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# Evolution of fluvial systems in salt-walled mini-basins: a review and new insights

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## *Abstract*

The preserved sedimentary expression of fluvial successions accumulated in salt-walled mini-basins records the complex history of basin subsidence, the style of sediment supply, and the pattern of sediment distribution in response to a range of fluvial processes throughout the evolution of such basins. Temporal and spatial variations in the rate of basin subsidence govern the generation of accommodation space, whereas the rate and style of sediment supply govern how available accommodation is filled; together these parameters act as principal controls that dictate the gross-scale pattern of fluvial sedimentation. Additional factors that influence fluvial stratigraphic architecture in salt-walled mini-basins are: (i) the trend and form of inherited basement lineations and faults that control the geometry, orientation and spacing of salt walls that develop in response to halokinesis; (ii) salt thickness and composition that dictate both the maximum potential basin-fill thickness within a developing mini-basin and the rate of evacuation (migration) of salt from beneath evolving mini-basins, leading to the growth of confining salt walls, uplift of which may generate surface topographic expression that influences fluvial drainage patterns; (iii) climate that dictates fluvial style and the processes by which sediment is distributed; and (iv) the inherited direction of drainage relative to the trend of elongate salt walls and locus of sediment supply that dictates how sediments are distributed both within a single mini-basin and between adjacent basins.

Examples of fluvial sedimentary architectures preserved in salt-walled mini-basins from a number of geographic regions are used to illustrate and document the primary controls that influence patterns of fluvial sediment accumulation. The distribution of fluvial architectural elements preserved within mini-basins follows a predictable pattern, both within individual basin depocentres and between adjoining basins: drainage pathways preferentially migrate to topographic lows within basins, such as developing rim-synclines, and away from topographic highs, such as uplifting salt walls or developing turtle-back structures.

This paper demonstrates a range of fluvial-halokinetic interactions through consideration of a series of case studies, which demonstrate current understanding of fluvial response to salt-walled mini-basin evolution and which highlight gaps in current understanding.

Keywords: fluvial, salt wall, mini-basin, architecture; stratigraphy; accommodation; subsidence

## *1. Introduction*

Globally, there exist in excess of 120 provinces in which evaporite basins are known to have been influenced by salt deformation (Hudec and Jackson, 2007; Fig. 1). Numerous studies have been previously conducted to demonstrate how various sedimentary environments are influenced by coeval halokinesis that results in high rates of basin subsidence (e.g., Prather, 1998), diversion of sediment transport pathways by uplifting topography (e.g., Kneller and McCaffrey, 1995; Banham and Mountney, 2013a), and reworking of uplifted sediments or diapir-derived detritus (e.g., Lawton and Buck, 2006). Numerous studies show how the effects of these phenomena are expressed in the preserved stratigraphic record: in deep-water environments, turbidity currents can be deflected, diverted or reflected by uplifting salt topography resulting in a complex arrangement of turbidite deposits (Kelling et al., 1979; Kneller and McCaffery, 1995; Byrd et al., 2004; Kane et al, 2012); in shallow-marine environments, enhanced rates of subsidence can locally increase sediment accumulation rates (Dyson, 2004; Kernen et al. 2012); in aeolian environments, surface topography arising from salt-wall growth can encourage dune-field construction, accumulation and preservation by shielding such environments from reworking by fluvial processes (Venus, 2013). Of these and other studies, only a modest number have attempted to document and account for the style of accumulation of fluvial successions in salt-walled mini-basins and show how fluvial systems can be diverted by salt-wall-generated topography. Despite having hitherto been the attention of only relatively few studies, understanding the detailed sedimentology and stratigaphy of fluvial successions preserved in salt-walled mini-basins is important since such successions act as economically important hydrocarbon reservoirs in several salt-basin provinces globally (Smith et al., 1993; Barde et al., 2002; Newell et al., 2012).

The aim of this paper is to review the current state of literature regarding controls on the style of accumulation of fluvial successions in salt-walled mini-basins and to highlight gaps in current understanding. Specific objectives are as follows: (i) to establish a standard set of terminology for the description of various attributes associated with the spatial and temporal evolution of salt-walled mini-basins; (ii) to highlight the numerous ways in which halokinetic and sedimentary processes can interact; (iii) to illustrate how these different styles of interaction are known to be expressed through examination of a series of reviewed case studies; (iv) to present a series of summary tectono-stratigraphic models with which to relate preserved fluvial stratigraphic architecture present in mini-basins to the principal halokinetic and sedimentary controls; (v) to show how such models can be used as predictive tools; and (vi) to discuss potential approaches to future research that will address issues that currently remain unresolved in this field of research.

This work is of broad appeal for the following reasons: (i) the terminology describing the attributes and style of infill of salt-walled mini-basins are currently poorly defined and this study provides clarification and discussion through development of a generic classification framework; (ii) this work identifies and discusses a series of controls that operate to determine the style of evolution of salt-walled mini-basins and the manner by which these basins become filled by fluvial successions; and (iii) this work distils our current understanding into a series of generic models that describe the influence of key controls on fluvial sedimentation for a variety of types of basin fill.

## *2. Terminology*

The terminology required for the description of basin subsidence, gross-style of basin fill and basin-fill state at any given time during the evolution of a series of salt-walled mini-basins is inherently complex because many dependent and independent variables are known to interact during the evolution of such systems. To resolve this issue, terminology describing the primary variables that govern mini-basin evolution and their fill states are defined here in an attempt to standardize descriptions of basin attributes (Fig. 2).

**Basin-fill thickness (T)** describes the current total thickness of accumulated sediment within a subsiding mini-basin. This thickness may vary across a single basin in cases where differential subsidence has generated variable accommodation; for example, a rim syncline structure (R) will locally increase accommodation, whereas accommodation will be less above a turtle-back structure (Tb).

**Maximum basin-fill thickness (M)** describes the maximum potential thickness of fill that can be accommodated by continued subsidence and accumulation within a mini-basin, and this is governed by both basement geometry and by the original thickness of salt present at the location of mini-basin formation. Where a mini-basin grounds on pre-salt basement strata, further subsidence is no longer possible and T = M, if the effects of additional sediment compaction are ignored.

**Remaining basin-fill potential (P)** describes the remaining thickness of salt beneath a subsiding mini-basin at a given point, and the maximum remaining distance the mini-basin can subside before it grounds on the sub-salt basement. Where a basin or part thereof grounds on the pre-salt basement, a salt weld is formed and remnant basin-fill potential (potential accommodation space) for that location is theoretically zero.

**Fill inheritance (F)** describes the pre-existing basin-fill thickness at the onset of a new sequence of sediment delivery, where the inherited fill state is described in relation to the onset of accumulation of a new stratigraphic sequence. Differential rates of subsidence within a single mini-basin can lead to spatial variation in inherited basin-fill thickness at the onset of accumulation of a later stratigraphic sequence. Such a situation might arise in response to the early grounding of one side of a mini-basin while the other side remains actively subsiding and able to accumulate additional strata. This can result in the development of a so-called "heel-toe" sediment-fill geometry (Kluth and DuChene, 2009), a common style of architectural expression. The sum of all inherited basin-fill is equal to basin-fill thickness (T).

**Available unfilled accommodation space (U)** refers to the vertical thickness of accommodation within a mini-basin that remains unfilled by sediment at a given point and for a specified time period, but which could potentially be filled with sediment without additional subsidence occurring. This can be a negative value if the basin-fill rises locally or temporarily above the regional base level.

**Basin-fill style (S)** describes the nature of the basin-fill and the distribution of fluvial elements in general, qualitative terms; for example, whether the basin-fill is relatively sand-prone or sand-poor. The distribution of fluvial elements may be heterogeneous at the scale of a mini-basin, giving rise to variations in the style of accumulated strata, potentially in orientations parallel or perpendicular to the trend of elongate basins, or vertically within the overall fill of a mini-basin. For example, the arrangement of stratal packages may exhibit heterogeneity such that groups of channel-fill elements might be clustered at certain stratigraphic intervals or at only one side of a mini-basin.

**Salt-wall height (W)** describes the relief of the salt-wall (or its directly overlying cover sediment) relative to that of the sediment fill-level in the adjoining mini-basin(s). When a salt wall (or its cover sediment) rises above the height of the surrounding basin plain to generate a topographic expression, it will be prone to erosion and reworking, potentially acting as a source for the generation of clastic detritus derived from reworked cover sediment, and diapir-derived detritus reworked from the salt wall itself. Such detritus may be reworked into the surrounding accumulating stratigraphy as part of the basin fill and recorded as lithofacies characterized by lithic clasts of local intraformational origin (e.g., mudstone rip-ups) or by clasts of reworked evaporite material (e.g., gypsum).

**Basin subsidence rate (R)** refers specifically to the rate at which the floor of the mini-basin subsides into the underlying salt. This value may vary in orientations both parallel and transverse to the axis of elongate mini-basins, as well as temporally.

**Salt-wall uplift rate (U)** refers to the rate at which a salt wall (or its directly overlying cover sediment) is uplifted above the sediment fill-level of the adjoining mini-basin. This rate of uplift may be modified by dissolution of the salt in the subsurface, effectively reducing the rate of uplift.

**Salt-weld formation or basin grounding** refers to the time that basin subsidence effectively ceases because the remaining thickness of salt beneath a subsiding mini-basin is insufficient to allow further flow because the remaining basin-fill potential (P) is effectively zero (cf., Hudec and Jackson, 2009) and a salt weld forms. The ability of salt to deform and flow is dependent on the thickness of salt, with shear rates in the salt tending to reduce with decreasing thickness of salt: the conditions that dictate the timing of salt-weld formation vary and are mainly dependent on the composition of the salt layers (Hudec and Jackson, 2009).

## *3. Controls on the style of stratigraphic in-fill of salt-walled mini-basins*

### 3.1 Initiation of salt-walled mini-basins

The initiation of mini-basin subsidence requires a number of prerequisites: (i) the presence of a salt layer (or layers) of sufficient thickness to allow halokinesis to occur (Trusheim, 1960; Hudec *et al*, 2009); and (ii) a mechanism to initiate halokinesis (Fig. 3), such as extension (Hodgson et al., 1992), compression (Jackson and Talbot, 1986; Brun and Fort, 2003), differential loading (Ings and Beaumont, 2010), or buoyancy (Trusheim, 1960). Where salt thickness is sufficient to allow the development of salt-walled mini-basins, the presence of pre-salt basement structures, including their geometry, trend, spacing and along-strike continuity commonly exert a significant control on the location and style of salt-wall growth (Doelling, 1988; Barde et al., 2002a; Trudgill, 2011). Sites of initiation of salt-wall growth and the orientation of such salt walls have been related to the distribution and orientation of various types of basement feature, including horst-and-graben structures or relict topography, which generate variations in the thickness of salt, of which triggers the onset of halokinesis (Smith et al, 1993; Hodgson et al, 1992). Salt walls typically form above horst and graben structures, although a differential thickness of salt may not itself be the sole factor responsible for the initiation of the growth of salt walls. Additionally, differential loading may exert a control on the spacing of salt walls, as a function of salt viscosity, salt thickness, and overburden density (Ings and Beaumont, 2010). Mini-basin size varies, in part as a function of initiation mechanism, with individual basins typically being 8 to 15 km wide, whereas intervening salt walls typically have widths of 1 to 2 km (Barde et al., 2002a; Goldsmith et al., 2003;Trudgill, 2011; Banham and Mountney, 2013a).

The ongoing growth of salt-walled mini-basins is maintained and driven by a buoyancy imbalance (Rayleigh-Taylor instability), where the overlying sediment has a greater density than that of the underlying salt (Fig. 3; Hudec et al, 2009; Ings and Beaumont, 2010). This density-driven process typically requires sediments to have a density of ~2500 kg $m^{-3}$, which equates to a burial depth of ~1000 to 2300 m to generate the mechanical compaction required to achieve this density for most clastic sediments (Jackson and Talbot, 1986; Hudec et al, 2009).

Initiation of salt-wall growth by other mechanisms has also been described by Hudec et al (2009), including: (i) lateral shortening of the salt layer due to the application of compressive stress, thereby creating a bathymetric high where salt is forced up and a bathymetric low in the top surface of the adjacent salt; (ii) thinning of the salt layer due to extensional stress, whereby stretching of the salt layer causes it to sag, forming a bathymetric low (Fig. 3); or (iii) flow of salt via creep down-dip under the influence of gravity, thereby creating a bathymetric low at the head of the original salt body; (iv) sediment loading, whereby overlying strata of variable thickness generate a significant differential load at a point on the underlying salt layer (Fig. 3); (v) sub-salt deformation, such as the creation of a roll-over basin by extension or folding by compression. Each of these mechanisms relies on the generation of a bathymetric low in the salt to allow for the accumulation of sediment, progressive accumulation of which, in turn, generates additional loading and enables buoyancy-driven withdrawal and lateral salt migration at depth, thereby leading to additional subsidence at the site of loading.

An alternative mechanism for the initiation of mini-basin generation is the viscous pressure ridge model proposed by Ings and Beaumont (2010), in which flow of overburden and underlying salt – driven by, for example, collapse of a continental margin or progradation of a delta or alluvial mega-fan – can result in the formation of a pressure ridge due to differential rates of flow within the underlying salt. The trapping of sediment by the formation of these viscous pressure ridges culminates in the development of a sediment succession that is sufficiently thick to create a Rayleigh-Taylor instability, allowing conventional buoyancy-driven subsidence to take over.

Once a sufficient density contrast threshold has been attained, whereby compaction of the overlying sediment has resulted in a mean sediment density that is greater than that of the underlying salt, load-driven displacement of the salt from beneath the incipient mini-basins will commence as salt flows into neighbouring growing salt walls (Ings and Beaumont, 2010).

The evolution of salt-walled mini-basins (or 'pods') has been described previously by Hodgson et al (1992) (Fig. 4). Initiation of salt-wall growth can be triggered by any one of the aforementioned mechanisms, before sediment loading of the salt eventually takes over as the driving mechanism of basin subsidence and salt wall growth (Hudec et al, 2009; Ings and Beaumont, 2010). Sediment accumulation in these basins continues by the process of down-building (Barton, 1933) until the basin grounds on the sub-salt basement, effectively preventing additional accumulation of sediments within the mini-basin. Later, axial migration or dissolution of salt from an uplifted swell, wall or stock can cause the salt uplift to collapse, thereby allowing secondary mini-basins to form over the crests of the former salt-wall highs (Hodgson et al, 1992). The typical rate of mini basin subsidence

Once initiated, mini-basins can subside at sustained rates of >1 km/Ma for several million years: for example, some Pliocene and Pleistocene examples have fills that are up to 8 km thick (Hudec et al, 2009). In some instances, rates of up to 10 km/Ma have been recorded in the Gulf of Mexico (Prather, 2000).

## 3.2 Parameters controlling subsidence and sedimentation rate

Many parameters are known to influence the style of sediment accumulation in salt-walled mini-basins: some are static (e.g., original salt thickness and composition) in that they do not vary throughout the episode of mini-basin subsidence; others are dynamic variables (e.g., climate and sediment delivery rate) that change over the course of mini-basin subsidence. Understanding these parameters is key to determining the history of subsidence and sedimentation within a salt mini-basin province, and for showing how this may have controlled fluvial drainage pathways, and subsequently how this influenced basin-fill evolution.

### Static parameters

Static parameters are controls that remain constant (i.e., temporally invariable) throughout the evolution of a mini-basin; such parameters typically exert a basin-scale control on system evolution and are normally set prior to the onset of halokinesis.

***Basement geometries.*** The trends of faults in the pre-salt basement, their spacing and geometry, together with the average dip of the pre-salt basement, act to control the spatial pattern of development and temporal sequence of evolution of growing salt walls (Fig. 5). Salt walls tend to develop at a site of change in salt thickness, such as commonly occurs across fault offsets (Doelling, 1988; Smith et al, 1992; Trudgill, 2011). Alternatively changes in salt thickness may occur where salt overlies buried topography, or may result from facies variations within the evaporite-bearing deposition units. Where basement trends are simple and follow a single trend, salt walls tend to evolve as elongate, linear and parallel features (e.g., the Permian Salt Anticline Region of the Paradox Basin, Utah; La Popa Basin, Mexico; parts of the Central Graben in the subsurface of the North Sea). By contrast, in situations where basement features are present that trend in different orientations, more complex pre-salt basement geometries tend to favour the evolution of salt walls arranged in polygonal patterns and with varying continuity (e.g., Pre-Caspian Basin, Kazakhstan) (Fig. 5). Basement dip, which can result in a variable basement depth across a basin, may determine basin-scale regional changes in the thickness of salt that accumulates. This, in turn, determines the maximum potential basin-fill thickness during the later development of salt-walled mini-basins.

***Total thickness of salt.*** The thickness of salt ultimately controls the maximum distance a mini-basin can subside before it grounds on the pre-salt basement. The total thickness of salt can vary across the basin (e.g., Paradox Basin, Trudgill, 2011; Central North Sea, Hodgson et al, 1992, Smith et al, 1993), resulting in adjacent mini-basins grounding at different times during the evolution of a mini-basin province. Mini-basin grounding results in a cessation of generation of further accommodation in that basin and once local available accommodation has been filled, sediment bypass into neighbouring basins will commence leading to a relative increase in sedimentation rate in mini-basins that may formerly have been relatively sediment-starved. Furthermore, thicker successions of salt tend to deform and flow at faster rates than thinner successions, meaning that higher rates of subsidence tend to occur in mini-basin provinces for which evaporite thicknesses are greatest (Hudec and Jackson, 2007).

***Evaporite properties.*** The composition and style of stratification of the evaporate-bearing units undergoing halokinesis exert a control on the shear rate of the salt as it deforms and flows within the subsurface. The presence of clastic or carbonate lithologies within an otherwise evaporite-dominated succession will tend to reduce the flow rate (Hite, 1968; Jackson and Talbot, 1986), thereby directly influencing the rate of subsidence of overlying mini-basins and the rate of uplift of adjacent salt walls. The composition of the evaporites undergoing deformation will also influence the timing of salt-weld formation because the presence of clastic and carbonate lithologies acts to hamper the ability of salt to flow, especially where the thickness of salt is substantially reduced.

**Dynamic parameters**

Dynamic parameters are controls that vary either spatially within or between mini-basins, or temporally over the duration of the evolution of one or a series of mini-basins. These factors can be allogenic or autogenic in origin and can influence the style of sedimentation at a range of scales.

***Geothermal controls.*** Geothermal gradient dictates the viscosity and density of salt (Jackson and Talbot, 1986). An increase in the geothermal gradient will act to reduce salt viscosity, thereby enabling it to flow at a faster rate. Decreasing the density of the salt will reduce the threshold required to allow buoyancy-driven subsidence to occur (Srivastava and Merchant, 1973; Jackson and Talbot, 1986).

