M2 reflect the locations of thickness
1409 measurements taken to calculate thinning rates, this location is consistent across all models (M2-M6), B)
1410 Interpreted versions of each model. Fault interpretations use displacement from initial neighbour

1411 methodology (see Figure 4, Supplementary Figure 2). Interpretations highlight the width of salt withdrawal
1412 basins (deformed zones), height of diapirs, fault and fracture distributions, stratigraphic thickness variation
1413 and variable halokinetic influence vertically and laterally, discussed in text.

1415 **Figure 6:** Uninterpreted and interpreted static image of the DEM with variable sedimentation (M5 and
1416 M6) after 4.6 Myr. Displayed output represent the central 3000 m of the original output model (located in
1417 Figure 3). A) Uninterpreted models, sedimentary layers are coloured and assigned letters for easier
1418 discussion. Crest, flank and undeformed zones mentioned in text reflect those shown in M2 (Figure 5). B)
1419 Interpreted versions of each model. Fault interpretations use displacement from initial neighbour
1420 methodology (see Figure 4, Supplementary Figure 2). Interpretations highlight the width of salt withdrawal
1421 basins (deformed zones), height of diapirs, fault and fracture distributions, stratigraphic thickness variation
1422 and variable halokinetic influence vertically and laterally discussed in text.

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1426 sediment is not shown in M1 as none is added. All cases have a similar top salt outside the deformed zone
1427 (shown by a bold black line). Diapir growth and width of the pre-kinematic overburden anticline are
1428 reduced in M2-M6 when compared to M1 due to overpressure caused by sedimentation (see Table 2). Grey
1429 lines indicate 200 metre divisions vertically. Located on Figure 3.

1431 **Figure 8:** Comparison of stratigraphic thickness for flank location across all models with sediment (M2-
1432 M6) see Figure 5 for location of flank profile. Thicknesses and bedding orientation highlighted. The flank
1433 location shows the most deformation in all models, greatest modulation is observed in M2, and least
1434 modulation is observed in M4, generally modulation decreases upwards, however in M6 bed dip increases
1435 in layers K and L. Models give an indication of the sedimentary thickness and bed rotation expected
1436 adjacent to salt diapirs under different sedimentation rates, aiding prediction of sediment thickness and bed
1437 rotation in the subsurface salt sediment interface.

1439 **Figure 9:** Schematic interpretations of M2-M6, focussed on the central 2000 m of the syn-kinematic
1440 stratigraphy, located in Figure 3. Bullet points highlight the key observations specific to each modelled
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1456 et al. 2021b.
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1460 the upper part of the increasing sedimentation model (located on Figure 6). H and A are theoretical
1461 stratigraphic profiles in the halokinetically-influenced and non-halokinetically-influenced zones
1462 respectively. B) Location map of the field analogue, for a full geological discussion see Cumberpatch et al.
1463 in 2021a). C) Outcropping Triassic evaporites on Bakio Beach, believed to be part of the Bakio Diapir. D)
1464 Non-halokinetic succession showing a classic progradational deep-water system (controlled by allogenic
1465 and autogenic processes), accompanied by an example of non-halokinetically influenced strata from the

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1468
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1474 highlights that while stratigraphic trap quality may be greater closer to the diapir, reservoir thickness and
1475 quality are likely higher further from the diapir, showing a 'trade-off' exists for subsurface energy
1476 exploration and production. Similar zonation is possible for all models.

1477

TABLE CAPTIONS

Table 1: Details of the different sedimentation patterns and rates used in the six experiments. S1-3 refer to sedimentation intervals.

Table 2: Comparison of diapir growth and overburden anticline width across all six experiments at T=4.6 Myr. Measurements are taken from model simulations (Figure 4, 5 and 6). Initial diapir height (1050 m at T=0 Myr) is subtracted from final diapir height to give diapir growth. M2-6 are compared to M1 to show the percentage reduction that occurs when different amounts of sediment are added. Pre-kinematic anticline measurements are taken from the stratal termination of layer A, in M1 they are taken from the top of the overburden at the location of the greatest change in dip. For M2-M6 the width of the overburden is compared to M1 to show the percentage reduction in overburden anticline width that occurs when different amounts of sediment are added. Syn-kinematic folded roof measurement are taken from location of greatest dip change in layer F (Supplementary Figure 1). Value only calculated where layers extend across the entire structure (M3-M6). Ratio of syn- to pre-kinematic folded roof thickness is calculated.

Table 3: Comparison of stratigraphic thinning across M2-M6. Bold headings indicate each model and show the total thinning for all stratigraphy, beneath each heading this is divided by sedimentary package. Layer A is excluded as it fills to base level in all models. Layers are subdivided based on observational differences (e.g. lateral extent across the model) and are broadly grouped into their sedimentation stage (S1, S2, S3). Percentage thinning and normalised thinning rates (%/m) are shown for U-C (Undeformed to Crest, i.e., the total model), U-F (Undeformed to Salt Flank) and F-C (Salt Flank to Crest). In layers which do not extend across the entire model no results exist for U-C and F-C. The final column displays the ratio of the normalised thinning rate between the Undeformed to Salt Flank and the Salt Flank to Crest sections, to show how much more thinning is observed adjacent to the diapir. Locations of undeformed, flank and crest measurements are shown on Figure 5, and are the same for all models to ensure direct comparison.

SUPPLEMENTARY MATERIAL

Supplementary Figure 1: Explanation of diapir growth, syn-kinematic folded roof and pre-kinematic folded roof measurements shown in Table 2, and described within the text. Diagram is based on M3, and terms are consistent across all models.

Supplementary Figure 2: Static image of element displacement relative to initial neighbours for all models at 4.6 Myr. Cold colours represent elements that are in contact with their original neighbour and hot colours suggest high displacement. This is used to show discontinuities and is a proxy for fault location. Note how the majority of the internal salt has remained in connection with its original neighbouring element, and the radial faults associated with salt withdrawal. The maximum displacement shows the relative movement of the salt to the neighbouring overburden and the high mobility of layers in close proximity to the diapir. An apparent lack of deformation is observed outside of the halokinetically deformed zone. Deformation in sediment is shown over the crest and in the flank region of each model (see Figure 5 for locations of these areas). Outside of these areas, less deformation is observed in the overburden, suggesting neighbours have remained in contact and overall packages have been rotated (consistent with bedding). Displacements relative to initial neighbour in the sedimentary sequence appear lower than the pre-kinematic overburden, because the overburden has been deforming for 4.6 Myr, while the sediment has only been deforming for 2.4 Myr. See Spence and Finch (2014) for further details on relative displacement from initial neighbour.