***Climate.*** Climate controls the evolution of salt walled mini-basins in several ways. Where meteoric water percolates into subsurface salt layers, “softening” of the salt ensues, leading to increased flow rates, enhanced rates of subsurface dissolution, a reduction in the overall rate of salt-wall uplift, or enhanced rates of mini-basin subsidence (Jackson and Talbot, 1986; Senseny et al, 1992). Climate is also a fundamental control that influences rates of weathering and erosion in fluvial catchments, fluvial discharge regime, style of sediment transport, and fluvial form at downstream sites of sediment deposition. Thus, climate exerts a significant influence on the ensuing style of fluvial sedimentation and generation of preserved sedimentary architecture.

***Sediment delivery direction.*** The orientation of inherited sediment delivery networks relative to the orientation of the trend of evolving salt walls exerts a fundamental control on the style of stratigraphic architecture preserved both within a single mini-basin and between neighbouring mini-basins. In situations where preferred drainage is aligned transverse to the trend of growing salt walls – and especially in cases where salt-wall uplift has been sufficient to generate a surface topographic expression – the style of fill of a series of adjacent mini-basins will tend to be manifest as a systematic proximal-to-distal fining away from the sediment source (Venus, 2013). By contrast, fluvial drainage systems aligned parallel to the trend of salt walls tend to result in basin-fill architectures that can change from sand-prone to sand-poor between neighbouring mini-basins, in situations where topography associated with growing salt walls is effective in confining fluvial fairways to a particular mini-basin, leaving others relatively sediment-starved (Banham and Mountney, 2013 a, b).

***Sediment delivery rate.*** The rate of sediment delivery, which is significantly controlled by external factors such as climate regime and bedrock geology in the catchment area, exerts a direct control on the rate at which accommodation in mini-basins becomes filled; evidence for such control is recorded in the architectural fill-style of the developing mini-basins. Furthermore, the rate of sediment delivery and infilling of accommodation also exerts an indirect control on the generation of new accommodation by driving additional subsidence due to loading that enhances rates of subsurface salt withdrawal from beneath evolving mini-basins. High rates of sediment delivery tend to favour rapid infilling of available accommodation, leading to significant reworking of earlier deposits by fluvial systems that migrate dynamically across alluvial plains and undertake repeated avulsions. Such activity tends to preserve fluvial expressions that are dominated by relatively coarse-grained lower parts of fluvial channel-fill elements, with reworking leading to considerable bypass of detritus farther downstream (Hardgrove et al, 2010). Such conditions favour the accumulation of relatively sand-prone basin-fill styles with the associated preservation of multi-storey channel complexes (Banham and Mountney, 2013a). By contrast, low rates of sediment delivery favour the accumulation and preservation of more complete fluvial depositional cycles arising from the cut, fill and migration of channels and the accumulation of surrounding floodplain elements since accommodation will more likely be available to promote preservation. Such conditions tend to favour the development of relatively sand-poor basin fills in which a greater proportion of argillaceous floodplain sediments are preserved and where channel-belts will tend to be isolated in otherwise overbank-dominated successions (cf. Bristow and Best, 1993; Banham and Mountney, 2013a).

***Dissolution rate.*** Salt dissolution by meteoric waters tends to enhance rates of mini-basin subsidence and retard rates of salt-wall growth. A reduction or even reversal of salt-wall uplift may result in the diminishment or elimination of surface topographic expression, resulting in a reduction in the amount of incision required by a fluvial system to maintain a drainage pathway across an actively uplifting salt wall and potentially eventually leading to the linkage of neighbouring mini-basins and a cessation of basin isolation.

## Feedback cycles

Several types of feedback cycles, which can be either positive or negative in behaviour, operate in subsiding salt-walled mini-basin provinces, and these have been shown to govern the rate and style of sediment accumulation within these basins.

Sediment loading of the basin is one example of a positive feedback cycle. Sediment accumulating in the basin loads the underlying salt, resulting in displacement of the salt into the neighbouring salt walls thereby inducing subsidence of the developing mini-basin into the underlying salt and the generation of additional accommodation in which additional sediment can accumulate. Load-induced compaction of strata in the lower parts of basin fills increases the overall density of the basin fill, thereby further increasing rates of salt displacement and driving further subsidence.

Growth of salt walls such that they generate a surface topographic expression that acts to impede fluvial drainage networks also illustrates an example of a positive feedback loop. Where an uplifting salt wall impedes fluvial drainage, ponding behind the elevated surface topography enhances the potential for sediment accumulation in the mini-basin upstream of the salt wall. Such accumulation increases the net sediment-fill thickness and the additional load on the underlying salt associated with this drives further salt displacement and migration into the growing salt wall, thereby maintaining a surface topographic expression that can act as a barrier to drainage.

The tendency for the rate of subsidence beneath evolving mini-basins to decrease as the time of salt-weld formation approaches and the remaining thickness of salt is reduced demonstrates one example of a negative feedback cycle. As the remaining thickness of salt decreases, so its ability to flow is retarded (Hudec and Jackson, 2007), culminating in the generation of a salt weld (or effective salt weld), at which time subsidence ceases.

## Fluvial Interactions

Surface topography generated by the combination of growth of salt walls with subsidence of adjacent mini-basins exerts a fundamental control on fluvial drainage pathways and therefore also on the resultant accumulated stratigraphic succession. The effect of this control is manifest in a number of different ways (Fig. 6). Preferred or inherited orientation of drainage networks relative to the trend of salt walls acts to determine the type and geometrical arrangement of preserved fluvial elements and their distribution both within and between mini-basins. Topography associated with uplifted salt walls can divert or deflect transverse-draining fluvial systems, or induce localised accumulation of sediment while fluvial systems attempt to incise across uplifted salt-walls. This can ultimately lead to drainage capture or diversion and the development of antecedent drainage networks. This can reduce sediment input into neighbouring basins, which inturn can lead to the formation of relatively sand-poor basins adjacent to relatively sand-prone basins. Where drainage pathways cross salt-wall-generated topography, the potential rate of fluvial incision must be greater than the rate of salt-wall uplift for the fluvial course to be maintained.

Active channels draining parallel to or across salt-wall-generated topography can migrate and encroach on to and rework sediment derived from the flanks of salt walls leading to the accumulation of beds composed of locally reworked intraformational clasts or, in some cases, diapir-derived detritus such as reworked clasts of gypsum, carbonate or clastic material associated with surface exposure of the uplifted salt (Lawton and Buck, 2006; Banham and Mountney, 2013a).

In the case of axial-draining fluvial systems for which confining salt-wall-generated topography is linear, elongate and continuous, individual mini-basins tend to be isolated from their neighbours, even where surface relief over the salt wall is minimal. This configuration potentially allows significantly different successions to accumulate between neighbouring basins, in terms of sediment and rate of accumulation.

Generation of topographic lows associated with the development of rim-synclines by the preferential withdrawal of salt from beneath the margins of mini-basins adjacent to salt walls (Barde et al, 2002b; Banham and Mountney, 2013a) can result in the capture of fluvial systems and their confinement to the edges of a single mini-basin (Fig. 6). As such, the distribution of preserved fluvial elements in areas close to the flanks of salt walls can vary markedly from those present in the central part of the same mini-basin (Andrie et al, 2012; Banham and Mountney, 2013a, b). Salt trapped beneath the centre of a mini-basin can result in formation of a turtle-back structure (sensu Barde et al, 2002), where subsidence rates are reduced relative to those of adjacent rim-synclines (Fig. 6). This can result in the generation of a relative high in the centre of a mini-basin that may limit the rate of sedimentation in such regions and may even potentially isolate two marginal rim-synclines to form sub-basins.

In isolated basins, where fluvial activity is limited due to preferential drainage into adjacent basins, active fluvial processes tend to be dominated by: (i) localised reworking and redistribution of sediment from uplifted salt-wall topography; (ii) delivery of sediment via the overspill of drainage pathways from adjacent basins; and (iii) the development of minor drainage pathways within the basin via supply along the basin axis, in some cases in the form of episodic non-confined flow rather than channelised flow (Abdullatif, 1989; Banham and Mountney, 2013b). Non-confined flows that give rise to depositional sediment bodies with thin but laterally extensive sheet-like elements and only minor channel elements are especially common in relatively isolated, sediment-starved basins under semi-arid climates (Rahn, 1967; Williams 1970; Benvenuti, 2005; Banham and Mountney, 2013a, b).

## Mini-basin sediment-fill style

The interplay between the rates of sediment supply and accommodation generation due to subsidence is a key factor that dictates basin-fill style (S) in evolving mini-basins (Fig. 7). For example, for a relatively high and constant rate of sediment supply (Fig. 7), fluvial strata within a slowly-subsiding basin will experience significant reworking as fluvial systems avulse and migrate laterally, preserving only the lower parts of channel-fill elements that become vertically stacked to form multi-storey channel complexes dominated by coarse-grained clastic deposits. In such cases, aggradation rates are relatively low and the middle and upper parts of channel-fill elements, including the sandy bedforms that typically characterise the middle parts of fluvial depositional cycles (Miall, 1996), will be prone to reworking as later channels are cut. In cases like this, where the rate of sediment delivery outpaces the rate of accommodation generation, an ***over-filled basin*** tends to develop, the fill of which is dominated by channel lag and gravel sheet elements arranged into multiple, vertically stacked thin sets and cosets separated by complex arrangements of erosional bounding surfaces (Fig. 7). Bypass of sediment farther downstream within the system as sand-dominated bedload and mud- and silt-dominated suspended load is significant.

In cases where the rate of sediment delivery is broadly in equilibrium with the rate of accommodation generation due to on-going subsidence, fluvial depositional cycles will tend to preserve a relatively complete record of the cut, infill, lateral migration and final abandonment of channel systems (Fig. 7). Such fluvial cycles tend to have an erosional base, a lower part dominated by gravel sheet elements, a middle part dominated by cross-bedded sand-dominated bedform elements, and an upper part dominated by ripple cross-laminated fine-grained sandstone elements (often with climbing-ripple strata) and argillaceous floodplain elements. In some cases, the down-cutting associated with the emplacement and lateral accretion of later channel elements will result in reworking of overbank (floodplain) elements and the transport of argillaceous sediment farther downstream. Over time, this will culminate in the formation of a ***filled basin***, the fill of which will tend to be dominated by a sand-prone basin fill style in which fining-upward depositional cycles are evident.

In cases where the rate of sediment delivery is outpaced by the rate of accommodation generation due to subsidence, the potential vertical aggradation rate will be high but if sedimentation rate does not fill available accommodation an ***under-filled basin*** will develop, the fill of which will be dominated by channelised fluvial elements encased within finer-grained elements of floodplain and overbank origin (Fig. 7). The preservation potential of floodplain packages is greatest in these types of basins, and relatively silt-prone and sand-poor basin-fill styles tend to accumulate as a consequence. Fluvial successions in under-filled basins are commonly intercalated with lacustrine successions, especially in humid-climate settings (Prochnow et al., 2006; Mathews et al, 2007). By contrast, in arid-climate settings, accommodation may remain unfilled and sabkha and playa systems may develop (Banham and Mountney 2013a).

## 4. Case Studies

The expression of fluvial systems accumulating in salt-walled mini-basins varies dramatically, both between and within halokinetic provinces (Bromley, 1981; Matthews et al., 2007; Venus, 2013; Banham and Mountney, 2013 a, b). The various styles of basin fill are herein explored through studies (Table 1) which demonstrate how the interplay between various controlling factors (Fig. 8) act to dictate drainage pathways, the generation of localised depocentres, and the style of fluvial system accumulation and preservation.

### 4.1Salt Anticline Region, SE Utah

The Salt Anticline Province of SE Utah is located in the foredeep of the Paradox Basin, which developed as a flexural foreland basin during the Pennsylvanian in response to loading of the continental plate by the Uncompahgre Uplift of the Ancestral Rocky Mountains (Barbaeu, 2003) (Fig. 9a). The Uncompahgre Uplift formed the northwest margin of the basin, adjacent to the foredeep, and acted as a source of sediment throughout much of the Pennsylvanian and Permian (Kluth and Coney, 1981, Barbeau, 2003), eventually leading to the progradation and accumulation of the alluvial mega-fan of the Cutler Group (Cain and Mountney, 2009, 2011). A series of transgressive-regressive cycles driven by eustatic sea-level changes during the Pennsylvanian (Goldhammer et al., 1984) resulted in the periodic partial isolation of the foredeep region from a larger epeiric sea-way by the forebulge of the Paradox Basin, and this repeated isolation resulted in the accumulation of a thick succession of evaporites (the Paradox Formation).

Pre-salt basement faults generated by brittle deformation associated with flexural down-warping are aligned northwest-to-southeast, parallel to the elongate trend of the uplifted Uncompahgre Front (Doelling, 1988; Barbeau, 2003; Trudgill, 2011). The accumulation of differential salt thicknesses across these basement faults likely initiated salt movement and controlled the location and orientation of the resultant salt walls, which grew to form linear features along the same northwest-southeast trend (Prommel 1923; Shoemaker and Newman, 1959; Doelling, 2002). Halokinesis and mini-basin development commenced in response to loading of salt of the Paradox Formation by accumulating fluvial strata of the overlying Honaker Trail Formation and Cutler Group during the late Pennsylvanian and Permian (Kluth and DuChene, 2009; Trudgill, 2011). Throughout the duration of sedimentation in the Salt Anticline Region, both the direction and rate of sediment supply varied substantially, and these changes are recorded by significant differences in the style of fill of the mini-basins by fluvial strata of the Permian Cutler Group, and the younger Triassic Moenkopi and Chinle formations (see below). Mini-basin subsidence and sediment accumulation continued throughout Permian (Venus 2013), Triassic (Mathews et al, 2007; Banham and Mountney, 2013), and locally into the Jurassic (Doelling, 1988; Bromley, 1991). Four fully developed salt-walled mini-basins developed between the Uncompahgre Front and the Paradox fore-bulge: the Fisher; Parriott; Big Bend (Mathews, 2007; Banham and Mountney, 2013); and Shafer basins (Venus, 2013). Additional mini-basins not described here are present elsewhere in the region, most notably along-strike from these primary basins (Trudgill, 2011).

**The Cutler Group**

The Undifferentiated Cutler Group, of predominantly Permian age, accumulated in the mini-basins of the Salt Anticline Region during a protracted episode characterised by relatively high rates of sediment delivery. Sediment was principally sourced from the eroding Uncompahgre Uplift, a region of significant regional elevation on the northeast flank of the Paradox Basin, and delivered southwest-wards into the Salt Anticline Region, perpendicular to the northwest-southeast trend of the evolving salt walls (Fig. 9b; Werner, 1974; Mack and Rasmussen, 1984; Cain and Mountney, 2009; Venus, 2013). The prevailing climate at this time was dominantly semi-arid (Werner, 1974; Cain and Mountney, 2009), though with evidence for more humid episodes at times (Cain and Mountney, 2009, 2011; Soreghan et al., 2009). Evidence for these climatic variations are recorded in part by the progradation and retrogradation of the Organ Rock Formation, a ~100 m-thick wedge of alluvial strata, which is the lateral equivalent of the Undivided Cutler Group and which interacts with an aeolian dune field in the distal part of the Paradox Basin, beyond the margin of the Salt Anticline Region (Cain and Mountney, 2009).

The transverse drainage orientation relative to the trend of the actively uplifting salt walls resulted in initial preferential deposition and accumulation of fluvial strata in the Fisher mini-basin that developed adjacent to the frontal thrust of the Uncompahgre Uplift, most proximal to the sediment source (Venus, 2013). Throughout most of the episode of accumulation of strata of the Cutler Group, the rate of delivery of sediment significantly outpaced the rate of subsidence, and an over-filled basin state developed in which the accumulating fluvial system was able to rapidly fill available accommodation in the developing mini-basins, sequentially from the most proximal Fisher Basin, and latterly into the Parriott and Big Bend basins (Kluth and DuChene, 2009; Trudgill, 2011). During the late Permian, the fluvial systems were episodically able to deliver sediment beyond the distal limits of the Salt Anticline Region, leading to progradation of the Organ Rock Formation (Cain and Mountney, 2009).

The styles of fluvial sediment fill within the subsiding mini-basins document an architectural expression that records high rates of sediment delivery that resulted in an over-filled basin state and the preservation of a sand- and gravel-prone fill-style (Venus, 2013). The total thickness of accumulated Cutler Group sediments in each mini-basin systematically decreases from the more proximal Fisher Basin to the more distal Shafer Basin across the Salt Anticline Region (Paz and Trudgill, 2009; Trudgill, 2010). Basin-fill styles demonstrate a progressive fining trend from the proximal Fisher Basin into the Parriott, Big Bend and finally into the Shafer basins (Cain and

Mountney, 2009, 2011; Venus, 2013). This occurred, in-part, due to systematic fining of the sediment in transport with increasing distance down-steam, and implies a decrease in fluvial energy and transport capacity that is typical of alluvial mega-fans or distributive fluvial systems, due to radial spreading and transmissions loss (Fisher and Nichols, 2007; Hartley et al., 2010; Weissmann et al., 2010). Additionally, downstream fining of the calibre of accumulated sediment was also influenced by episodic salt-wall uplift, which episodically resulted in repeated diversion of fluvial drainage pathways to orientations parallel to the trend of the salt walls and the ponding of flood-water and sediment behind growing salt-wall topography (Venus, 2013). These “pond” elements are characterised by non-channelised elements containing high proportions of mica, suggesting an episodic damming of flood-waters that resulted in the accumulation of deposits from slow-flowing or standing water in areas directly upstream of uplifted salt-wall topography.

The damming of floodwater required the emergence of localised relief associated with growing salt-wall topography and such episodes record the episodic transition to a temporarily under-filled basin state. Ponding of sediment behind salt walls likely corresponded to episodes of decreased fluvial activity at times of heightened climatic aridity: fluvial deposits indicative of such conditions are characterised by surfaces with desiccation cracks in fine-grained strata and sedimentary structures such as climbing ripples and trough cross-bedding that record palaeoflow indicators that are diverted or even reversed compared to the dominant south-westerly trend (Venus, 2013). Despite evidence to show that salt-wall topography influenced fluvial drainage pathways, few salt clasts are preserved in the Cutler Group accumulations that form the main fill of the mini-basins, and this demonstrates that the salt walls themselves were unlikely to have breached the land surface.

Episodic resurgence of fluvial activity led to overtopping of salt walls and such events likely corresponded to more humid climatic episodes. For such events, palaeoflow indicators record transport directly across buried salt walls suggesting the burial of any earlier surface topographic expression. Localised reworking of fluvial strata from atop salt walls is demonstrated by an increase in the occurrence of intraformational rip-up clasts in sediment accumulations directly downstream from the buried but slowly growing salt walls (Cain and Mountney, 2009; Venus, 2013).