Supplementary Table 1: Summary of geological and mechanical parameter values used in the DEM in this study and a justification for their use.

Supplementary Table 2: Comparison of stratigraphic thinning across different stratigraphy approaching north Pierce diapir (Figure 11). Packages are compared to a specific sedimentation event in a specific model, e.g., the Mid-Miocene – Pleistocene is perhaps analogous to the final sedimentary sequence (S3) in the increasing sedimentation model. Percentage thinning and normalised thinning rates (%/m) are shown

1531   for U-C (Undeformed to Crest), U-F (Undeformed to Salt Flank) and F-C (Salt Flank to Crest). The
1532   positions of the U, F, and C stratigraphic profiles are spaced at the same distance as in modelled examples
1533   (Figure 5). In layers which do not extend across the entire model no results exist for U-C and F-C. An
1534   average seismic velocity of 2000 m/s was used for approximate depth conversion of the time-migrated
1535   seismic data for thinning rate calculations, we are aware this is an over simplification of seismic velocity,
1536   which varies with depth and lithology however it is suitable to give a comparison to modelled values.
1537   Colours subdivide different stratigraphy, and relate to Figure 11.
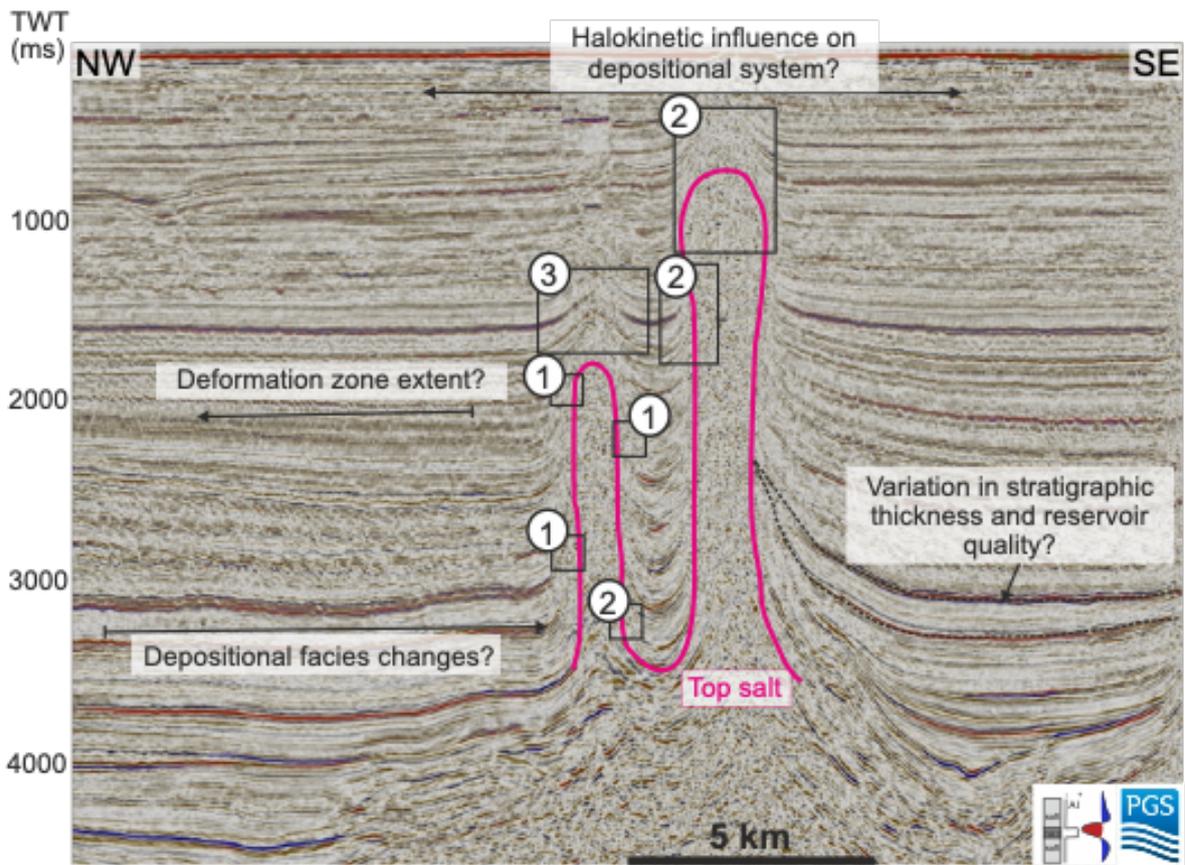
Figure 1: Seismic reflection cross-section from the Pierce diapirs, Eastern Central Graben, UK North Sea (located in Figure 10) highlighting some of the key questions and uncertainties of subsurface interpretation in salt-influenced basins. Mapping top salt, Visualizing pinch-outs and onlap geometries in areas of lower-quality data adjacent to the salt (1), confidently mapping top salt (2) and studying the lateral and vertical extent of syn-kinematic deformation and sedimentation (3), are challenging. Seismic data courtesy of PGS (MegaSurvey Plus 3D seismic data).