**The Moenkopi Formation**

The resumption of fluvial accumulation in the Salt Anticline Region of the Paradox Basin in the early Triassic coincided with a significant change in the climatic and drainage regime of the region. Sediment accumulation during this period occurred under the influence of an arid to hyper-arid regime (Blakey and Ranney, 2008; Banham and Mountney, 2013b). Palaeodrainage from the Uncompahgre Highlands had diminished significantly by the onset of deposition of the Moenkopi Formation, and instead a new dominant drainage pathway had become established that was sourced in the San Luis and Defiant Upwarp region to the southwest (Fig. 9c), and which drained north-westward through the Salt Anticline Region (Stuart et al*,* 1972; Blakey, 1974; Banham and Mountney, 2013b). Sediment was delivered axially into what was by then a series of well-developed northwest-southeast-trending mini-basins: the Fisher, Parriott and Big Bend Basins, each of which was apparently isolated from its neighbouring basins in terms of drainage pathways.

Subsidence rates between the basins varied throughout accumulation of the Moenkopi Formation, in part due to the inheritance of basin-fill geometries from the previously accumulated deposits of the Cutler Group. These inherited basin-fill geometries, where the basin-fill thicknesses varied both within and between basins, were important in controlling the style, timing and rate of mini-basin subsidence throughout accumulation of the Moenkopi Formation. The basin-fill thickness was thinnest adjacent to the Uncompahgre Uplift (where Cutler Group sediments were preferentially deposited in the earlier phase of basin fill) and thickened towards more central parts of the Paradox Basin (Banham and Mountney, 2013a). Variable rates of both subsidence and sediment supply resulted in the accumulation of both sand-prone (filled) and sand-poor (under-filled) basin-fill styles in neighbouring basins for the same stratigraphic levels. The Fisher Basin, which had little remaining remnant basin-fill potential (i.e., available accommodation) due to it approaching salt-weld formation by the onset of the early Triassic, experienced low rates of subsidence (Banham and Mountney, 2013a). This, coupled with high rates of sediment delivery (both from the San-Luis Uplift region and from the remnant Uncompahgre Highlands as a secondary source), resulted in the accumulation of a sand-prone basin-fill style, apparently early in the history of accumulation of the Moenkopi Formation. As the basins evolved during the early Triassic, the rate of sediment delivery to the Fisher Basin progressively diminished, in-part due to the final denudation of the Uncompahgre Uplift, resulting in the basin-fill style becoming progressively less sand-prone upwards.

In the adjacent Parriott Basin, the basin-fill potential varied spatially, due to the variable thickness of the inherited basin-fill: basin-fill potential was least on the side of the basin closest to the Uncompahgre Front where a thick succession of Cutler Group sediments had accumulated, and greatest on the distal margin of the basin. This resulted in preferential subsidence on the distal margin of the Parriott Basin, culminating in formation of a rim-syncline in the latter stages of accumulation of the Moenkopi Formation. This rim-syncline, which formed a locus of subsidence and which apparently formed a topographic low, acted as a preferential drainage corridor, leading to the accumulation of stacked fluvial channel-fill elements during the final phases. During accumulation of the middle and upper parts of the Moenkopi Formation, high rates of salt-wall uplift, combined with an increase in climatic aridity resulted in a relative reduction in the rate sediment supply (Banham and Mountney, 2013a) and this allowed the Castle Valley Salt Wall, which separated the Parriott and Big Bend basins, to breach the land-surface (Lawton and Buck, 2006). Detritus derived from surface exposure of this salt wall was subsequently reworked into discrete gypsum-clast-bearing units, which are preserved in the succession around the flanks of the Castle Valley salt wall, in both the Parriott and Big Bend basins. The Big Bend Basin, which occupied a position further from the Uncompahgre Uplift, had the greatest basin-fill potential at the onset of accumulation of the Moenkopi Formation and the rate of sediment supply to this basin is interpreted to have been slightly higher than that to the Parriott Basin because a greater proportion of channel elements are preserved throughout the stratigraphic succession, especially in the upper part where the in-fill of a pronounced rim-syncline is dominated by amalgamated channel-fill elements, indicating preferential concentration of drainage pathways in this topographic low.

**The Chinle Formation**

By onset of accumulation of the Chinle Formation, the Fisher and Parriott Basins had effectively grounded, allowing only very limited additional accumulation in these basins, in response to slow, load-induced sediment compaction. The palaeodrainage direction remained axial to the salt walls, towards the northwest. The final burial of the remnants of the Uncompahgre Uplift signified a final cessation of sediment derived from the northeast (Trudgill, 2011). Climate during accumulation of the Chinle Formation changed from humid to semi-arid (Fig. 9d; Prochnow et al, 2006). This shift in climate is recorded by a change in fluvial style: architectural elements in the lower part of the formation are characterised by thick and well-developed palaeosols that are intercalated with thick channel-fill elements, the internal fill of which is dominated by asymptotic-based cross-stratified sets indicative of accumulation of a coarse-grained meandering river system in which lateral accretion processes dominated (Hazel, 1994); the upper part of the formation, records laminated sand sheet elements and sandy bedforms containing crudely-bedded conglomerate and coarse-grained sandstone trough cross-stratified sets and scour fills, indicative of accumulation of a gravelly low-sinuosity fluvial system (Hazel, 1994).

Preferential accumulation of fluvial deposits of the Cutler Group and Moenkopi Formation in the basins closest to the Uncompahgre Uplift limited the remnant basin-fill potential of the Fisher and Parriott basins, which by the Late Triassic had effectively grounded. However, in the Big Bend basin, lower rates of subsidence throughout evolution of this basin, and the retention of residual salt beneath this basin, allowed continued localised subsidence into the Late Triassic. This is reflected by variations in thickness in units of the lower part of the Chinle Formation, which vary from <10 m thick in the northern-most part of the Paradox Basin, to over 50 m thick in localised depocentres, such as parts of the Big Bend Basin, and near the Cane Creek Anticline of the Shafer Basin to the southwest of the town of Moab (Fig. 9d; Matthews et al, 2007). These localised depocentres record the final phases of subsidence associated with salt displacement from beneath mini-basins where axial variations of salt thickness or rate of salt flow resulted in localised grounding relatively late in the history of evolution of the Salt Anticline Region and the concomitant accumulation of a thicker succession.

In addition to variations in preserved thickness, angular discordances between the Chinle and Moenkopi formations, together with intraformational unconformities within the Chinle Formation, indicate ongoing tilting of strata by halokinesis in some parts of the succession (Matthews et al, 2007). Areas of relatively high rates of subsidence in basin centres were typically poorly drained, resulting in accumulation of lacustrine elements, especially during the lower part of the Chinle Formation (Matthews et al, 2007). Multi-storey channel elements and palaeosols accumulated toward the basin margins during episodes of halokinetic quiescence, with higher proportions of channel elements and palaeosol maturity apparently increasing with the duration of quiescence (Prochnow et al, 2006; Matthews et al, 2007). In the upper part of the Chinle Formation, an increase in aridity is recorded by a change from a meandering to a braided system (Hazel, 1994). During this episode palaeodrainage was oblique to salt-wall axes; aggradation of the basin fill to the level where available accommodation was filled temporarily allowed cross-salt-wall drainage, before renewed uplift of the salt walls resulted in deflection of the drainage pathways parallel to salt walls (Mathews et al, 2007).

### 4.2 Pre-Caspian Basin

Salt-walled mini-basins of the Pre-Caspian salt tectonic province are situated in the Pre-Caspian Basin, which developed as a rift basin during the Devonian (Pairazian, 1999; Barde et al., 2002a) (Fig. 10a). Throughout the Carboniferous, approximately 2000 m of carbonate strata recording reef development accumulated in the basin at a time when it was largely starved of clastic sediment input (Schamel, 1995). This episode of carbonate accumulation ceased when the Pre-Caspian Basin became partially isolated from the regional sea due to uplift of the developing Ural Mountains to the east during the Late Carboniferous (Barde et al., 2002a; Volozh et al, 2003). This restriction of marine water circulation resulted in repeated desiccation of the basin, culminating in the Middle Permian in the accumulation of up to 4500 m salt during the Kungurian to Kazanian (Garalla and Marsky, 2000; Barde et al., 2002a). Onset of halokinesis was linked to further uplift of the Urals, either by lateral shortening and orogenic collapse (Ings and Beaumont, 2010; Newell et al., 2012) or by sediment loading induced by accumulation of clastic sediments derived from the Urals. The orientation of the developing salt walls was controlled either by pre-existing basement trends inherited from the original onset of the Pre-Caspian Basin or from the ensuing uplift of the Urals (Barde et al., 2002a; Brown et al., 2004). These events resulted in the development of complex basement trends, which are expressed by the distribution of salt-wall geometries: linear salt walls with a north-south trend typically developed in the east and these follow basement faults sympathetic to the trend of the Urals, in the rest of the Pre-Caspian Basin, basement faults are typically oriented northeast-southwest and southeast-northwest, having been generated during the initial evolution of the basin (Barde et al., 2002a). These competing basement trends resulted in the generation of salt walls with polygonal geometries throughout the rest of the basin (Barde et al, 2002a; Volozh et al, 2003).

Sediment accumulation within the mini-basins occurred from Late Permian through to present, with rates of halokinesis having decreased significantly since the Triassic. Up to 6 km of sediment has accumulated in the western mini-basins (Barde et al., 2002; Newell et al., 2012). Most subsurface studies have focused on the Permian and Triassic parts of the basin fill since these host significant hydrocarbon plays (Barde et al., 2002a; O'Hearn et al., 2003, Volozh, et al., 2003); more recently, field-based studies have additionally been undertaken where outcrop allows (Newell et al, 2012).

Sediment was delivered into the Pre-Caspian Basin by fluvial systems emanating from the Ural Mountains (Newell et al., 2012) that drained transverse to the north-south trend of the linear mini-basins in the eastern part of the region. Climate during the Tatarian (Upper Permian) was semi-arid to sub-humid, whereas Triassic strata accumulated under a more arid climatic regime (Newell et al., 1999). The polygonal salt walls in much of the basin exerted a significant control on fluvial pathways and sediment distribution, resulting in the evolution of multiple of sedimentary environments within adjacent basins (Fig. 10b). Where fluvial systems incised into uplifting salt walls and maintained their drainage pathways, braided river and associated facies dominated the basin fill (Barde et al., 2002b). Where mini-basins became endorheic, due to uplift of a salt wall on the downstream margin of the basin, intra-basin lakes developed, with the fluvial systems terminating as lacustrine fan-deltas (cf., Nichols and Fisher, 2007). This is expressed as a basin-fill style that is sand-prone at its upstream margin but which is dominated by heterolithic siltstone, mudstone and potentially lacustrine organic-rich argillaceous strata in its central part. Where mini-basins remained isolated due to diversion of drainage pathways or high rates of salt-wall uplift driven by displacement of salt from beneath adjacent basins, the resulting accumulation is dominated by a sediment-starved basin-fill style characterised by the accumulation of continental evaporites. Locally, salt-glaciers derived from gravity collapse of salt-wall segments that breached the ground surface and conglomerate horizons derived from the reworking of clastic material from the flank of the uplifting salt wall or diapir itself contributed coarse detritus to the margins of such basins (cf., Lawton and Buck, 2006; Buck *et al.*, 2010).

Basin-fill style between the mini-basins varies dramatically, as recorded by subsurface well-log data (Barde et al., 2002b). This is a function of several factors: inheritance and capture of pre-existing drainage pathways as mini-basins developed; maintenance of salt-wall height at a rate of uplift that preserved a long-lived topographic expression capable of effectively partitioning neighbouring mini-basins and localised differences in the rate of sediment delivery and accumulation to fill available accommodation in adjacent mini-basins. Presently, there is a lack of detailed sedimentological description for the basin-fill styles of the mini-basins in the Pre-Caspian Basin.

### 4.3 North Sea – J Block/Skagerrak Formation

During the Middle Permian, ongoing subsidence of the northern Rotliegend Basin (located in the subsurface beneath the present-day North Sea) allowed the Zechstein Sea to flood into the basin from the north via a sill or narrow inlet (Ziegler, 1975). Repeated closure of this inlet, coupled with prevailing arid conditions and high evaporation rates, led to restricted recharge of the fluids circulating in the basin and culminated in accumulation of a series of thick evaporate sequences of the Zechstein Group (Smith and Taylor, 1992). The total thickness of these evaporite layers in the Central Graben region exceeds 1.5 km (Fig. 11a) (Stewart and Clarke, 1999; Glennie et al, 2004). Initiation of halokinesis occurred during the Early Triassic in response to differential loading of the salt by prograding clastic fluvial wedges from the north, and by thin-skinned extension causing reactive diapirism (Stewart and Clarke, 1999). It is likely that basement faults exerted a fundamental control on the orientation and distribution of salt walls, which follow the primary NNW-SSE-oriented fault trends of the Central Graben (Fig 11a) (Hodgson et al, 1992; Peacock, 2004). The duration of halokinesis in the Central Graben was dictated by the thickness of salt beneath the mini-basins, with basins developed over palaeo-highs where the salt was thinner, typically grounding on the pre-salt basement during the Early- to Middle-Triassic, and salt basins formed over thicker successions of salt grounding in the late Triassic or Jurassic (Smith et al., 1993). In some instances, salt walls began to collapse due to secondary axial migration away from the wall or in response to dissolution after grounding of the adjacent mini-basins. These processes led to the development of secondary mini-basins over the original salt-wall crests (Smith et al, 1993).

The Triassic fill of the Central Graben consists of two main formations: the Lower Triassic Smith Bank Formation, which is of fluvio-lacustrine origin (Smith et al, 1993; Goldsmith et al, 2004); and the Middle- to Upper-Triassic Skagerrak Formation, which is divided into the Judy, Joanne, and Josephine sandstone members, and the Julius, Jonathan and Joshua mudstone members (Fig 11a, Goldsmith et al, 1995). Provenance of the lower parts of the Skagerrak Formation demonstrates a sediment source almost exclusively from the Shetland Platform, with sediments in the upper parts of the formation having been sourced from both Scotland and Fennoscandia (Mange-Rajetzkey, 1995). This temporal change in sediment provenance could have arisen in response a change in climate, a change in tectonic regime on the Fennoscandian margin, or change in rate of halokinesis that could have resulted in a change of configuration of sediment supply (cf., Moenkopi Formation). The Judy and Joanne sandstone members form two of the main hydrocarbon plays of the Triassic syn-halokinetic sequence in the

Central Graben of the North Sea (Goldsmith et al, 1995; McKie, 2011).

The Judy Sandstone Member is interpreted to be a dryland terminal fluvial system, characterized by: stacked, low-sinuosity, high width-depth ratio channel-fill elements; terminal splay and flood-out elements; and ephemeral playa lake elements (McKie et al., 2010; McKie, 2011). Throughout the Middle- to Late-Triassic, sediment was delivered axially in an orientation parallel to the uplifted salt walls, with surface topography above growing salt walls apparently causing neighbouring mini-basins to develop in isolation. This resulted in contrasting styles of basin fill between adjacent mini-basins (Fig. 11b; Hodgson et al., 1992; Smith et al., 1993). Temporary closure of basins resulted in the accumulation of ephemeral lakes or salt pans, with much of the fill characterised by argillaceous heterolithic facies, the total thickness of which was partly dependent on the duration of isolation. During these episodes of drainage diversion, fluvial pathways were apparently concentrated preferentially along the axes of some basins, resulting in the coeval development of sand-prone basin fill styles in certain mini-basins, yet notably sand-poor fill-styles in neighbouring mini-basins.

## 4.4 La Popa Basin

La Popa Basin, northern Mexico, developed during the Late Jurassic as a pull-apart basin associated with movement on the Coahuila-Tamaulipas transform and roll-back and eventual foundering of the Mezcalera Plate (Dickinson and Lawton, 2001; Andrie *et al.*, 2012), in which the Minas Viejas evaporite accumulated (Fig. 12a) (Rowan et al, 2012). The onset of halokinesis occurred during the Late Jurassic, though the initiation mechanism remains unresolved, with (i) extension, thermal and loading subsidence that caused tilting, or (ii) differential loading by sediments derived from the Coahuila Platform each having been proposed as the likely cause (Rowan et al, 2012). Sediment accumulation continued in La Popa Basin from the Late

Jurassic to the Eocene, with the Cretaceous and Palaeocene stratigraphy being dominated by marine sedimentation (Lawton et al, 2001). The Hidalgoan Orogeny (Cretaceous-Palaeogene) generated the Sierra Madre Fold-Thrust belt, which was the principal source region of sediment to the La Popa Basin to the south (Rowan et al, 2012). The fluvial Carroza Formation accumulated during the late Eocene at the end of a major marine regression (Andrie et al, 2012), and was later deformed during formation of the Carroza syncline by continued salt displacement and crustal shortening by the Hidalgoan Orogenic event (Rowan et al, 2012).

The distribution of fluvial elements in the Carroza Formation demonstrates how halokinetic processes can control the position of drainage pathways adjacent to uplifting salt walls, and study of these has led to development of models with which to predict the distribution of sand bodies and the reservoir potential (Fig. 12b; Andrie et al., 2012). The Lower Member of the Carroza Formation is characterised by wide but isolated channel-fill complexes, with varying palaeocurrents (Andrie et al., 2012). Sedimentation rates were relatively high compared to subsidence rates during initial phase of deposition, allowing for unconfined migration of the fluvial system across the basin. During accumulation of the Middle Carroza Member, fluvial elements began preferentially stacking around the position of the Carroza syncline hinge, which by this time had migrated toward the flank of La Popa salt wall. Palaeodrainage direction became preferentially aligned parallel to this northwest-southeast trending salt wall, indicating that sufficient topography had been generated to capture drainage pathways. During accumulation of the Middle Carroza Member, the rate of sedimentation relative to that of subsidence decreased, allowing the uplifting salt wall to generate surface topography and resulting in the capture of drainage pathways. The topographic expression eventually became sufficient to trigger debris flows from salt-wall flanks. Alignment and stacking of fluvial channel elements indicates that drainage pathways were confined to the developing rim syncline adjacent to the salt wall at this time. During accumulation of Upper Carroza Member, deposition of channel elements became restricted to the immediate flank of the salt wall, as controlled by the migrating axis of the Carroza Syncline. The progressive migration of the Carroza syncline axis and associated development and migration of a rim-syncline controlled the locus of drainage and the progressive shift of the accumulation of fluvial channel elements towards the flank of La Popa salt wall over time.

## *5. Discussion*

The interplay between the rate of mini-basin subsidence and the rate of sediment accumulation, including variation in these over space and time, is a key control on the development of fluvial systems and their preserved successions in salt-walled mini-basins. Mini-basin subsidence rates are determined by the rate of underlying salt evacuation, which will vary over time in response to changes of sediment loading and proximity to grounding on the pre-salt basement, typically resulting in a non-linear subsidence history. Rates of sediment delivery, which are themselves partly controlled by external factors such as climate and source area, dictate sediment type, composition, availability and delivery rate. Dominant drainage pathway and the nature of any sediment bypassing that may occur (itself controlled by subsidence rates) can result in significant variations in the style of sediment delivery and rate of accumulation between and within mini-basins. Figure 13 demonstrates how the interplay between sediment supply rate and subsidence rate can control basin-fill style.

In cases where rates of sediment delivery and subsidence are balanced, avulsion coupled with lateral migration of channels tends to rework and remove the majority of fine-grained overbank elements, resulting in high proportions of sand-prone channel-fill elements. Where the rate of subsidence outpaces sediment delivery, the fluvial system will aggrade to preserve near-complete fluvial fining-upward cycles, and avulsion and lateral migration are less likely to rework or remove these overbank elements. Such conditions result in sandy channel-fill elements becoming vertically isolated from each other. Such a situation may culminate in the formation of perennial lakes,which can lead to accumulation of thick lacustrine successions. Conversely, where the rate of sediment delivery significantly outpaces the rate of subsidence, lack of available accommodation means that deposited strata are prone to reworking, thereby preserving only the lowermost parts of fluvial depositional cycles, such as pebbly basal lags. Many such deposits have poor reservoir potential. Where rates of sediment delivery greatly exceed rates of subsidence, fluvial systems can potentially overspill confining salt walls, resulting in drainage diversion into adjacent basins, especially where salt walls are arranged in polygonal patterns. Alternatively, where growing salt walls generate a long-lived surface expression, fluvial systems may become confined within a single mini-basin, leaving adjacent basins isolated from the primary source of sediment input. These isolated mini-basins can potentially become dominated by evaporitic processes and salt flats might form (Goodall, 2000), whereas thick palaeosols and possibly coal swamps may develop in more humid settings. Eventual overtopping of salt walls by fluvial systems will result in incision and diversion of drainage into adjacent under-filled mini-basins; such events may either temporarily or permanently change the fill-style of the neighbouring basin.

## Evolution of Fluvial Systems in salt-walled mini-basins

Consideration of various case-studies has allowed for a series of generic evolutionary models to be proposed for both linear and polygonal salt-walled mini-basins, and these depict the expected sequence of basin-scale sedimentary architectures (Figs 14, 15).

### Linear basins

**Initiation** of salt-walled mini-basin growth (Fig. 14a) triggered by sediment loading in combination with a tectonic mechanism results in initial diversion of fluvial systems in cases where they are unable to respond sufficiently quickly to downcut through growing salt highs. Mini-basins closest to the sediment source will tend to fill quicker and therefore tend to be sand- or gravel-prone; conversely, more distal mini-basins may develop relatively sand-poor basin-fill styles.

**Rapid subsidence** may occur due to a rapid influx of sediment, which will act to load the salt beneath the developing basin (Fig. 14b). A reduction in the rate of sediment supply, for example due to a change to a more arid climatic regime, may contribute to fluvial diversion during episodes when the rate of sediment delivery is outpaced by subsidence and salt wall uplift. Mini-basins may become isolated for short episodes, developing under-filled basin segments, or for longer episodes where the style of basin infill may record isolation (e.g., evaporate basins). The development of turtle-back structures and rim synclines may act to partition drainage pathways within a single mini-basin.

**Temporary sediment supply shutdown** driven by a region-wide shift in climatic regime may significantly reduce fluvial activity resulting in reduced sediment accumulation rates in all mini-basins across a province (Fig. 14c). Such a climate change may also coincide with a change of sediment source region. Such a change may account for the development of adjacent sand-prone and sand-poor basins. When salt wall uplift rates outpace sediment accumulation, salt walls may breach the land surface, creating a source of detritus that can be locally reworked.

**Basin grounding** will eventually occur once subsiding mini-basins have effectively displaced the underlying salt, and this will prevent further subsidence (Fig. 14c). Due to the differential subsidence history of adjacent mini-basins or the onset of salt-wall collapse, subsidence may continue locally and may lead to the development of secondary basin atop collapsing salt walls. A shift to more humid conditions can result in the development of meandering fluvial systems, with ephemeral or perennial lakes developing in some basins.

## Polygonal basins

**Initiation** of polygonal salt walls in response to multiple basement trends can lead to development of blind or closed (i.e., isolated) mini-basins, with fluvial systems becoming confined to a series of adjacent basins (Fig. 15a). Blind basins can be characterised by lacustrine elements with deltas developing on lake margins.

**Rapid subsidence** and ensuing salt-wall uplift can disrupt drainage through the region, with previously open basins becoming closed or blind (Fig. 15b). Some isolated basins can become dominated by evaporitic or aeolian processes, especially those distal to sediment source area in arid settings.

**Drainage diversion** may occur when sediment accumulation in a closed basin out-strips subsidence rates, allowing the fluvial system to breach a salt wall and incise a new drainage pathway into a neighbouring basin (Fig. 15c).

**Onset of basin grounding** will tend to occur first in those mini-basins that experienced high rates of sediment delivery and therefore rapid subsidence (Fig. 15d). Partial collapse of certain salt walls in the aftermath of grounding will tend to promote unrestricted inter-basin drainage, whilst other salt walls may continue to grow and therefore prevent sediment delivery to adjacent basins. Heel-toe geometries with intraformational unconformities may develop where subsidence switches from one side of a single mini-basin to the other.

## Gaps in understanding

Despite having been the subject of numerous studies since the early 1990s (e.g., Bromley, 1991), there remain significant gaps in our understanding of the mechanisms governing the accumulation of fluvial stratigraphy in salt-walled mini-basins. Of the 120 evaporite provinces documented by Hudec and Jackson (2007) that are known to have undergone salt deformation, many examples that demonstrate syn-halokinetic evolution of fluvial systems exist, both in the subsurface and in outcrop. The majority of recent detailed outcrop studies have been conducted in the Paradox Basin of Utah, with fewer studies undertaken in other outcropping basins, such as the Pre-Caspian Basin, or La Popa Basin.

Of the case studies considered in this study, 7 examples accumulated under semi-arid or arid climates, which is reflected by a dominance of braided fluvial networks, with evidence for evaporitic processes and development of calcisols and aridisols. This may reflect the fact that most outcropping salt basins, which were later mobilised to form salt-walled mini-basin provinces, developed between the late Pennsylvanian and Permian, with fluvial sediment accumulation occurring during globally arid period of the Permian and Triassic. Studies of meandering fluvial systems preserved in ancient salt mini-basins are under-represented in the literature, in part due to a lack of recognition of suitable outcrops for study.

Spatial variations in rates of subsidence along mini-basin axes and the creation of local depocentres have yet to be studied in detail. Such spatial variations could result in local accumulation of lacustrine deposits, or could act as a mechanism for controlling the location of nodal avulsions, flood-outs or points of convergence of high-aspect-ratio channels, thereby controlling the distribution of sand-prone channel-fill elements or thin sheet-like heterolithic elements.

Detailed analysis of the controls on drainage pathways in polygonal salt-walled mini-basins also requires further study. Switching of drainage pathways and total or partial isolation of certain mini-basins reflects local changes in sediment delivery. Locally increased rates of accumulation can lead to accelerated rates of subsidence. This may drive positive feed-back cycles that promotes or accelerates mini-basin development.

## *6. Conclusions*

Evolution of salt-walled mini-basins and the ensuing accumulation of fluvial strata within these basins is an inherently complex process with multiple factors controlling sediment distribution, both within and between mini-basins. The style of basin fill can evolve independently between neighbouring mini-basins of equivalent age, most notably where drainage networks are aligned parallel to elongate salt walls such that they become partitioned from each other by growth of salt-related surface topography. Alternatively, where the preferred orientation of fluvial drainage is transverse to linear salt walls, mini-basins between salt walls may fill sequentially with increasing distance from the sediment source. Polygonal networks of salt walls for which surface topographic expression is present may result in the development of partially closed mini-basins occupied by lacustrine systems and fed by fan deltas.

Where rates of salt-wall uplift and mini-basin subsidence, which in combination define the rate of generation of accommodation, are matched or exceeded by the sediment delivery and accumulation, mini-basins become filled and surface topographic expression is overwhelmed, allowing fluvial systems to flow without significant interference and potentially allowing for correlation between separate mini-basin stratigraphic fills.

By reviewing a series of case studies, common examples of fluvial stratigraphic response to different types of salt-walled mini-basins have been identified; a series of generic models have been synthesised that demonstrate the expected evolutionary history for linear and polygonal salt-walled mini-basins. Generic basin evolution models demonstrate the inherent complexities present within mini-basins separated by elongate, linear salt walls and how basin-fill style might vary over time and space depending on sediment delivery (Fig. 14a-d). Such controls dictate how fluvial architectural elements are distributed throughout the course of basin evolution. Figure 15a-d demonstrates a typical evolutionary sequence for a polygonal arrangement of salt-walled mini-basins, showing the predicted distribution of fluvial architectural elements and how drainage pathways can be diverted by various controls operating within the mini-basin province.

These case demonstrate that, fluvial facies and architectural-element distributions can be predicted both within and between mini-basins. The predictive models presented here are of value in assessing the distribution of sand-prone elements within subsurface reservoirs. However, improved techniques for understanding architectural-element distribution and prediction of climate regimes will require good well control and high-quality seismic to predict the probable locations of sand fairways for systems known only from the subsurface.


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## *References*


Abdullatif, O.M., 1989. Channel-fill and sheet-flood facies sequences in the ephemeral terminal River Gash, Kassala, Sudan. Sedimentary Geology 63, 171-184.

Andrie, J.R., Giles. K.A., Lawton, T.F., Rowan, M.G., 2012. Halokinetic-sequence stratigraphy, fluvial sedimentology and structural geometry of the Eocene Carroza Formation along La Popa salt weld. La Popa Basin, Mexico. In: Alsop, G.I., Archer, S.G., Hartley, A.J., Grant, N.T., Hodgkinson, R. (Eds), Salt

Tectonics, Sediments and Prospectivity. Special Publication 363, Geological Society, London, pp. 59-79.

Banham, S.G. Mountney, N.P., 2013a. Controls on fluvial sedimentary architecture and sediment-fill state in salt-walled mini-basins: Triassic Moenkopi Formation, Salt Anticline Region, SE Utah, USA. Basin Research. DOI: 10.1111/bre.12022.

Banham, S.G. Mountney, N.P., 2013b. Climatic versus halokinetic control on sedimentation in a dryland fluvial succession. Sedimentology. DOI: 10.1111/sed.12064

Barbeau, D.L., 2003. A flexural model for the Paradox Basin: implications for the tectonics of the Ancestral Rocky Mountains. Basin Research 15, 97-115

Barde, J-P., Garalla, P., Harwijanto, J., Marsky, J., 2002a. Exploration at the eastern edge of the Precaspian basin: Impact of data integration on Upper Permian and Triassic prospectivity. American Association of Petroleum Geologists, 86, 399-415.

Barde, J-P., Chamberlain, P., Galavazi, M., Gralla, P., Harwijanto, J., Marsky, J., Van Den Belt, F., 2002b. Sedimentation during halokinesis: Permo-Triassic reservoirs of the Saigak Field, Precaspial Basin, Kazakhstan. Petroleum Geoscience 8, 177-187.

Barton, D.C., 1933. Mechanisms of formation of Salt Domes with special reference to Gulf Coast Sat Domes of Texas and Louisiana. American Association of Petroleum Geologists Bulletin 17, 1025-1083.

Benvenuti, M., Carnicelli, S., Ferarri, G., Sagri, M., 2005. Depositional processes in latest Pleistocene and Holocene ephemeral streams of the Main Ethiopian Rift (Ethiopia). In: Blum, M.D., Marriott, S.B., Leclair, S.F. (Eds), Fluvial Sedimentology VII, Special Publication, 35, International Association Sedimentologists, pp. 277-294.

Blakey, R.C., 1974. Stratigraphy and depositional analysis of the Moenkopi Formation, southeastern Utah. Utah Geological and Mineral Survey Bulletin 104.

Blakey, R.C., Ranney, W., 2008. Ancient Landscapes of the Colorado Plateau. Grand Canyon Association, Grand Canyon, Arizona.

Bromley, M.H., 1991. Architectural features of the Kayenta Formation (Lower Jurassic), Colorado Plateau, USA: relationship to salt tectonics in the Paradox Basin. Sedimentary Geology 73. 77-99.

Brown, D., Alvarez-Marron, J., Perez-Estaun, A., Gorozhanina, Y., Puchkov, V., 2004. The structure of the south Urals foreland fold and thrust belt at the transition to the Precaspian basin. Journal Geological Society, London 161, 813-822.

Brun, J.-P., Fort, X., 2003. Compressional salt tectonics (Angolan margin). Tectonophysics 382, 129-150.

Buck, B.J., Lawton, T.F., Brock, A.L., 2010. Evaporitic paleosols in continental strata of the Caroza Formation, La Popa Basin, Mexico: record of Paleogene climate and salt tectonics. Geol. Soc. America Bull. 122, 1011-1026.

Byrd, T., Kristian, M., Jaffey, N., Vines, J., Buddin, T., Ekstrand, E., 2004. 3D Structural Restoration of Ponded-Basin Turbidite Reservoirs in a Linked Extensional-Compressional Salt Basin, Holstein Field, Deep-Water Gulf of Mexico. In: Post, P., Olson, D., Lyons, K., Palmes, S., Harrison, P., Rosen, N. (Eds), Salt sediment interactions and Hydrocarbon Prospectivity; Concepts, Applications, and Case Studies for the 21st Century. GCSSEPM 24th Annual Conference, pp. 965-996.

Cain, S.A., Mountney, N.P., 2009. Spatial and temporal evolution of a terminal fluvial fan system: the Permian Organ Rock Formation, South-east Utah, USA. Sedimentology 56, 1774-1800.

Cain, S.A., Mountney, N.P., 2011 Downstream changes and associated fluvial-aeolian interaction in an ancient terminal fluvial fan system: the Permian Organ Rock Formation, SE Utah. In: Davidson, S., Leleu, S., North, C. (Eds.), From Rivers to Rock Record. Special Publication 97 SEPM, pp. 165-187.

Craggs, S., Keighley, D., Park, A., Waldron, J.W.F., 2013. Stratigraphy and salt tectonics of Mississippian-Pennsylvanian strata of the northern Cumberland Basin, Maringouin Peninsula, southeastern New Brunswick. Atlantic Geology, 49, ##-##.

Doelling, H.H., 1988. Geology of salt valley anticline and Arches National Park, Grand County, Utah. In: Doelling, H.H., Oviatt, C.G., Huntoon, P.W. (Eds.) Salt Deformation in the Paradox Region. Utah Geological Survey Bulletin 122, pp. 7-58.

Dickinson, W.R., Lawton, T.F., 2001. Carboniferous to Cretaceous assembly and fragmentation of Mexico. Geological Society of America Bulletin 133, 1142-1160.

Dyson, I.A., 2004. Interpreted Shallow and Deep-Water Depositional Systems of the Beltana Mini-Basin in the Northern Flinders Ranges. South Australia. In: Post, P., Olson, D., Lyons, K., Palmes, S., Harrison, P., Rosen, N. (Eds), Salt sediment interactions and Hydrocarbon Prospectivity; Concepts, Applications, and Case Studies for the 21st Century. GCSSEPM 24th Annual Conference, pp. 593-614.

Fisher, J.A., Nichols, G.J., 2007. Interpreting the stratigraphic architecture of fluvial systems in internally drained basins. Journal of the Geological Society, London 170, 57-65.

Glennie, K., Higham, J., Stemmerik, L., 2003. Permian. In: Evans, D., Graham, C., Armour, A., Bathurst P. (Eds.), The Millennium Atlas: Petroleum Geology of the Central and Northern North Sea, pp. 91-103.

Goldhammer, R.K., Oswald, E.J., Dunn, P.A., 1991. Hierarchy of stratigraphic forcing: Example from Middle Pennsylvanian self carbonates of the Paradox basin. Kansas Geological Survey Bulletin 233, 361-413.

Goldsmith, P.J., Rich, B., Standring, J., 1995. Triassic correlation and stratigraphy in the South Central Graben, UK North Sea. In: Boldy, S.A.R. (Ed.), Permian and Triassic Rifting in Northwest Europe. Special Publication, 91. Geological Society, London, pp. 123-143.

Goldsmith, P.J., Hudson, G., Van Veen, P., 2003. Triassic. In: Evans, D., Graham, C., Armour, A., Bathurst P. (Eds.), The Millennium Atlas: Petroleum Geology of the Central and Northern North Sea, pp. 105-127.

Goodall, T.M., North, C.P., Glennie, K.W., 2000. Surface and subsurface sedimentary structures produced by salt crusts. Sedimentology 47, 99-118.

Gralla, P., Marsky, J., 2000. Seismic reveals new eastern Precaspian target. Oil & Gas Journal 98, 86-89.

Hardgrove, C., Moersch, J., Whisner, S., 2010. Thermal imaging of sedimentary features on alluvial fans. Planetary and Space Science 58, 482-508.

Harrison, J.C., Jackson, M.P.A., 2013. Exposed evaporate diapirs and minibasins above a canopy in central Sverdrup Basin, Axel Heiberg Island, Arctic Canada. Basin Research, in press. Doi: 10:1111/bre.12037.

Hazel, J.E., 1994. Sedimentary Response to Intrabasinal Salt Tectonism in the Upper Triassic Chinle Formation, Paradox Basin, Utah. US Department of the Interior, U.S Geological Survey Bulletin 2000-F.

Hite, R.J., 1968. Salt Deposits of the Paradox Basin, Southeast Utah and Southwest Colorado. In: Mattox, R.B., Holster, W.T., Ode, H., McIntire, W.L., Short, N.M., Taylor, R.E., Van Siclen, D.C. (Eds), Saline Deposits, Special Papers 88, Geological Society America.

Hartley, A.J., Weissmann, G.S., Nichols, G.J., Warwick, G.L., 2010. Large Distributary Fluvial Systems: Characteristics, Distribution, and Controls on Development. Journal of Sedimentary Research 80, 167-183.

Hodgson, N.A., Farmsworth, J., Fraser, A.J., 1992. Salt-related tectonics, sedimentation and hydrocarbon plays in the Central Graben, North Sea, UKCS. In: Hardman R.F.P. (Ed), Exploration Britain: Geological Insights for the Next Decade. Geological Society Special Publication 67, London, pp. 31-63.

Hudec, M.R., Jackson, M.P.A., 2007. Terra inferma: Understanding salt tectonics. Earth Science Reviews 82, 1-28.

Hudec, M.R., Jackson, M.P.A. Schultz-Ela, D.D., 2009. Paradox of minibasin subsidence into salt: clues to the evolution of crustal basins. Geological Society America Bulletin 121, 201-221.

Ings, S.J., Beaumont, C., 2010. Shortening viscus pressure ridges, a solutution to the enigma of initiating salt ‘withdrawal’ minibasins. Geology 38, 339-342.