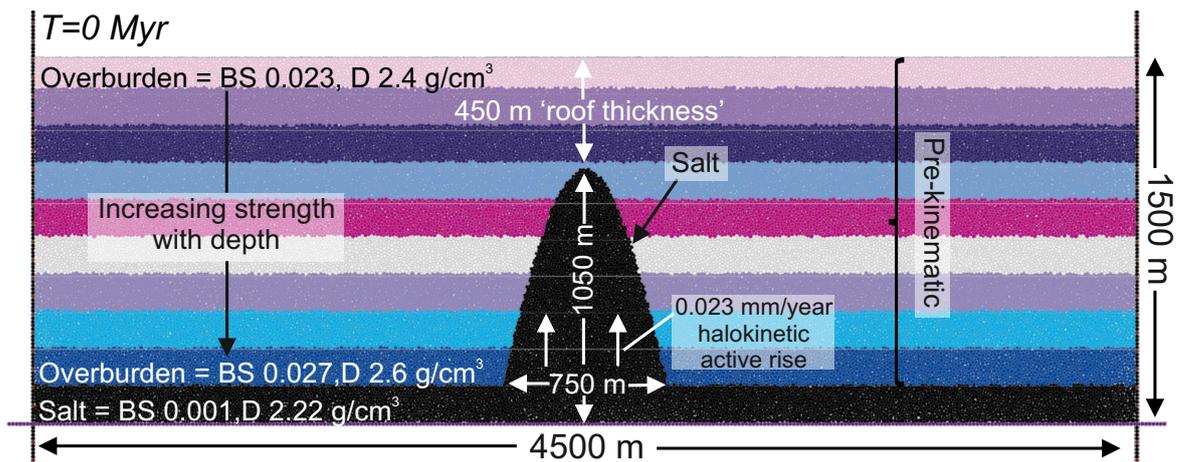
Figure 2: Initial set up of the DEM (T=0) and key parameters. Initial geometry and rise rates taken from North Pierce (Davison et al. 2000). See text for discussion and Table 1 and Supplementary Table 1 for further details.
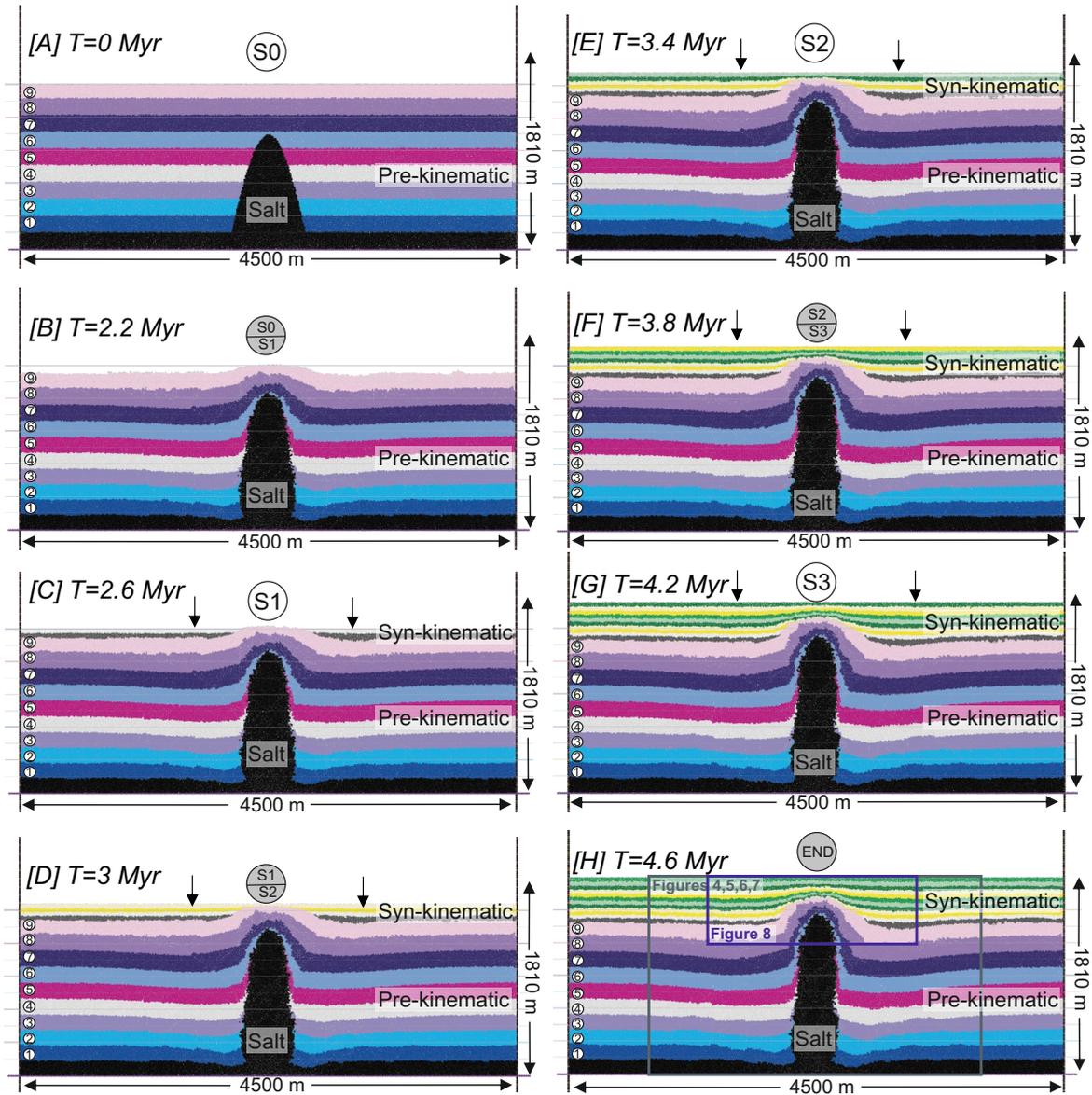
Figure 3: Model 3, the intermediate aggradation experiment with a sedimentation rate of 0.3mm/year. Diapir growth rate is continuous and constant throughout (0.023 mm/year). Outputs begin at the start of the experiment during a stage of no sedimentation (T=0 Myr, circle S0 (sedimentation stage 0)). At T=2.2 Myr (B) the diapir has generated surface topography and sedimentation begins to be added (S0/S1). Subsequent outputs are generated at 0.4 Myr intervals across the three equal stages of sedimentation (open circles S1-S3) from 2.2 to 4.6 Myr. Active sediment input is constant in stages 1-3 and the boundaries are highlighted by the grey circle (e.g. T=3 Myr represents the end of sedimentation 1 and the beginning of sedimentation event 2). The final output is at 4.6 Myr (H), following the final sedimentation from S3. Due to minimal deformation in the outer section of all models, for clarity, subsequent figures will focus on only the central 3000 m around the diapir (shown by the grey outline, T=4.6 Myr). Subsequent figures are shown at T=4.6 (H).
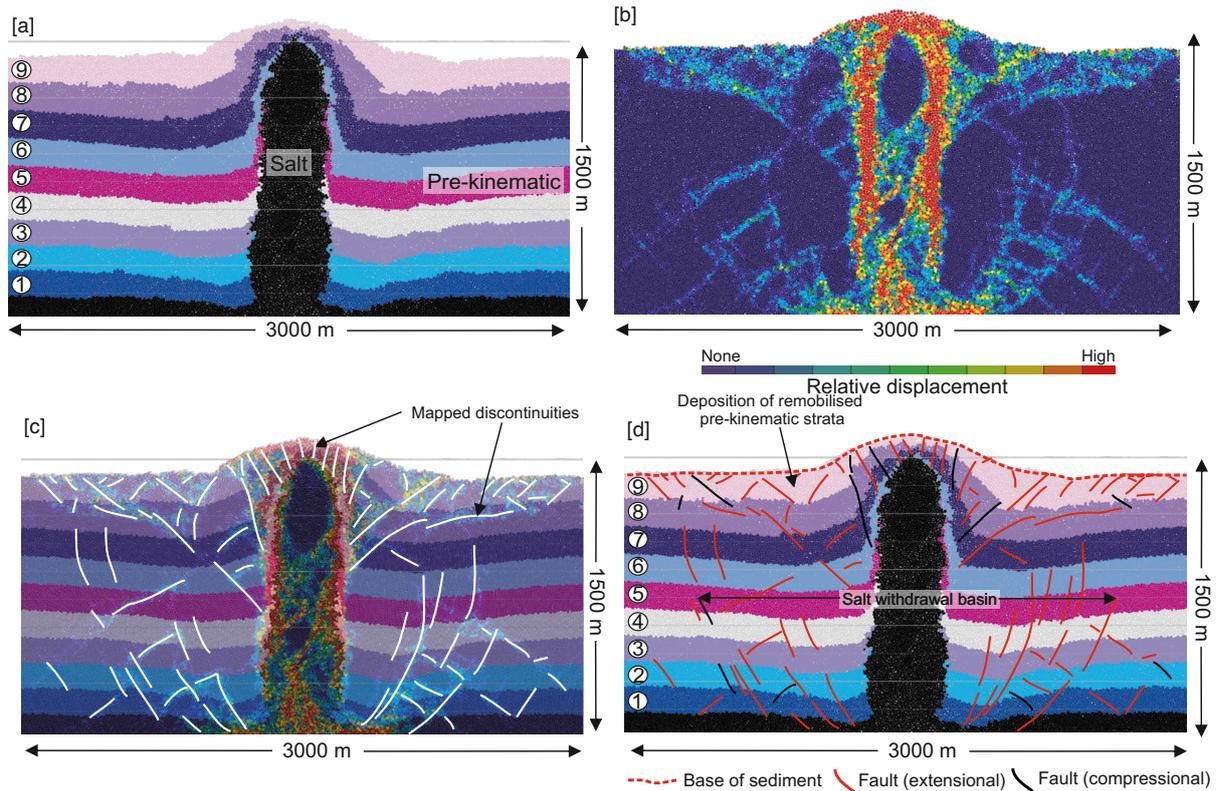
2

Figure 4: M1, the model with no sedimentary fill, after 4.6 Myr. Displayed outputs represent the central 3000 m of the original output model (located in Figure 3). A) Model output, uninterpreted, pre-kinematic layers are numbered for clarity, and ease of subsequent discussion. B) Static image of element displacement relative to initial neighbours for M1 after 4.6 Myr (provided for M1-M6 in Supplementary Figure 2). Cold colours represent elements that are in contact with their original neighbour and hot colours indicate high displacement. This is used to show discontinuities and is a proxy for fault location. C) (A) overlain with (B), interpreted with locations of discontinuities, highlighting the methodology used to interpret structures for all models. Note how the majority of the internal salt has remained in connection with its original neighbouring element and the radial faults are associated with salt withdrawal. The maximum displacement shows the relative movement of the salt, the neighbouring overburden, and the high mobility of layers in close proximity to the diapir. Note the lack of deformation outside of the salt withdrawal basin. D) Interpreted static DEM for 4.6 Myr, faults are taken from discontinuities in (C), and overlain onto (A).

3
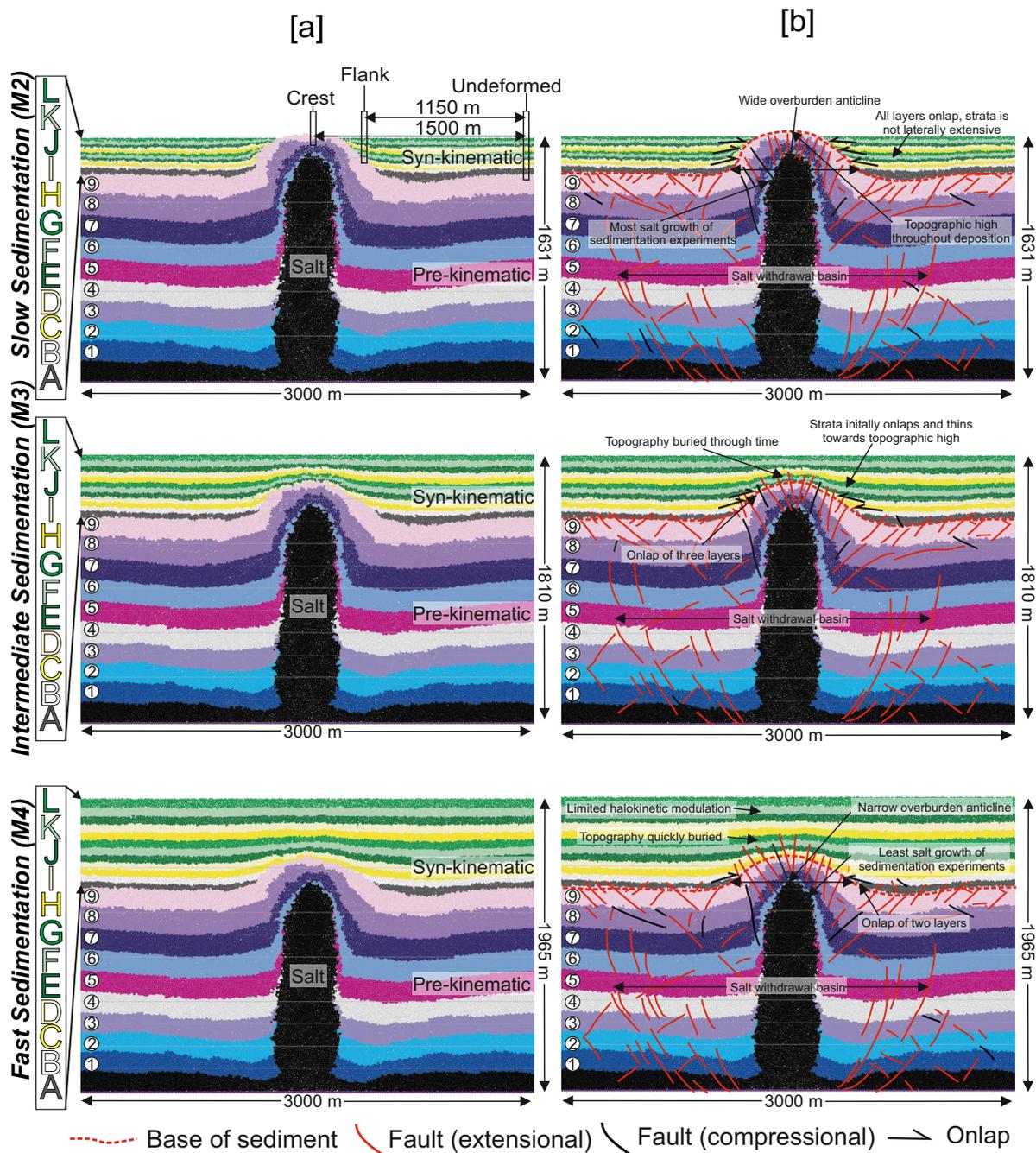
Figure 5: Uninterpreted and interpreted static images of the DEM for M2-M4 (constant sedimentation rate models) after 4.6 Myr. A) Uninterpreted models, sedimentary layers are coloured and assigned letters for easier discussion. Crest, flank and undeformed labels for M2 reflect the locations of thickness measurements taken to calculate thinning rates, this location is consistent across all models (M2-M6), B) Interpreted versions of each model. Fault interpretations use displacement from initial neighbour methodology (see Figure 4, Supplementary Figure 2). Interpretations highlight the width of salt withdrawal basins (deformed zones), height of diapirs, fault and fracture distributions, stratigraphic thickness variation and variable halokinetic influence vertically and laterally, discussed in text.