Jackson, M.P.A., Talbot, C.J., 1986. External shapes, strain rates, and dynamics of salt structures. Geological Society America Bulletin 97, 305-323.

Kane, I.A., McGee, D.T., Jobe, Z.R., 2012. Halokinetic effects on submarine channel equilibrium profiles and implications for faces architecture: conceptual model illustrated with a case study from Magnolia Field, Gulf of Mexico. Geological Society Special Publication 363, London, pp. 289-302.

Kelling, G., Maldonado, A., Stanley, D.J., 1979. Salt Tectonics and Basement Fractures: Key Controls of Recent Sediment Distribution on the Balearic Rise, Western Mediterranean. Smithsonian Contributions to the Marine Sciences, 3.

Kernen, R.A., Giles, K.A., Rowan, M.G., Lawton, T.F., Hearon, T.E., 2012. Depositional and halokinetic-sequence stratigraphy of the Neoproterozoic Wonoka Formation adjacent to Patawarta allochthonous salt sheet, Central Flinders Ranges, South Australia. In: Alsop, G.I., Archer, S.G., Hartley, A.J., Grant, N.T., Hodgkinson, R. (Eds.), Salt Tectonics, Sediments and Prospectivity. Geological Survey Special Publication, 363, London, pp. 81-105.

Kluth, C.F., Coney, P.J., 1981. Plate tectonics of the Ancestral Rocky Mountains. Geology 9, 10-15.

Kluth, C.F., Du Chene, H.R., 2009. Late Pennsylvanian and early Permian structural geology and tectonic history of the Paradox Basin and Uncompahgre Uplift, Colorado and Utah. In: Houston, W.S., Wray, L.L., Moreland, P.G. (Eds.), The Paradox Basin Revisited – New Developments in Petroleum Systems and Basin Analysis. Rocky Mountain Association of Geologists, Denver, CO, pp. 178-197.

Kneller, B., McCaffrey, W.D., 1995. Modelling the effects of salt-induced topography on deposition from Turbidity Currents. In: Travis, C.J., Vendeville, B.C., Harrison, H., Peel, F.J., Hudec, M.R., Perkins, B.E. (Eds), Salt, Sedimentation, and Hydrocarbons. GCSSEPM Foundation 16th Annual Research Conference, pp. 137-145.

Lawton, T.F., Buck, B.J., 2006. Implications of diapir-derived detrits and gypsic palaeosols in Lower Triassic strata near the Castle Valley salt wall, Paradox Basin, Utah. Geology 34, 885-888.

Lawton, T.F., Vega, F.J., Giles, K.A., Rosales-Dominguez, C., 2001. Stratigraphy and Origin of the La Popa Basin, Nievo Leon and Coahuila, Mexico. In: Bartolin, C., Buffler, R.T., Cantu-Chapa, A. (Eds) The western Gulf of Mexico Basin: Tectonics, sedimentary basins, and petroleum systems. AAPG Memoir 75, 219-240.

Mack, G.H., Rasmussen, K.A., 1984. Alluvial-fan sedimentation of the Cutler Formation (Permo-Pennsylvanian) near Gateway, Colorado. Geological Society of America Bulletin 95. 109-116.

Mange-Rajetzkey, M.A., 1995. Subdivision and correlation of monotonous sandstone sequences using high-resolution heavy mineral analysis, a case study: the Triassic of the Central Graben. In: Dunay, R.E., Hailwood, E.A. (Eds) Non-biostratigraphical Methods of Dating and Correlation. Geological Society Special Publication 89, London, pp. 23-30.

Matthews, W.J., Hampson, G.J., Trudgill, B.D. Underhill, J.R., 2007. Controls on fluvio-lacustrine reservoir distribution and architecture in passive salt diapir provinces: insights from outcrop analogue. American Association of Petroleum Geologists Bulletin 91, 1367-1403.

McKie, T., Jolley, S.J., Kirstensen, M.B., 2010. Stratigraphic and structural compartmentalization of dryland fluvial reservoirs: Triassic Heron Cluster, Central North Sea. In: Jolley, S.J., Fisher, Q.J., Ainsworth, R.B., Vrolijk, P.J., Delisle, S. (Eds) Reservoir

Compartmentalization. Geological Society, London, Special Publication 347, pp. 165-198.

McKie, T., 2011. Architecture and Behavior of Dryland Fluvial Reservoirs, Triassic Skagerrak Formation, Central North Sea. In: Davidson, S.K., Leleu, S., North, C.P. (Eds), From River to Rock Record, SEPM Special Publication 97, pp. 189-214.

Newell, A.J., Tverdokhlebov, V.P., Benton, M.J., 1999. Interplay of tectonics and climate on a transverse fluvial system, Upper Permian, Southern Uralian Foreland Basin, Russia. Sedimentary Geology 127, 11-29.

Newell, A.J., Benton, M.J., Kearsey, T., Taylor, G., Twitchett, R.J., Tverdokhlebov, V.P., 2012. Calcretes, fluviolacustrine sediments and subsidence patterns in Permo-Triassic salt-walled minibasins of the south Urals, Russia. Sedimentology 59. 1659-1676.

Nichols, G.J., Fisher, J.A., 2007. Processes, facies, and architecture of fluvial distributary system deposits. Sedimentary Geology 195, 75-90.

O'Hearn, T., Elliott, S., Samsonov, A., 2003. Karachaganak Field, Northern Pre-Caspian Basin, Northwest Kazakhstan. In: Halbouty, M.T. (Ed), Giant oil and gas fields of the decade 1990-1999. Memoir 78, American Association of Petroleum Geologists, pp. 237-250.

Pairazian, V.V., 1999. A review of the petroleum geochemistry of the Precaspian Basin. Petroleum Geoscience 5, 351-369.

Paz, M., Trudgill, B., Kluth, C., 2009. Salt System Evolution of the Northern Paradox Basin. Search and Discovery Article #30078.

Peacock, D.C.P., 2004. The post-Variscan development of the British Isles within a regional transfer zone influenced by orogenesis. Journal of Structural Geology 26, 2225-22231.

Prather, B.E., Booth, J.R., Steffens, G.S., Graig, P.A., 1998. Classification, Lithologic Calibration, and Stratigraphic Succession of Seismic Facies of Intraslope Basins, Deep-Water Gulf of Mexico. American Association of Petroleum Geologists Bulletin 82, 701-728.

Prather, B.E. 2000. Calibration and visualization of depositional process models for above-grade slopes: a case study from the Gulf of Mexico. Marine and Petroleum Geology, 17, 619-638.

Prochnow, S.J., Atchley, S.C., Boucher, T.E., Nordt, L.C., Hudec, M.R., 2006. Influence of salt withdrawal subsidence on palaeosol maturity and cyclic fluvial deposition in the Upper Triassic Chinle Formation: Castle Valley Utah. Sedimentology, 53, 1319-1345.

Prommel, H.W.C., 1923. Geology and structure of portions of Grand and San Juan counties, Utah. American Association of Petroleum Geologists Bulletin 4, 384-399.

Rahn, P.H., 1967. Sheetfloods, Streamfloods, and the formation of Pediments. American Association of Petroleum Geologists Bulletin 57, 593-604.

Rowan, M.G., Lawton, T.F., Giles, K.A., 2012. Anatomy of an exposed vertical salt weld and flanking strata, La Popa Basin, Mexico. In: Alsop, G.I., Archer, S.G., Hartley, A.J., Grant, N.T., Hodgkinson, R. (Eds.) Salt Tectonics, Sediments and Prospectivity. Geological Society, London, Special Publication 363, pp. 33-57.

Schamel, S., Pilifossov, V.M., Votsalevsky, E.S., 1995. Geological controls on style of salt piercements in the Pricaspian Basin, Kazakhstan: contrasts with the Gulf of Mexico. In: Travis, C.J., Vendeville, B.C., Harrison, H., Peel, F.J., Hudec, M.R., Perkins, B.E. (Eds) Salt, Sedimentation, and Hydrocarbons. GCSSEPM Foundation 16th Annual Research Conference, pp. 253-255.

Senseny, P.E., Hansen, F.D., Russell, J.E., Carter, N.L., Handin, J.W., 1992. Mechanical Behaviour of Rock Salt: Phenomenology and Micromechanisms. International Journal of Rock Mechanics and Mining Sciences & Geomechanics Abstracts 29, 363-378.

Shoemaker, E.M., Newman, W.L. 1957. Moenkopi Formation (Triassic? and Triassic) in Salt Anticline Region, Colorado and Utah. American Association of Petroleum Geologist Bulletin 42, 775-789.

Smith, R.I., Hodgson, N., Fulton, M.M., 1993. Salt control on Triassic reservoir distribution, UKCS Central North Sea. In: Parker, J.R. (Ed.), Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference, Geological Society, London, pp. 547-557.

Smith, D.B., Taylor J.C.M., 1992. Permian. In: Cope, J.C.W., Ingham, J.K., Rawson, P.F. (Eds) Atlas of Palaeogeography and Lithofacies. Geological Society, London, Memoir 13, pp. 87-95.

Soreghan, G.S., Soreghan M.J., Sweet, D.E., More, K.D., 2009. Hot fan or cold outwash? Hypothesized proglacial deposition in the upper Paleozoic Cutler Formation, western tropical Pangea. Journal of Sedimentary Research 79, 495-522.

Srivastava, K.K., Merchant, H.D., 1973. Thermal expansion of alkali halides above 300°K. Journal of Physics and Chemistry of Solids 34, 2069-2073.

Stewart, J.H., Poole, F.G., Wilson, R.F., Cadigan, R.A., 1972. Stratigraphy and origin of the Triassic Moenkopi Formation and related strata in the Colorado Plateau region. US Department of the Interior, US Geological Survey Professional Paper 691.

Stewart, S.A., Clarke, J.A., 1999. Impact of salt on the structure of the Central North Sea hydrocarbon fairways. Petroleum Geology Conference Series 5, 179-200.

Trudgill, B.D., 2011. Evolution of salt structures in the northern Paradox Basin: controls on evaporite deposition, salt wall growth and supra-salt stratigraphic architecture. Basin Research 23, 208-238.

Trusheim, F., 1960. Mechanism of salt migration in northern Germany. American Association of Petroleum Geologists Bulletin 44, 1519-1540.

Venus, J.H., 2013. Tectono-stratigraphic evolution of fluvial and aeolian systems in a salt mini-basin province during changing climatic conditions: Permian Undifferentiated Cutler Group, South East Utah, USA. Unpublished PhD Thesis, University of Leeds, UK.

Volozh, Y., Talbot, C., Ismail-Zadeh, A., 2003. Salt structures and hydrocarbons in the Pricaspian Basin. American Association of Petroleum Geologists 87, 313-334.

Waldron, J.W.F., Rygel, M.C., 2010. Evaporite Tectonics, Transpression and Transtension in the development of the Maritimes Basin, Atlantic Canada. Salt Tectonics, Sediments and Prospectivity, GSP-SEPM Conference Abstract, 20-22 January 2010, 16.

Weissmann, G., Hartley, A.J., Nichols, G.J., Scuderi, L.A., Olsen, M., Buehler, H., Banteah, R., 2010. Fluvial form in modern continental basins: Distributive fluvial systems. Geology 38, 39-42.

Werner, W.G., 1974. Petrology of Culter Formation (Pennsylvanian-Permian) near Gateway, Colorado, and Fisher Towers, Utah. Journal of Sedimentary Petrology, 44 292-298.

Williams, G.E., 1970. The Central Australian Stream Floods of Febuary-March 1967. Journal of Hydrology 11, 185-200.

Ziegler, P.A., 1975. Geologic Evolution of the North Sea and its Tectonic Framework. American Association Petroleum Geologists Bulletin 59, 1073-1097.

## *Captions*

**Fig. 1:** Overview of halokinetic provinces world-wide. Light grey indicates halokinetic province not covered in this study. Dark grey denotes province mentioned in this study. G: German case studies; LP: La Popa Basin; NB: New Brunswick; NS: North Sea; PC: Precaspian Basin; Px: Paradox Basin. Modified after Hudec and Jackson (2007).

**Fig. 2:** Description of basin-fill attributes defining basin-fill thickness, fill style, pre-existing basin fill, and remaining subsidence potential of the basin. These parameters can vary both between mini-basins and within a single mini-basin. **T = Basin-fill thickness**, which can vary within a single basin, e.g., features such as turtle-back structures & rim synclines. **F = Fill inheritance**, which records the state of basin-fill at the onset of a subsequent episode of deposition, and which can vary spatially across a mini-basin due to variations in differential subsidence rate or existing basin fill-thickness. **M = Maximum basin-fill level** (fill potential) is determined by the original thickness of salt and can vary due to the presence of a dipping basement or the presence of pre-salt basement structures. **P = Remnant basin-fill potential**, describes the salt remaining beneath an evolving mini-basin and can vary across a basin due to differential subsidence or due to sub-salt basement geometries. **S = Basin-fill style** is a general concept describing the overall nature of the sediment fill (e.g. sand-prone or sand-poor). Sh = horizontal fill style; Sv = vertical fill style. **U = Available accommodation** (space remaining unfilled) and can be negative if the basin fill becomes elevated above a “baseline of erosion”. **W= Salt-wall height** above “regional” elevation.

**Fig. 3:** Common mechanisms for the initiation of halokinesis. (a) Buoyancy-driven halokinesis, where density of the overburden initiates and drives halokinesis. (b) Differential loading, where halokinesis is driven by varying thickness and density of overburden, created by features such as accumulation of a prograding alluvial wedge. (c) Initiation of halokinesis due to extension, where thin-skinned tectonics creates differential thicknesses of salt. Modified after Jackson and Talbot (1986).

**Fig. 4:** Model demonstrating the evolution of salt-walled mini-basins throughout various stages of evolution. Modified after Hodgson et al (1992).

**Fig. 5:** Common salt wall and mini-basin geometries. Geometries of developing salt walls are controlled by factors, including: the trend of pre-existing basement faults and topography; the direction of tectonic

extension or shortening; the direction of differential loading. Where a single basement trend exists, linear salt walls tend to develop and these separate elongate, linear mini-basins. Where multiple basement trends exist or where tectonism occurs in an orientation that is oblique to the trend of pre-existing basement structures, polygonal walled mini-basins tend to develop.

**Fig. 6:** Fluvial interactions with uplifted salt walls and subsiding mini-basins. Sediments can be delivered transversely or axially into the mini-basins and this may act to dictate the ensuing basin-fill style. The development of polygonal salt walls can further add to stratigraphic complexity.

**Fig. 7:** Basin-fill style, demonstrating the development of over-filled, filled, and under-filled basin styles. The model depicts a system subject to a constant rate of sediment supply rate, but for adjacent mini-basins that undertake variable rates of subsidence.

**Fig. 8:** Schematic depiction of the action of a suite of controlling parameters to dictate the geometry and style of infill of salt-walled mini-basins. These factors govern rates of sediment accumulation and basin subsidence. Modified after Banham and Mountney (2013b).

**Fig. 9:** Salt Anticline Region, southeastern Utah. (a) Overview of map and stratigraphy of the Paradox Basin (b) General depositional model depicting the style of fluvial accumulation of the Permian Cutler Group, where transverse delivery of sediment from the Uncompahgre Uplift into the foredeep of the Paradox Basin resulted in preferential filling of basins closest to the Uncompahgre Uplift prior to overspill into more distal mini-basins. (c) General depositional model depicting the style of fluvial accumulation in the Lower Triassic Moenkopi Formation and showing the development of adjacent sand-prone and sand-poor basins of similar ages. Modified after Banham and Mountney (2013b). (d) General depositional model depicting the style of fluvial accumulation in the Upper Triassic Chinle Formation, where episodic uplift of salt walls influenced sediment accumulation across the region. In addition, climate variation during influenced fluvial style and preserved stratigraphic expression.

**Fig. 10:** Pre-Caspian Basin, Kazakhstan. (a) Overview map and stratigraphic column (After Barde *et al.*, 2002a, b). (b) General depositional model depicting fluvial sediment accumulation in a series of polygonal salt-walled mini-basins in the Pre-Caspian Basin. Basin-fill style is highly variable: drainage pathways can become entrenched in some mini-basins, thereby preventing sediment delivery into neighbouring basins. Sediment-starved basins tend to be characterised by evaporitic or lacustrine sedimentation. Modified after Barde et al (2002b).

**Fig. 11:** Central Graben, North Sea. (a) Overview map of Central Graben & stratigraphic column (Modified after Goldsmith et al., 1995; Glennie et al, 2003; Goldsmith et al., 2003). (b) General depositional model depicting fluvial sediment accumulation during the Middle and Late Triassic in the Central Graben of the North Sea. Drainage pathways may bifurcate or become isolated as a result of salt-wall uplift, which controls the developing stratigraphic succession. Differential subsidence rates may lead to the development of lacustrine intervals within the succession. Modified after Hodgson et al. (1992).

**Fig. 12:** La Popa Basin, Mexico. (a) Overview maps of La Popa Basin and stratigraphic column (After Dickinson and Lawton, 2001; Andrie et al., 2012). For more accurate maps, see Andrie et al., 2012. (b) General depositional model depicting fluvial sediment accumulation in the Carroza Formation, La Popa Basin, Mexico. Development of a rim-syncline adjacent to La Popa salt wall controlled the location of drainage pathways throughout the latter stages of deposition of the Carroza Formation. Modified after Andrie et al. (2012).

**Fig. 13:** Conceptual model demonstrating how the interplay between rates of subsidence and sediment supply act to control the developing basin-fill style. Balanced rates of subsidence and sediment supply generate filled (sand-prone) basins; in cases where the rate of subsidence outpaces the rate of sediment supply, under-filled basins with argillaceous fills tend to develop and available accommodation may remain partially unfilled; in cases where the rate of sediment supply outpaces the rate of subsidence, over-filled basins with gravel-prone fills tend to develop and accommodation is filled leading to downstream bypassing of excess sediment.