4

Figure 6: Uninterpreted and interpreted static image of the DEM with variable sedimentation (M5 and M6) after 4.6 Myr. Displayed output represent the central 3000 m of the original output model (located in Figure 3). A) Uninterpreted models, sedimentary layers are coloured and assigned letters for easier discussion. Crest, flank and undeformed zones mentioned in text reflect those shown in M2 (Figure 5). B) Interpreted versions of each model. Fault interpretations use displacement from initial neighbour methodology (see Figure 4, Supplementary Figure 2). Interpretations highlight the width of salt withdrawal basins (deformed zones), height of diapirs, fault and fracture distributions, stratigraphic thickness variation and variable halokinetic influence vertically and laterally discussed in text.

Figure 7: Diagrammatic comparison of static 4.6 Myr DEM across all models. Comparison between top salt (solid line), top of the overburden (dashed line) and top of the sediment (dotted line). The top of sediment is not shown in M1 as none is added. All cases have a similar top salt outside the deformed zone (shown by a bold black line). Diapir growth and width of the pre-kinematic overburden anticline are reduced in M2-M6 when compared to M1 due to overpressure caused by sedimentation (see Table 2). Grey lines indicate 200 metre divisions vertically. Located on Figure 3.
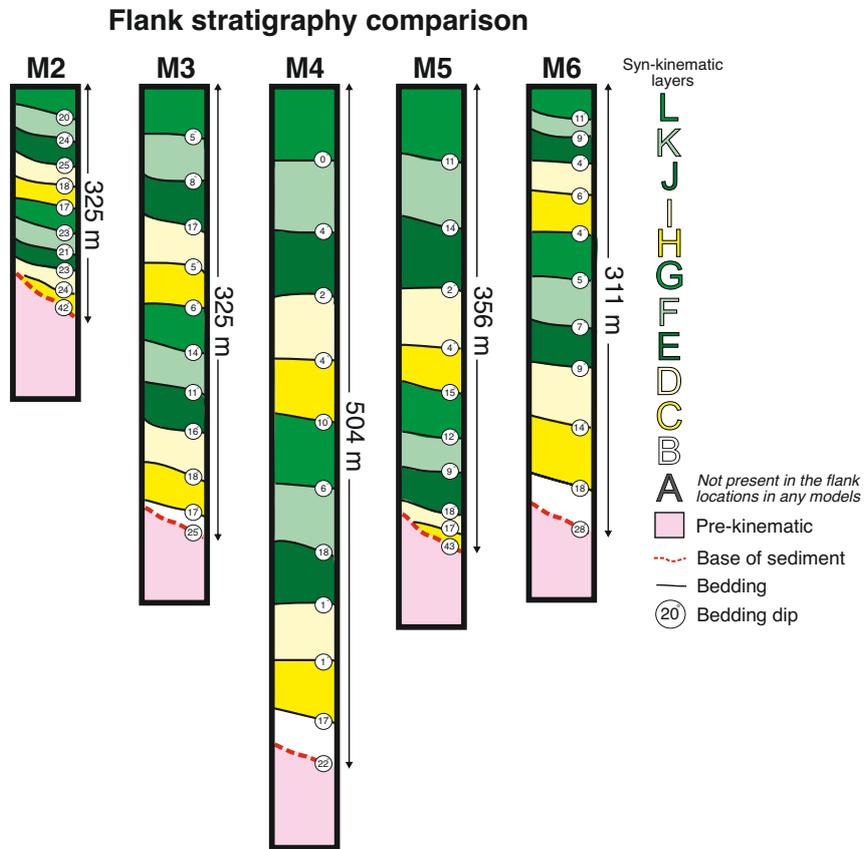
Figure 8: Comparison of stratigraphic thickness for flank location across all models with sediment (M2-M6) see Figure 5 for location of flank profile. Thicknesses and bedding orientation highlighted. The flank location shows the most deformation in all

models, greatest modulation is observed in M2, and least modulation is observed in M4, generally modulation decreases upwards, however in M6 bed dip increases in layers K and L. Models give an indication of the sedimentary thickness and bed rotation expected adjacent to salt diapirs under different sedimentation rates, aiding prediction of sediment thickness and bed rotation in the subsurface salt sediment interface.

A) Slow Sedimentation (M2)

Exposed pre-kinematic overburden

600 m

2000 m

Syn-kinematic

Pre-kinematic
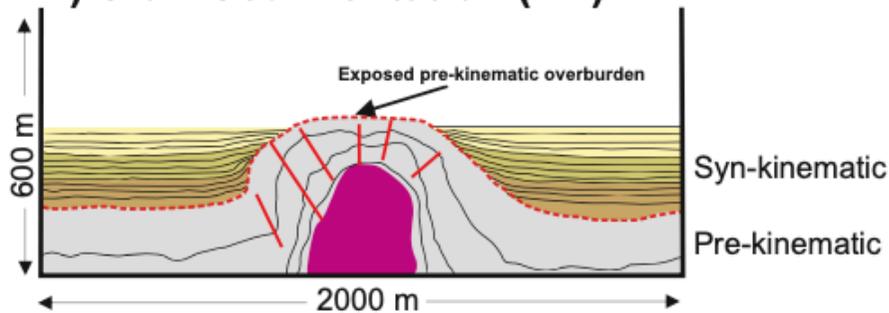
- Non-extensive layers
- Diapir forms topographic high
- Thinning towards the diapir
- Onlap onto the overburden
- Wide overburden anticline
- Long-lived salt influence

B) Intermediate Sedimentation (M3)

600 m

Thinning of layers towards high

2000 m

Syn-kinematic

Pre-kinematic
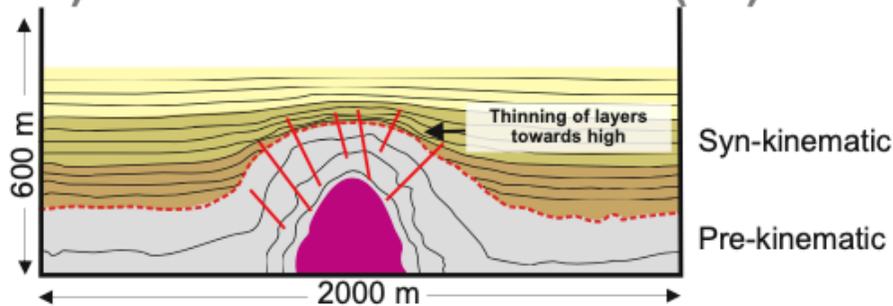
- Initially non-extensive layers
- Early overburden onlap
- Layers thin towards diapir
- Percentage thinning decreases upwards
- Halokinetic influence decreases upwards

C) Fast Sedimentation (M4)

Minimal salt modulation in upper stratigraphy

600 m

Diapir growth inhibited

2000 m

Syn-kinematic

Pre-kinematic
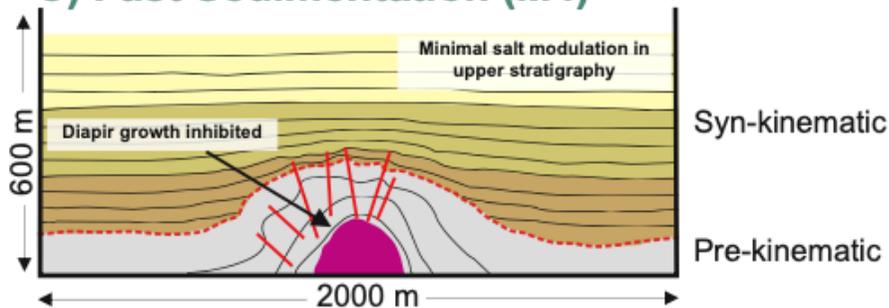
- Laterally extensive layers
- Diapir quickly buried
- Initial onlap, then thinning
- Halokinetic influence decreases upwards
- Narrow overburden anticline
- Reduced salt growth

D) Increasing Sedimentation (M5)
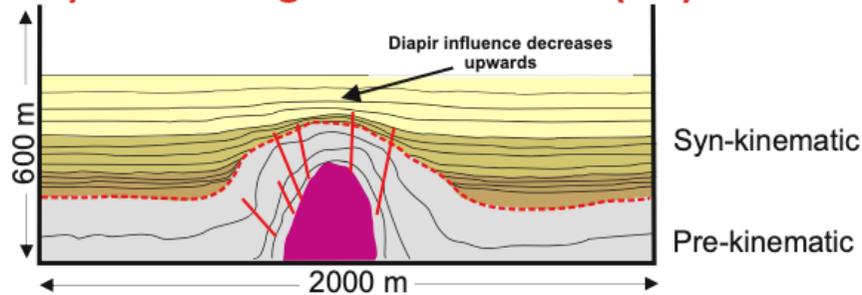
- Sedimentation rate rises to simulate progradation
- Initially non-extensive onlapping layers
- Then thinning of layers
- Halokinetic influence decreases upwards
- Top layer not deformed

Diapir influence decreases upwards

600 m

2000 m

Syn-kinematic

Pre-kinematic

E) Decreasing Sedimentation (M6)
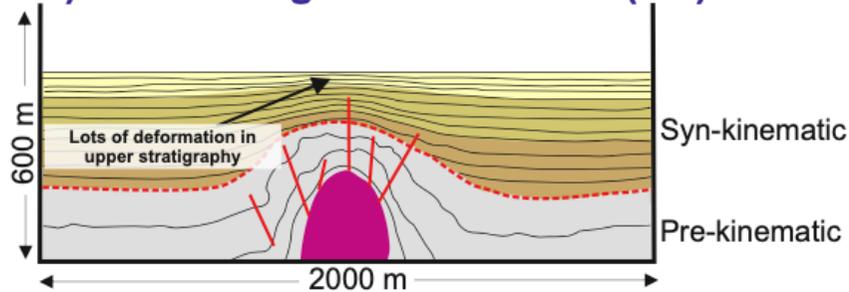
- Sedimentation rate decreases to simulate retrogradation
- First layer non-extensive
- Thinning rate increases upwards
- Halokinetic influence increases upwards
- Top layers thin rapidly towards high
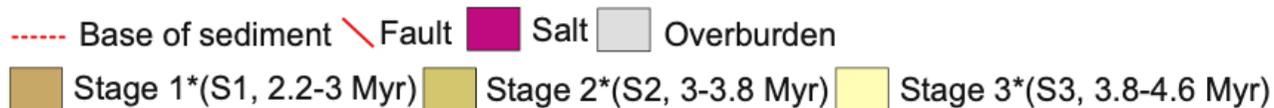
Lots of deformation in upper stratigraphy

600 m

2000 m

Syn-kinematic

Pre-kinematic

----- Base of sediment ＼ Fault ▉ Salt ☐ Overburden

▉ Stage 1*(S1, 2.2-3 Myr) ▉ Stage 2*(S2, 3-3.8 Myr) ☐ Stage 3*(S3, 3.8-4.6 Myr)

Figure 9: Schematic interpretations of M2-M6, focussed on the central 2000 m of the syn-kinematic stratigraphy, located in Figure 3. Bullet points highlight the key observations specific to each modelled scenario. Crestal faults are simplified from Figures 4 -6 to better visualise syn-kinematic stratigraphy.
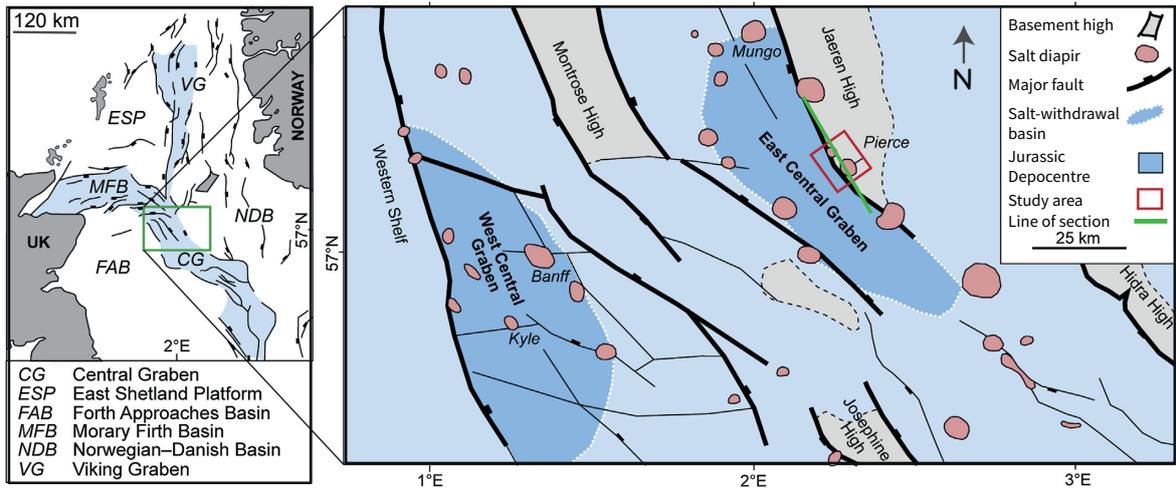
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A Slow Sedimentation (M2)