**Fig. 14:** Models depicting the evolution of linear salt-walled mini-basins. See text for further explanation.

**Fig. 15:** Models depicting the evolution of polygonal salt-walled mini-basins. See text for further explanation.

**Figure 1 SWMB overview**

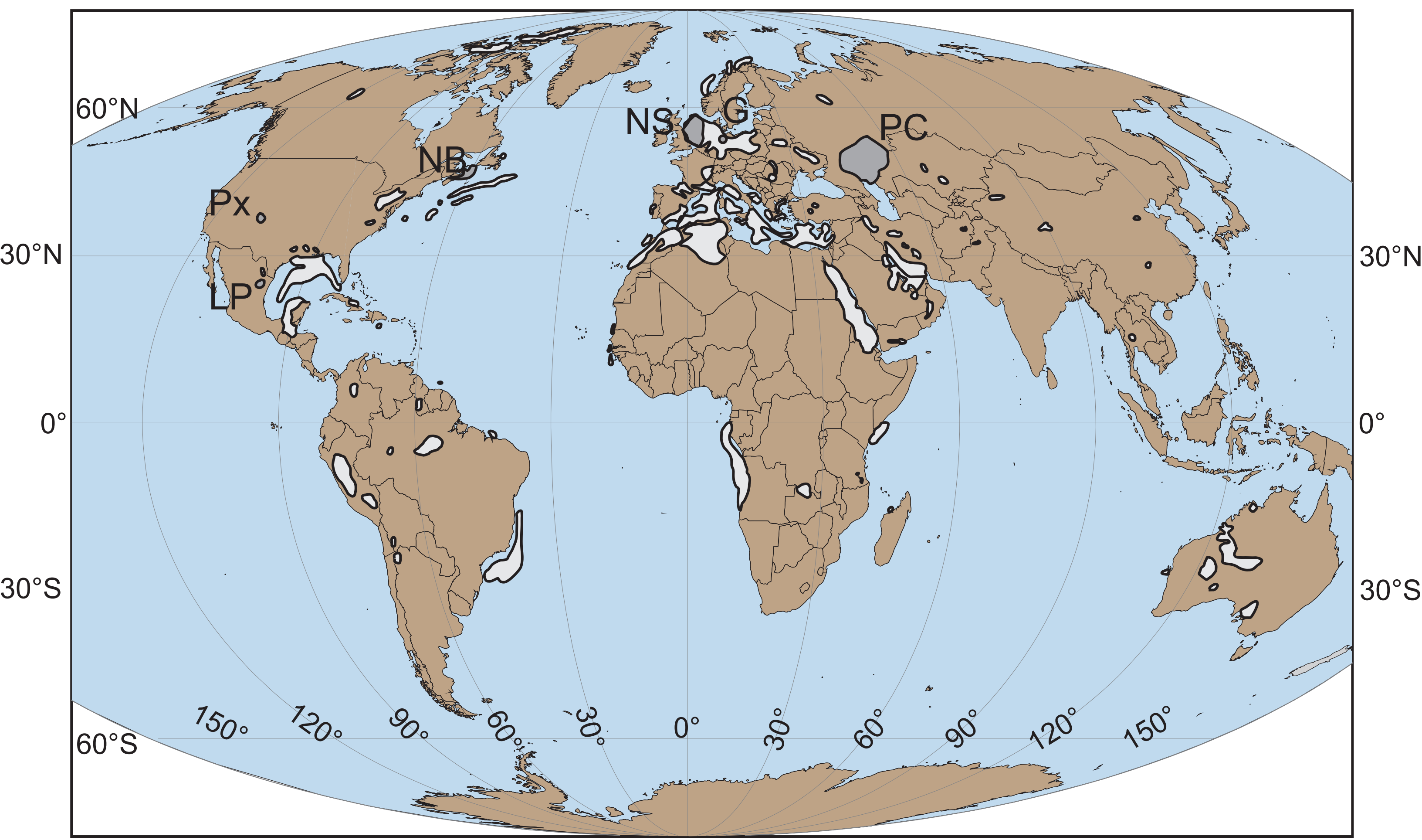


Figure 2 Basin fill atributes

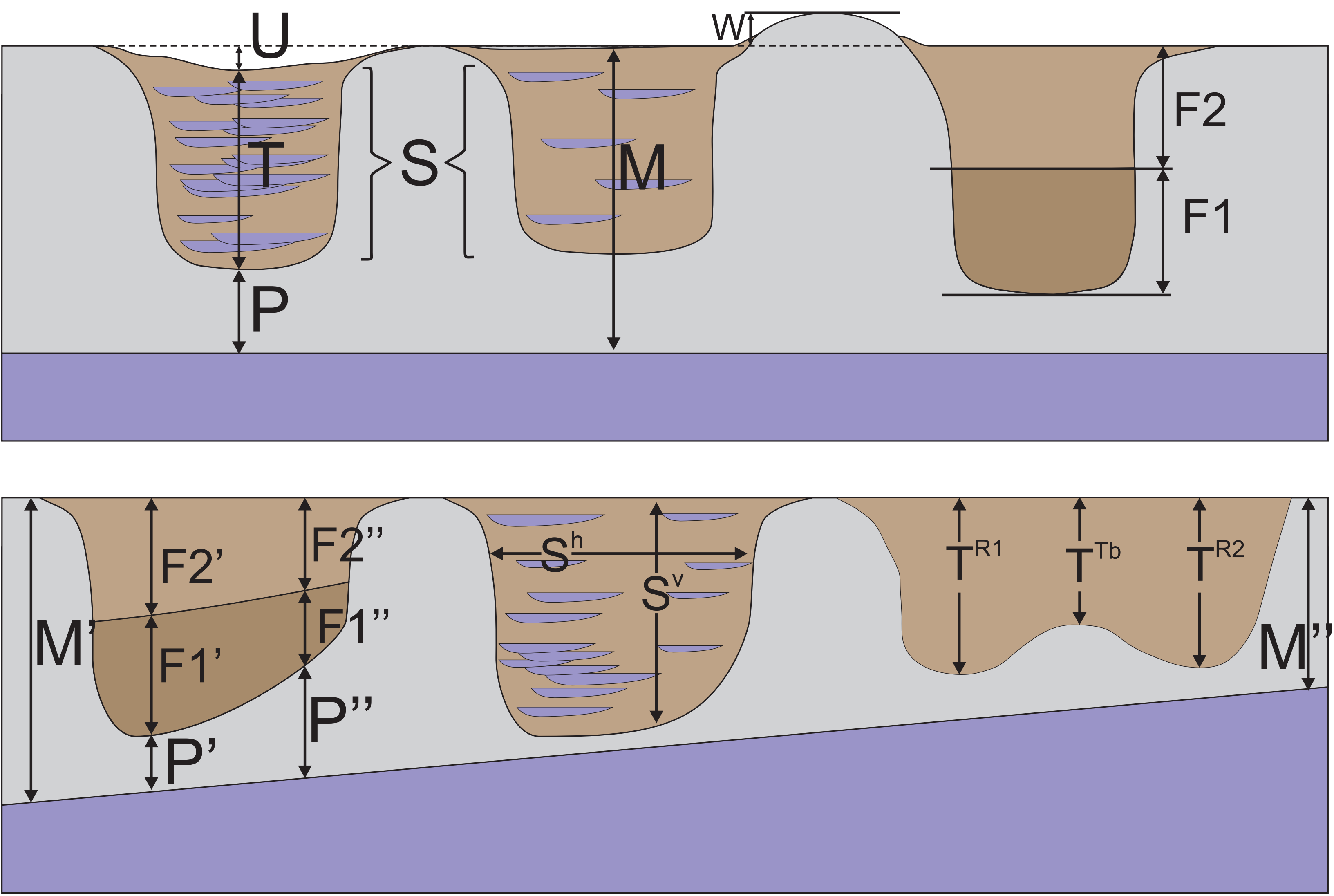


Figure 3 Mechanisms

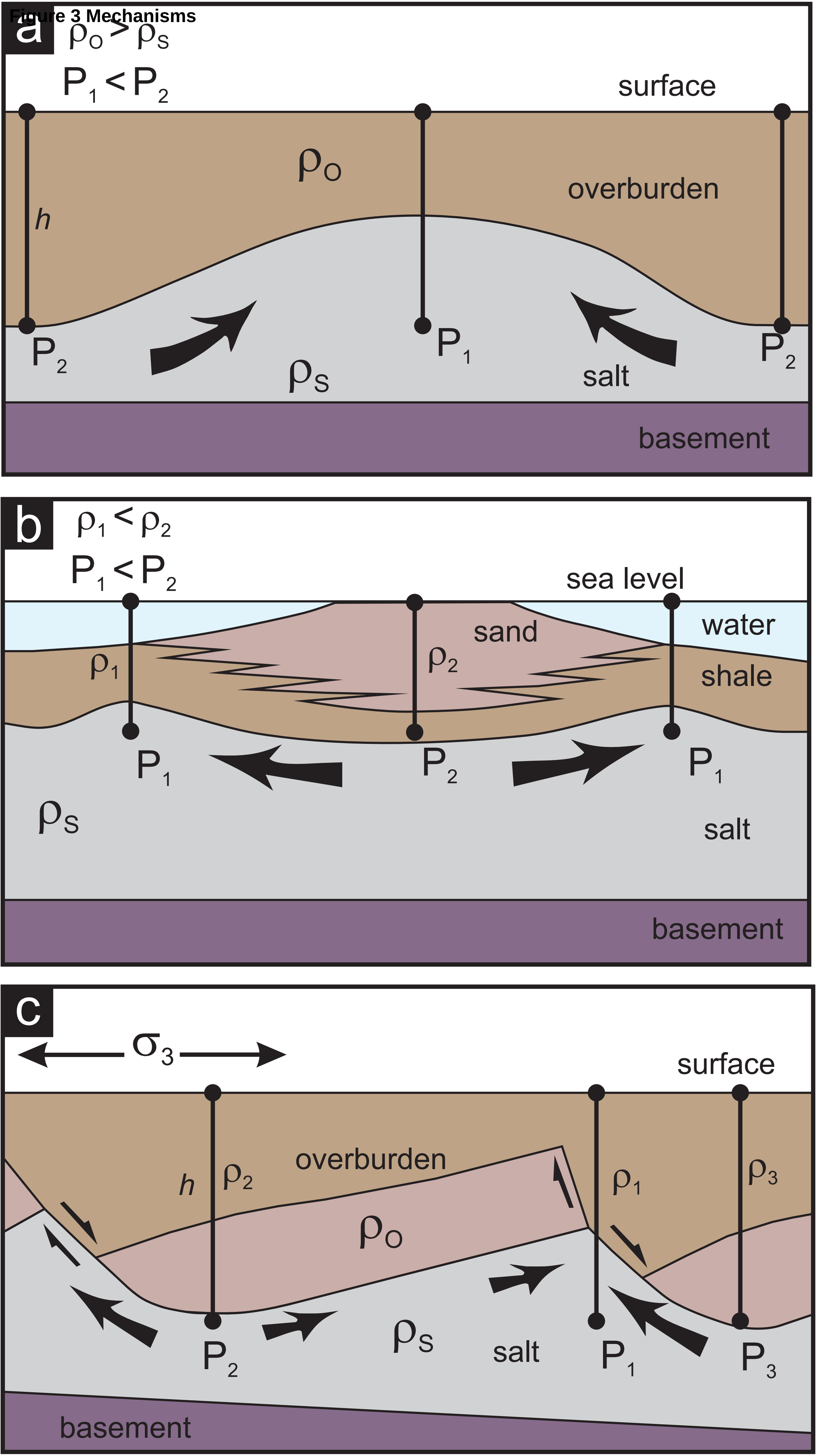