600 m

2000 m

Syn-kinematic

Pre-kinematic

B Intermediate Sedimentation (M3)

600 m

2000 m

Syn-kinematic

Pre-kinematic

C

TWT (ms)

NW

Undeformed
Flank
Crest

SE

Seabed

1000

S3

Pleistocene

Pliocene
Mid-Miocene

S3

2000

S2

S2
S1

Oligocene

Eocene

3000

S1

S3
S2
S1

Tay
Top Lista
Top
Cretaceous
Base Chalk
Base
Cretaceous
Jurassic

Top Salt

4000

Base Salt

5 km

D Increasing Sedimentation (M5)

600 m

2000 m

Syn-kinematic

Pre-kinematic

Key for all models

- - - - Base of sediment
— Fault
■ Salt
□ Overburden

Stage 1*(S1, 2.2-3 Myr)
Stage 2*(S2, 3-3.8 Myr)
Stage 3*(S3, 3.8-4.6 Myr)

Text and outline coloured by model

Figure 11: Subsurface example of model application comparing some of the modelled results to stratigraphy from the Pierce Field, Eastern Central Graben, UK North Sea. A,B & D) Models 2, 3 and 5 which are analogous to different parts of the stratigraphy around the Pierce diapirs. C) Interpreted time-migrated seismic profile across the Pierce diapirs. S1, S2, S3 highlight the likely sequences for comparison to models. The colour of the text represents which model those packages could represent (e.g. red represents M5). The location of the undeformed, flank and crest stratigraphic locations used for thinning rate calculations are shown for north Pierce, and are the same spacing as those used for model calculations (Figure 5; Table 3). Seismic data courtesy of PGS (MegaSurvey Plus 3D seismic data) from Cumberpatch et al. 2021b.
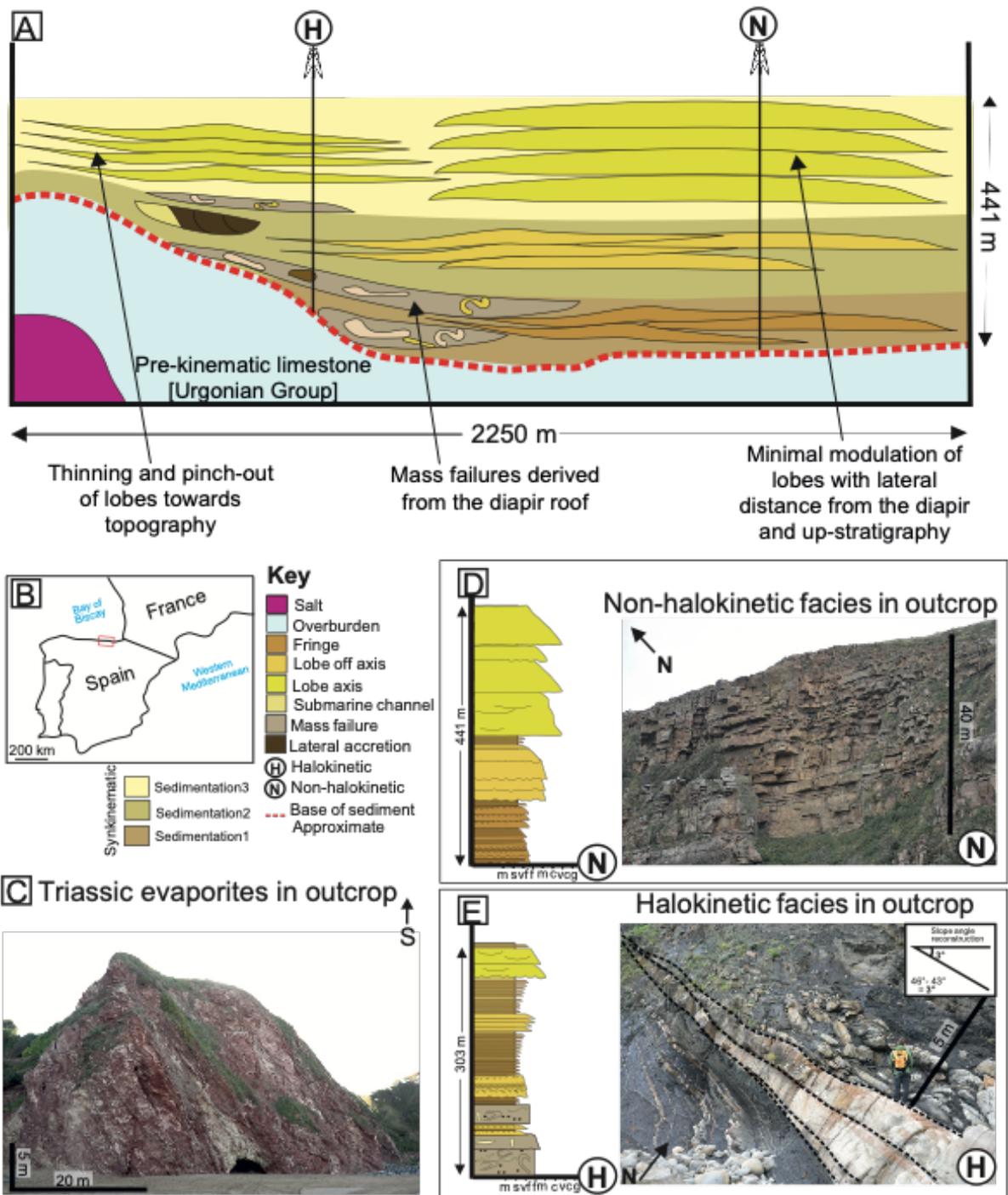
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13

Zone Name	Bedding Orientation	Reservoir Implications	Trap Implications
Minimal Modulation (1)	Limited deformation, horizontal beds	Good quality, thickest sandstones	Stratigraphic trap (e.g. lobe in mudstone)
Halokinetic (2)	Broadly horizontal	Only minimal thinning occured	Purely stratigraphic trap required
Onlap (3)	Rotated towards salt diapir	Reservoirs are thinned, lower net to gross	Onlap stratigraphic traps
Salt Flank (4)	Significantly rotated towards salt diapir	Greatest reservoir thinning, muddier sands	Stratigraphic-structural and onlap trap, column height high risk
Salt Crest (5)	Almost horizontal	Reservoir layers are thinned or missing	Subtle folding over crest could create differential compaction traps

Figure 13: Halokinetic zonation scheme shown for M3. The model is divided into 5 zones, 4 of which experience some form of halokinetic influence. Minimal deformation zones 1 and all stratigraphy outside of it show virtually no modulation by salt diapirism. Halokinetic influence increases towards the diapir-cored high, and changes from thinning in the 'halokinetic' zone to onlap and abrupt pinch out in the 'onlap' and 'salt flank' zones. Thinned and reduced stratigraphy are observed over the diapirs crest. The table highlights that while stratigraphic trap quality may be greater closer to the diapir, reservoir thickness and quality are likely higher further from the diapir, showing a 'trade-off' exists for subsurface energy exploration and production. Similar zonation is possible for all models.

Model	Sedimentation Type	Sedimentation Interval		
		S1 (2.2-3.0 Myr) [22000 – 30000 time steps]	**S2** (3.0-3.8 Myr) [30000 – 38000 time steps]	**S3** (3.8-4.6 Myr) [3800 – 46000 time steps]
		Sedimentation Rate (mm/year)		
M1	None	0.0	0.0	0.0
M2	Slow consistent sedimentation (Aggradation)	0.15	0.15	0.15
M3	Intermediate consistent sedimentation (Aggradation)	0.30	0.30	0.30
M4	Fast consistent sedimentation (Aggradation)	0.45	0.45	0.45
M5	Increasing sedimentation (Progradational)	0.15	0.30	0.45
M6	Decreasing sedimentation (Retrogradational)	0.45	0.30	0.15

Table 1: Details of the different sedimentation patterns and rates used in the six experiments. S1-3 refer to sedimentation intervals.

Model	Diapir growth (upward movement) (m)	Percentage reduction compared to base case (M1)	Pre-kinematic folded roof width (m)	Percentage reduction compared to base case (M1)	Syn-kinematic folded roof width (m)	Ratio of syn- : pre-kinematic folded roof (m)
M1, No sedimentation	425	n/a	961	n/a	n/a	n/a
M2, Slow sedimentation rate	393	8	770	20	n/a	n/a
M3, Intermediate sedimentation rate	363	15	760	21	839	1.1
M4, Fast sedimentation rate	297	30	734	24	890	1.21
M5, Increasing sedimentation rate	368	14	750	22	723	0.96
M6, Decreasing sedimentation rate	346	19	753	22	872	1.16

Table 2: Comparison of diapir growth and overburden anticline width across all six experiments at T=4.6 Myr. Measurements are taken from model simulations (Figure 4, 5 and 6). Initial diapir height (1050 m at T=0 Myr) is subtracted from final diapir height to give diapir growth. M2-6 are compared to M1 to show the percentage reduction that occurs when different amounts of sediment are added. Pre-kinematic anticline measurements are taken from the stratal termination of layer A, in M1 they are taken from the top of the overburden at the location of the greatest change in dip. For M2-M6 the width of the overburden is compared to M1 to show the percentage reduction in overburden anticline width that occurs when different amounts of sediment are added. Syn-kinematic folded roof measurement are taken from location of greatest dip change in layer F (Supplementary Figure 1). Value only calculated where layers extend across the entire structure (M3-M6). Ratio of syn- to pre-kinematic folded roof thickness is calculated.

Model	U-C (%)	U-C (%/m)	U-F (%)	U-F (%/m)	F-C (%)	F-C (%/m)	U-F:F-C
Slow (M2)	-	-	34.4	0.029	-	-	-
Intermediate (M3)	54.5	0.037	19.2	0.017	35.3	0.1	1:6.1
Layers B & C [~S1]	-	-	26.4	0.023	-	-	-
Layers D-H [~S2]	60.1	0.040	14.5	0.011	45.6	0.13	1:10.3
Layers I-L [S3]	23.2	0.015	6.3	0.005	16.9	0.048	1:8.8
Fast (M4)	32.5	0.022	11.6	0.010	20.9	0.060	1:5.9
Layer B [Early S1]	-	-	47.0	0.041	-	-	-
Layers C & D [Late S1]	38.4	0.026	12.9	0.011	25.5	0.073	1:6.5
Layers E-H [S2]	29.7	0.019	8.6	0.007	21.1	0.06	1:8
Layers I-L [S3]	2.35	0.0016	1.04	0.0009	1.3	0.037	1:4.1
Increasing (M5)	58.3	0.039	19.3	0.017	38.9	0.11	1:6.6
Layers B-D [S1]	-	-	61.1	0.053	-	-	-
Layer E [Early S2]	-	-	24.8	0.0022	-	-	-
Layers F-H [Late S2]	60.6	0.040	15.9	0.014	44.8	0.128	1:9.2
Layers I-L [S3]	21.5	0.014	6.7	0.006	14.7	0.042	1:7.2
Decreasing (M6)	52.4	0.035	18.6	0.016	33.9	0.097	1:5.9
Layer B [Early S1]	-	-	37.4	0.017	-	-	-
Layers C&D [Late S1]	39.6	0.026	10.3	0.009	29.3	0.084	1:9.3
Layers E-H [S2]	42.3	0.028	11.4	0.009	30.8	0.088	1:8.8
Layers I-L [S3]	43.9	0.029	19.5	0.017	24.5	0.069	1:4.1

Table 3: Comparison of stratigraphic thinning across M2-M6. Bold headings indicate each model and show the total thinning for all stratigraphy, beneath each heading this is divided by sedimentary package. Layer A is excluded as it fills to base level in all models. Layers are subdivided based on observational differences (e.g. lateral extent across the model) and are broadly grouped into their sedimentation stage (S1, S2, S3). Percentage thinning and normalised thinning rates (%/m) are shown for U-C (Undeformed to Crest, i.e., the total model), U-F (Undeformed to Salt Flank) and F-C (Salt Flank to Crest). In layers which do not extend across the entire model no results exist for U-C and F-C. The final column displays the ratio of the normalised thinning rate between the Undeformed to Salt Flank and the Salt Flank to Crest sections, to show how much more thinning is observed adjacent to the diapir. Locations of undeformed, flank and crest measurements are shown on Figure 5, and are the same for all models to ensure direct comparison.