**Figure 4 Podology Model**

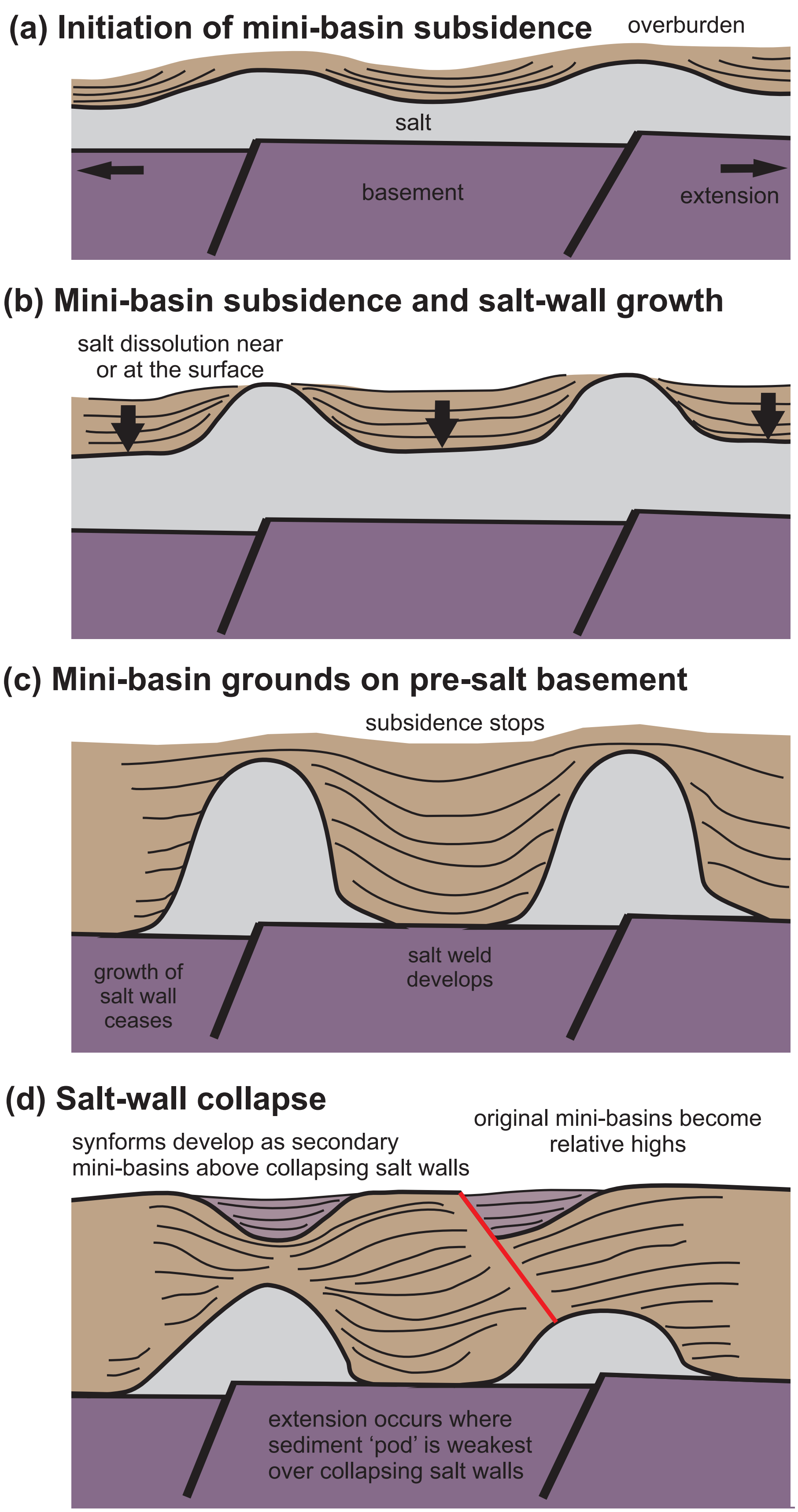


**Figure 5 SW geometries**

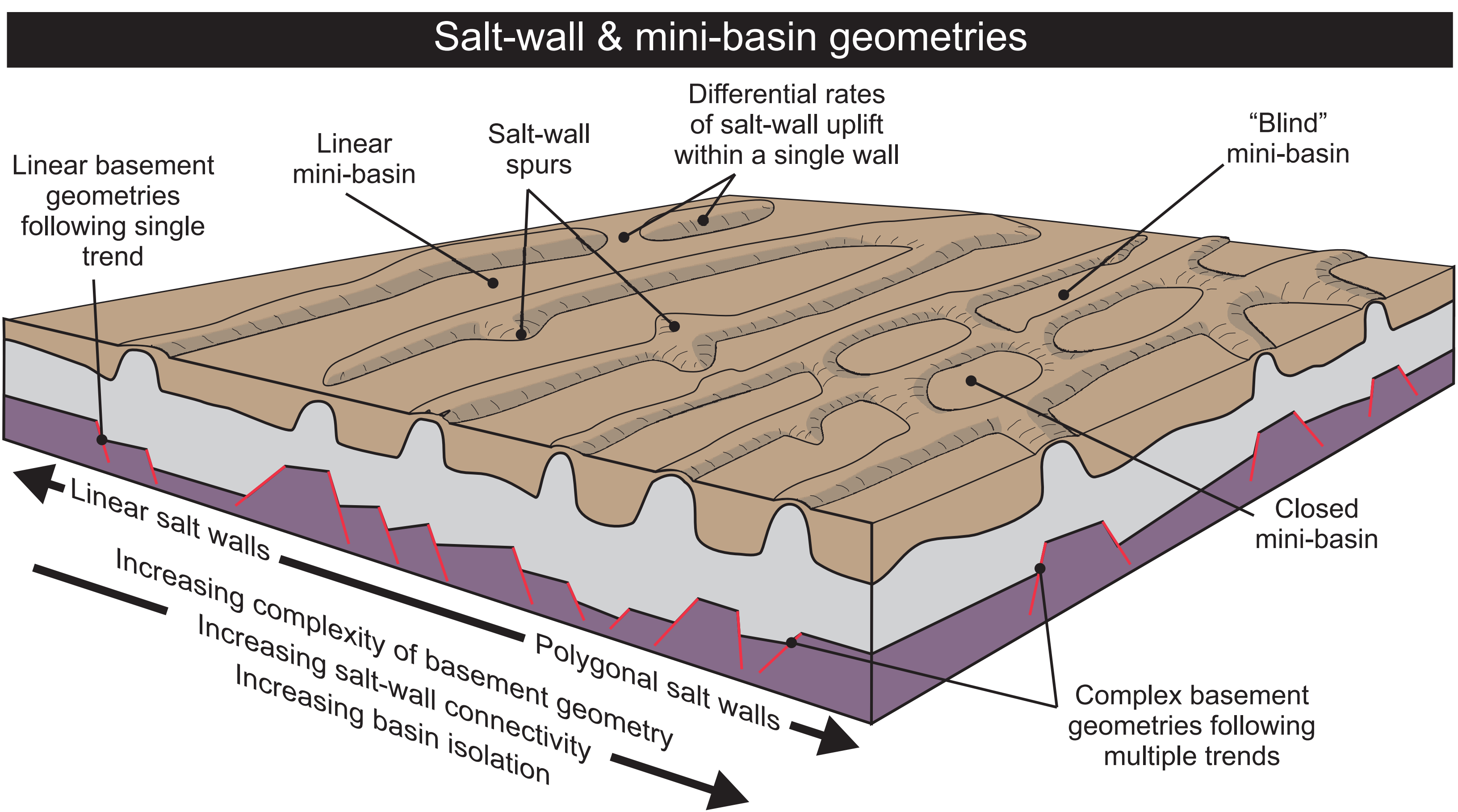


**Figure 6 Drainage pathways**

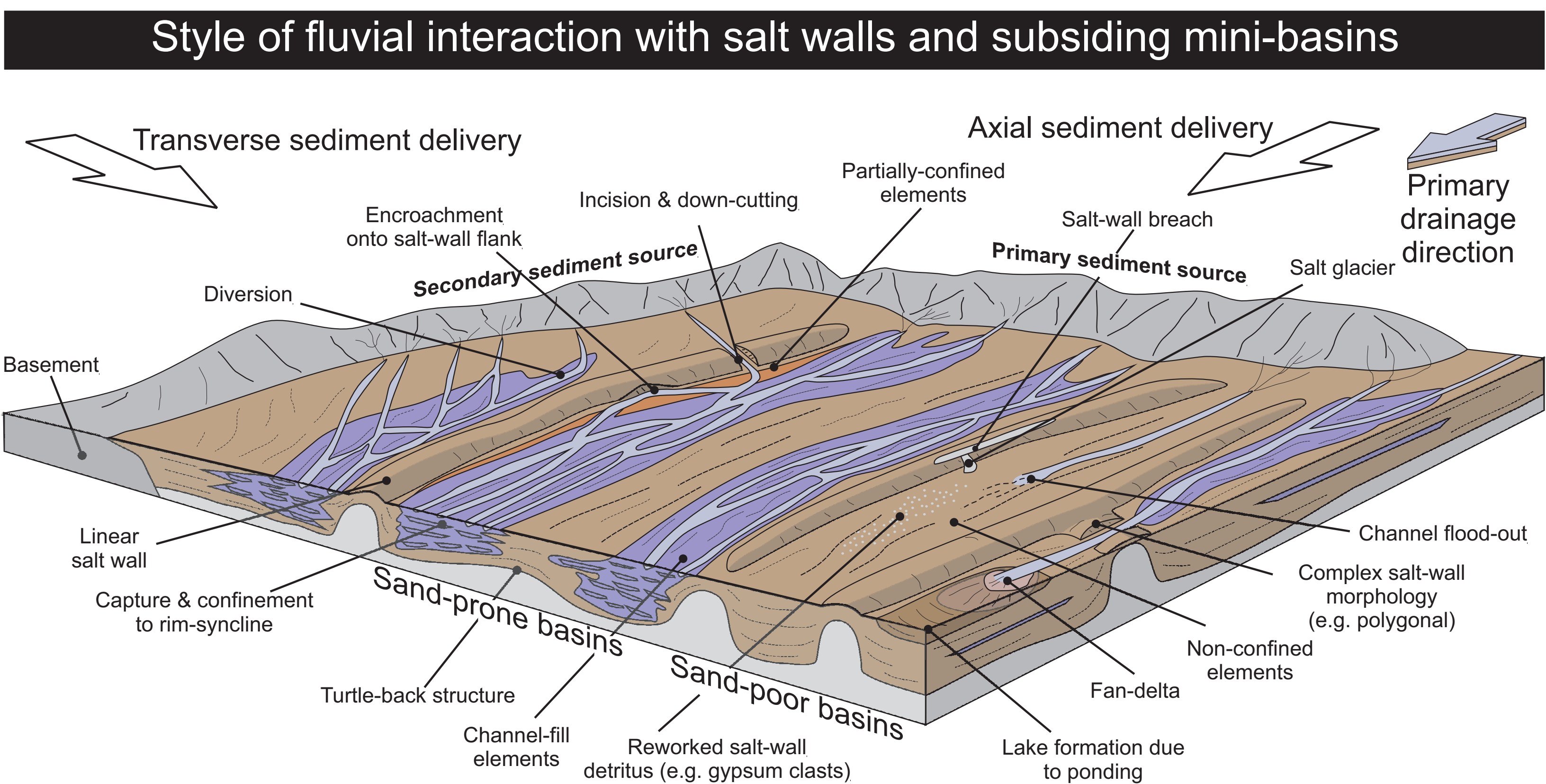


**Figure 7 Fill states**

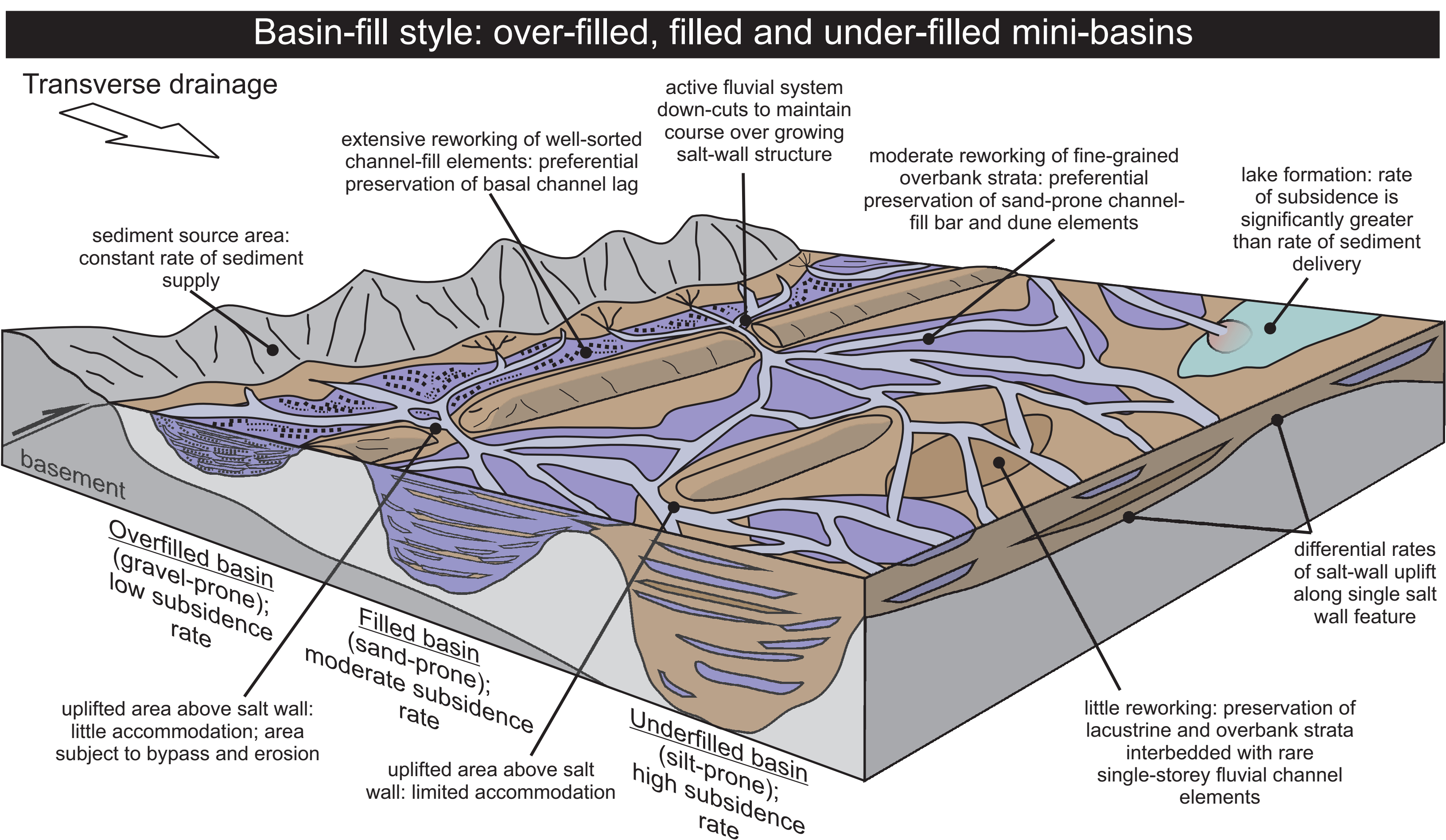


**Figure 8 Interactions**

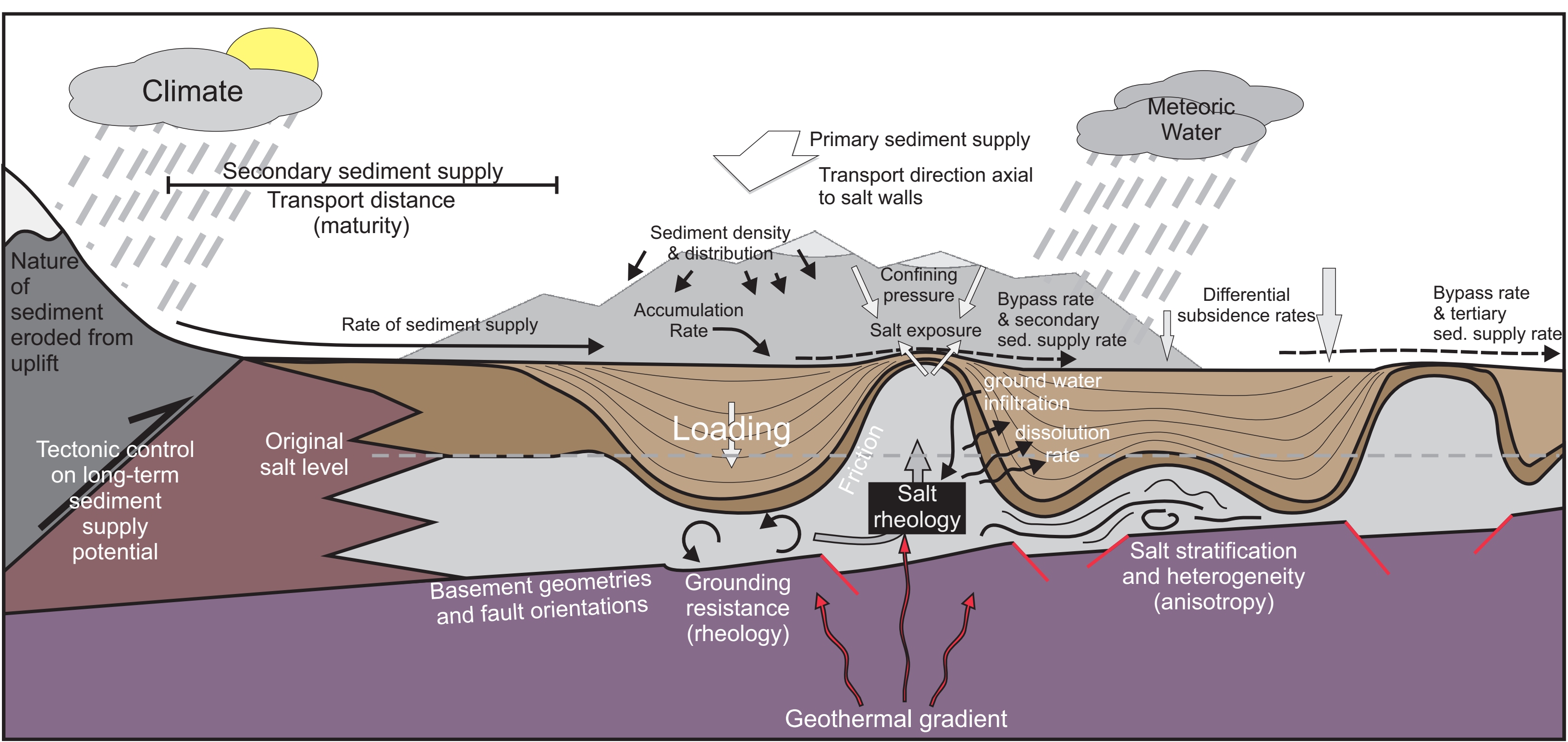


**Figure 9a Overview map**

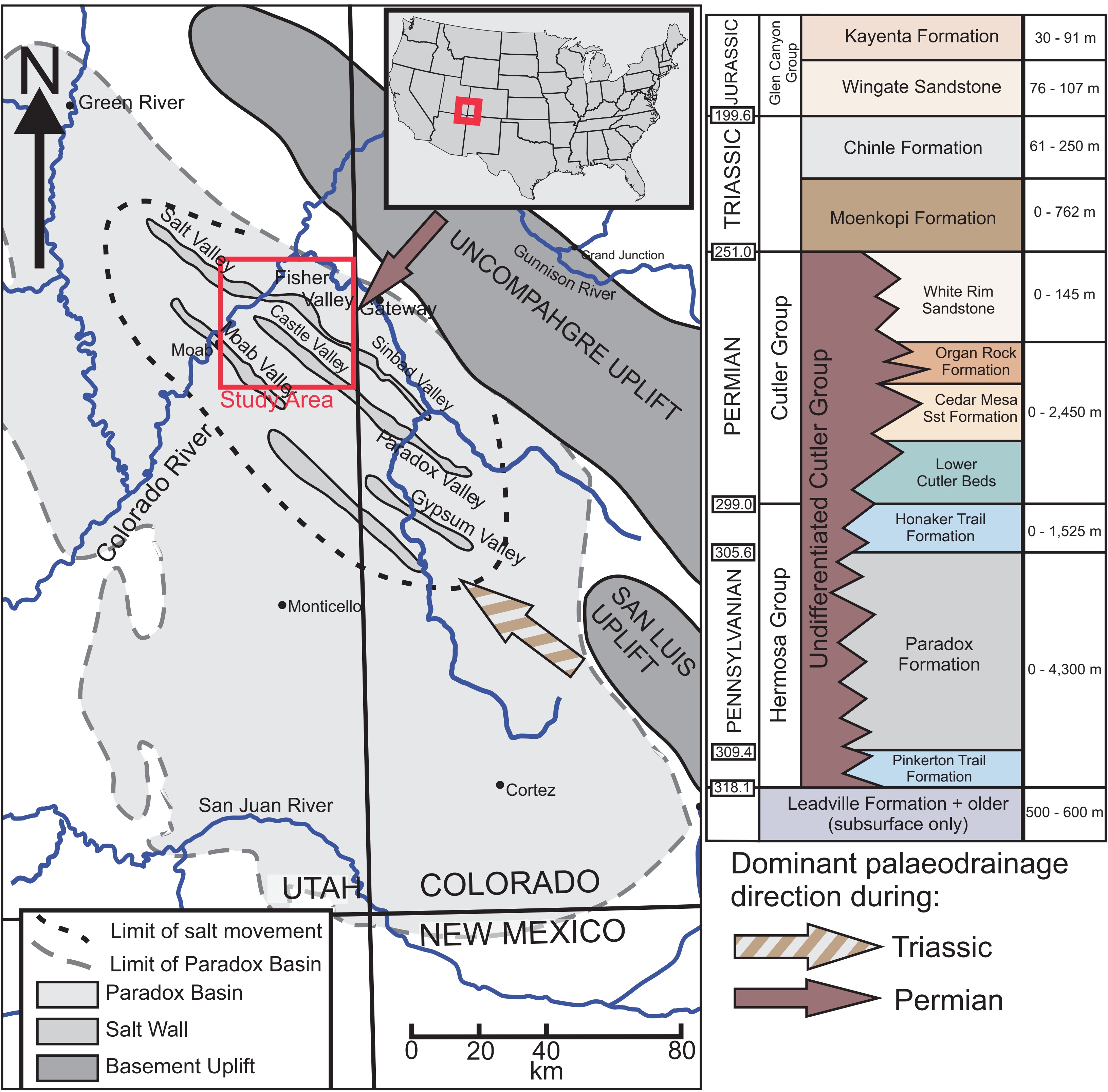
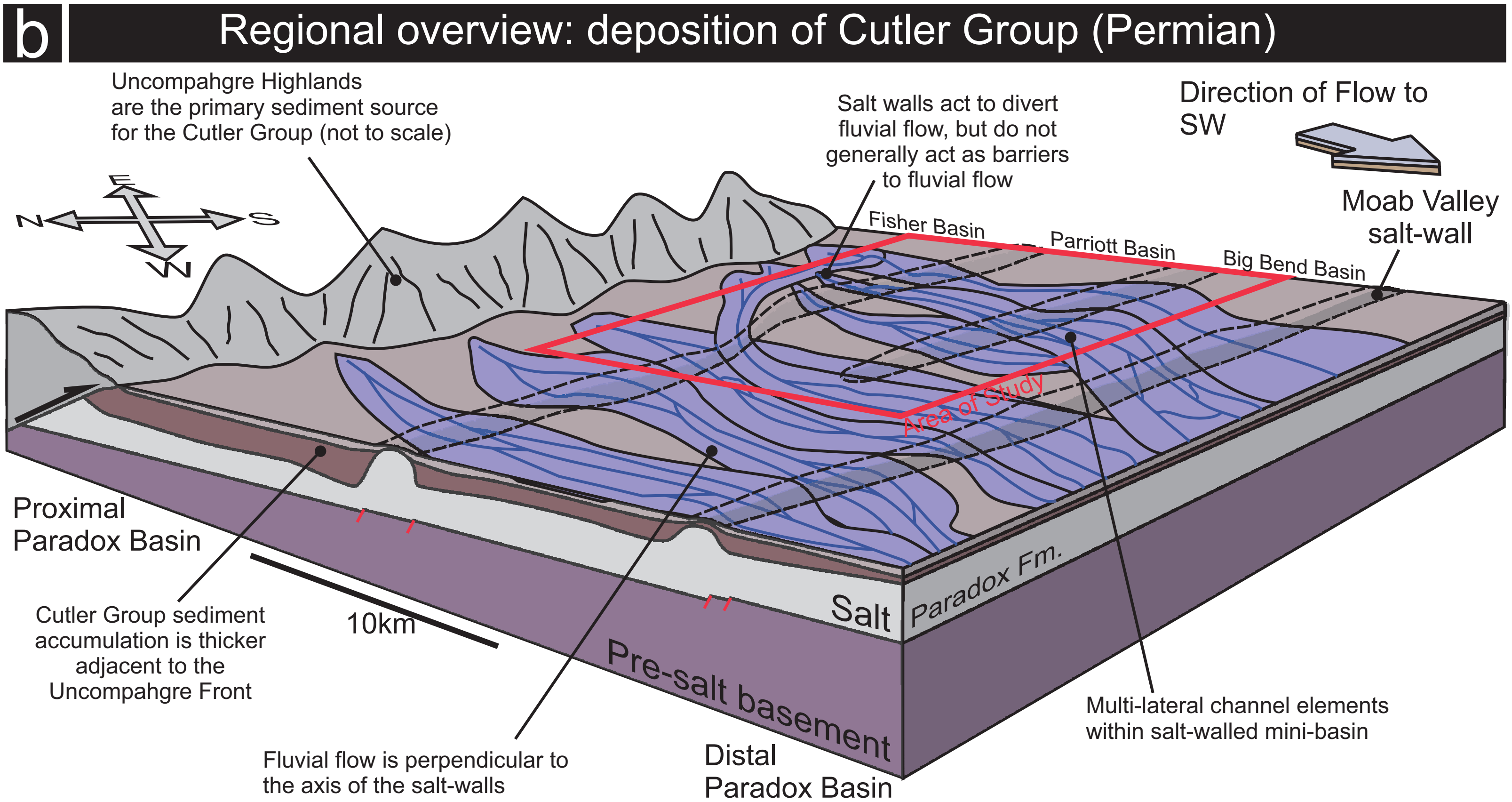


**Figure 9bc Cutler Moenkopi**



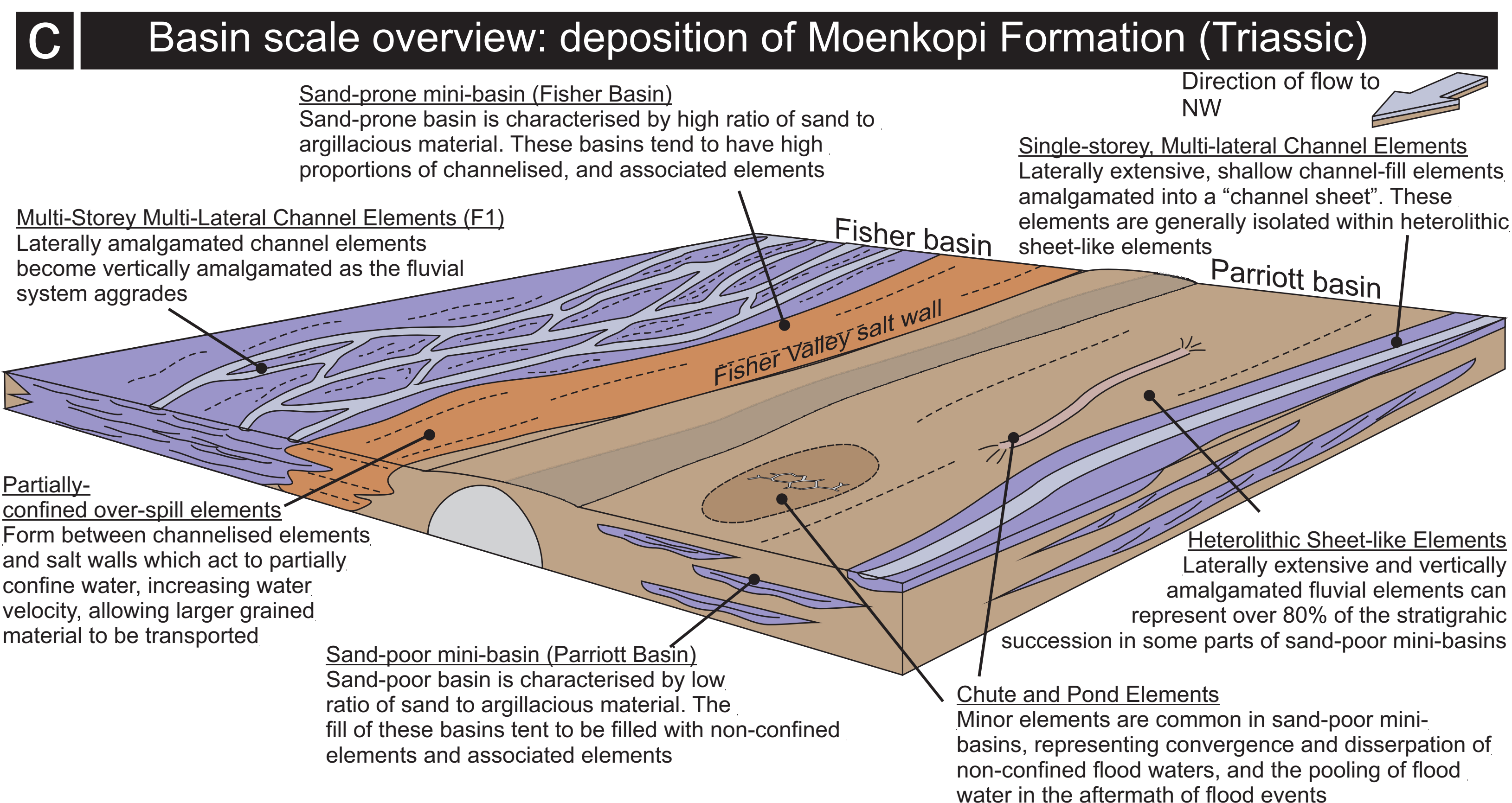



**Figure 9d Chinle**

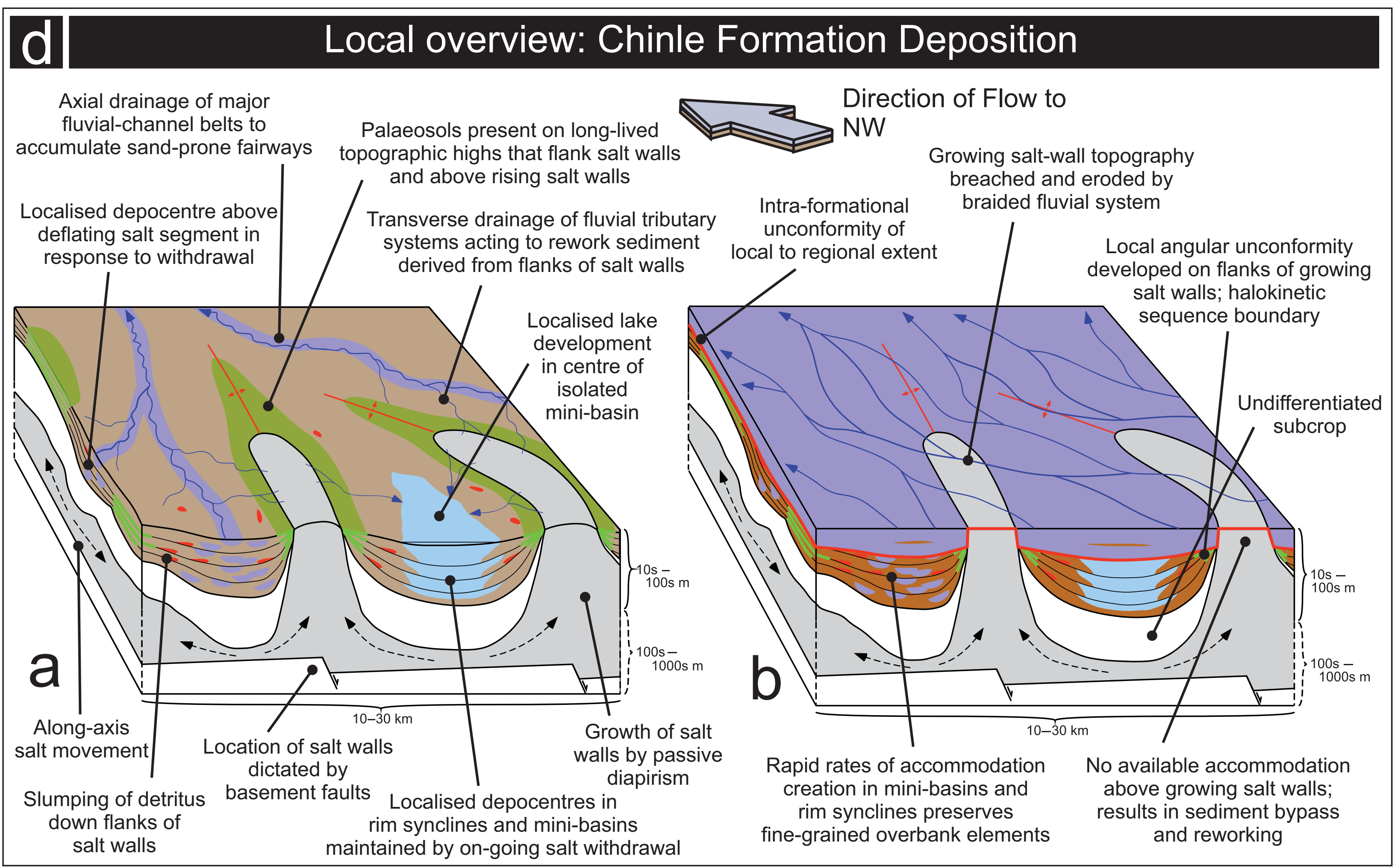


**Figure 10 a Pre Caspian Overview**

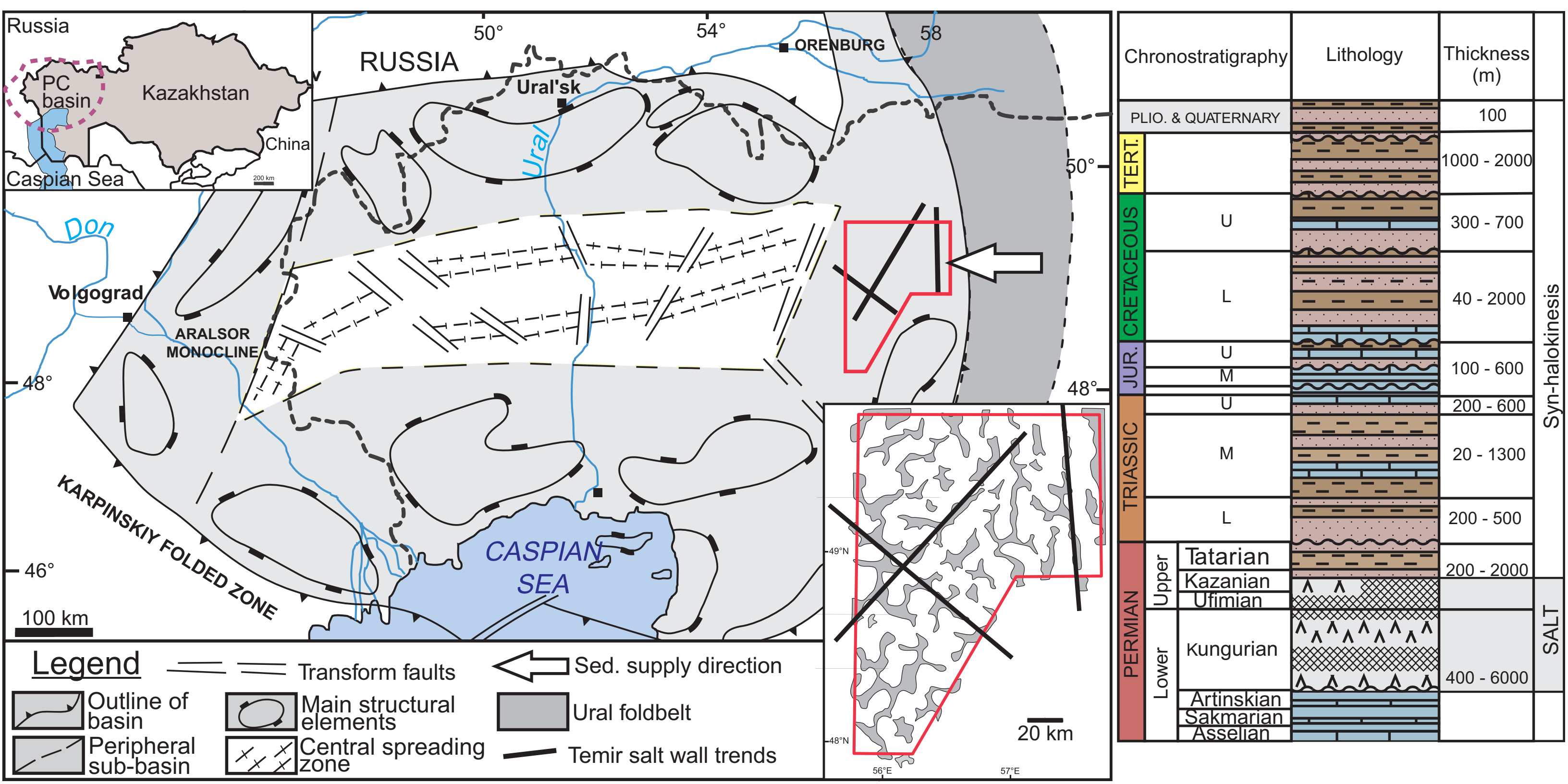


**Figure 10 b PCaspian model**

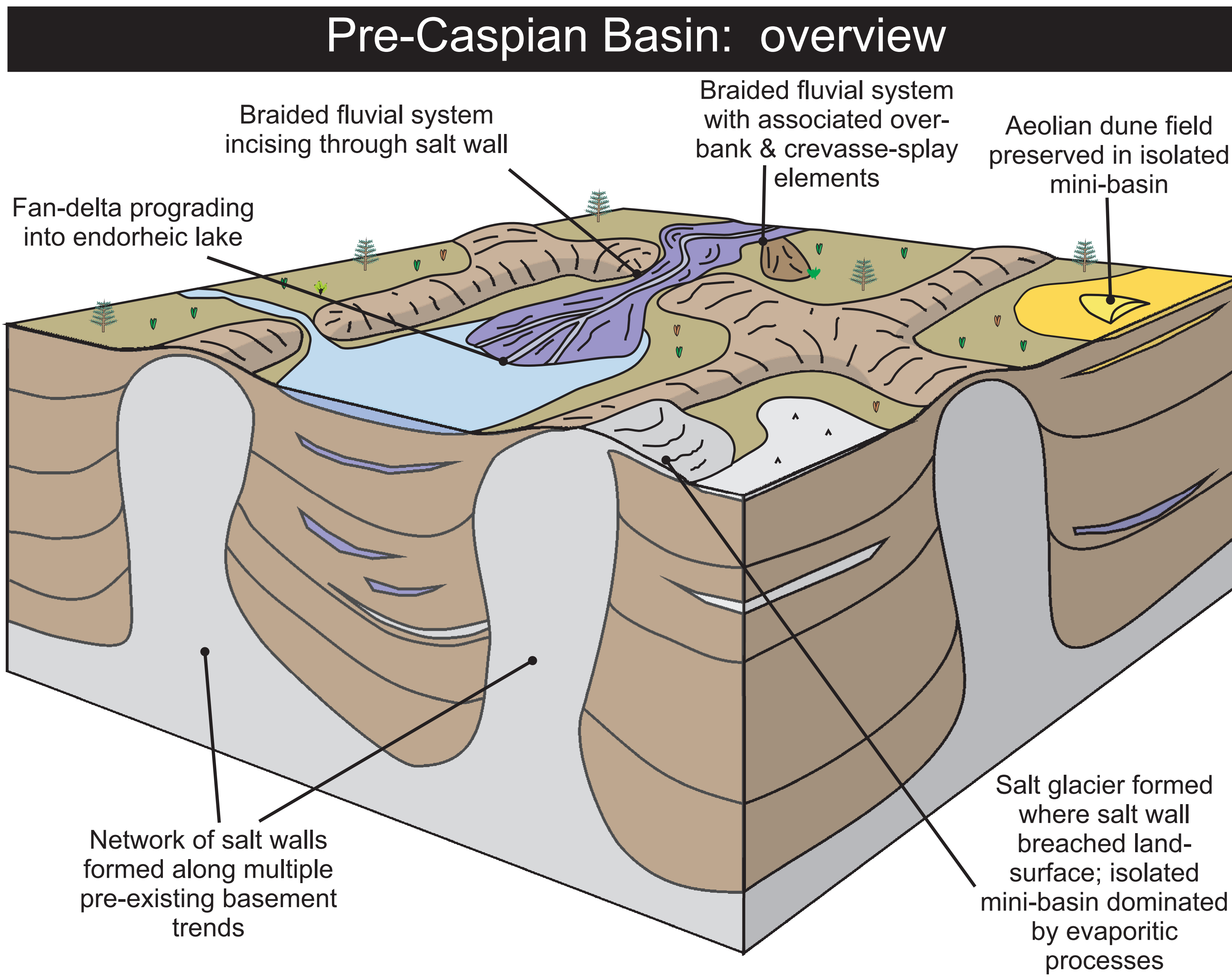


**Figure 11 a North Sea Map**

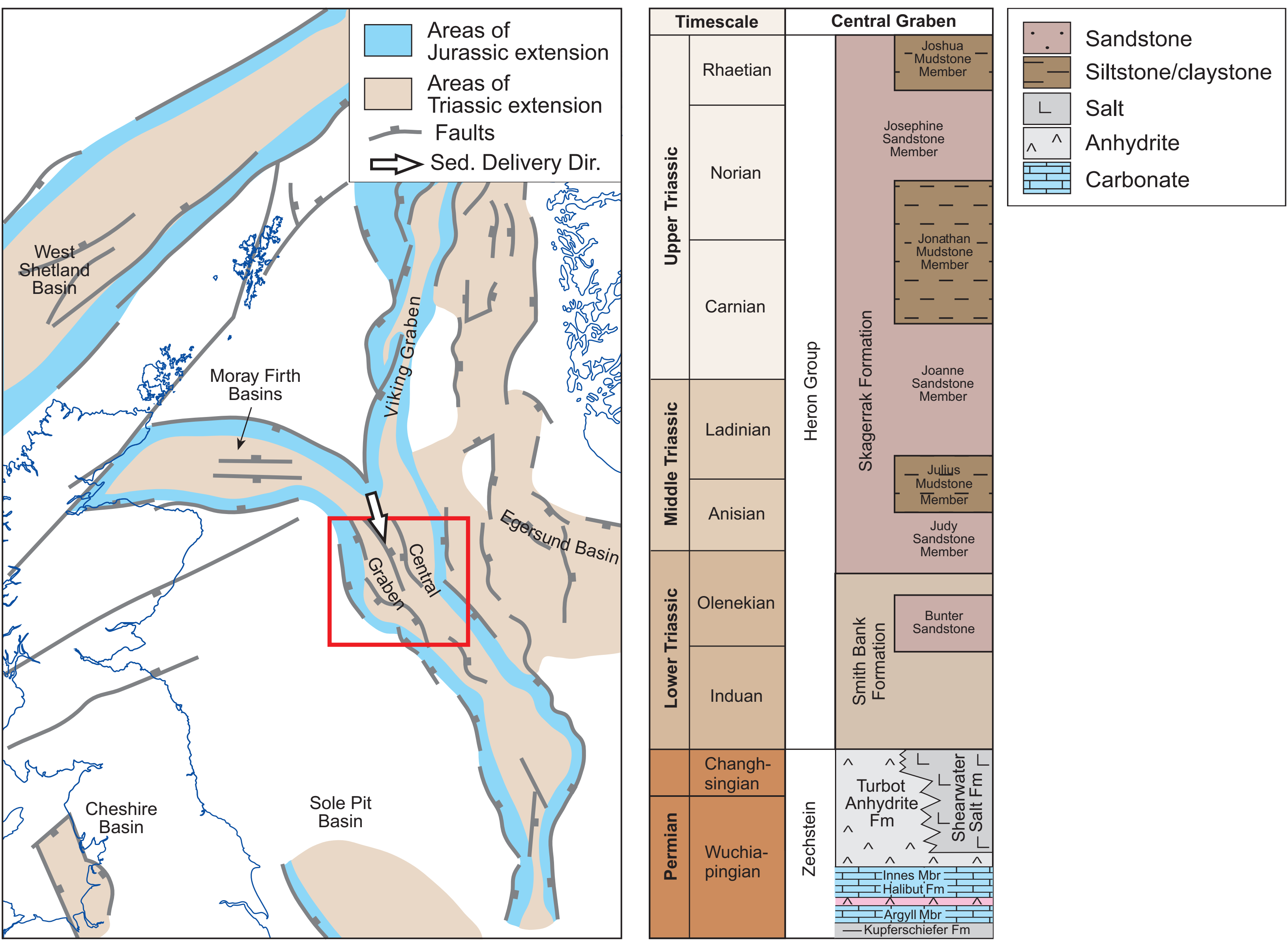


Figure 11 b NS Mini basins

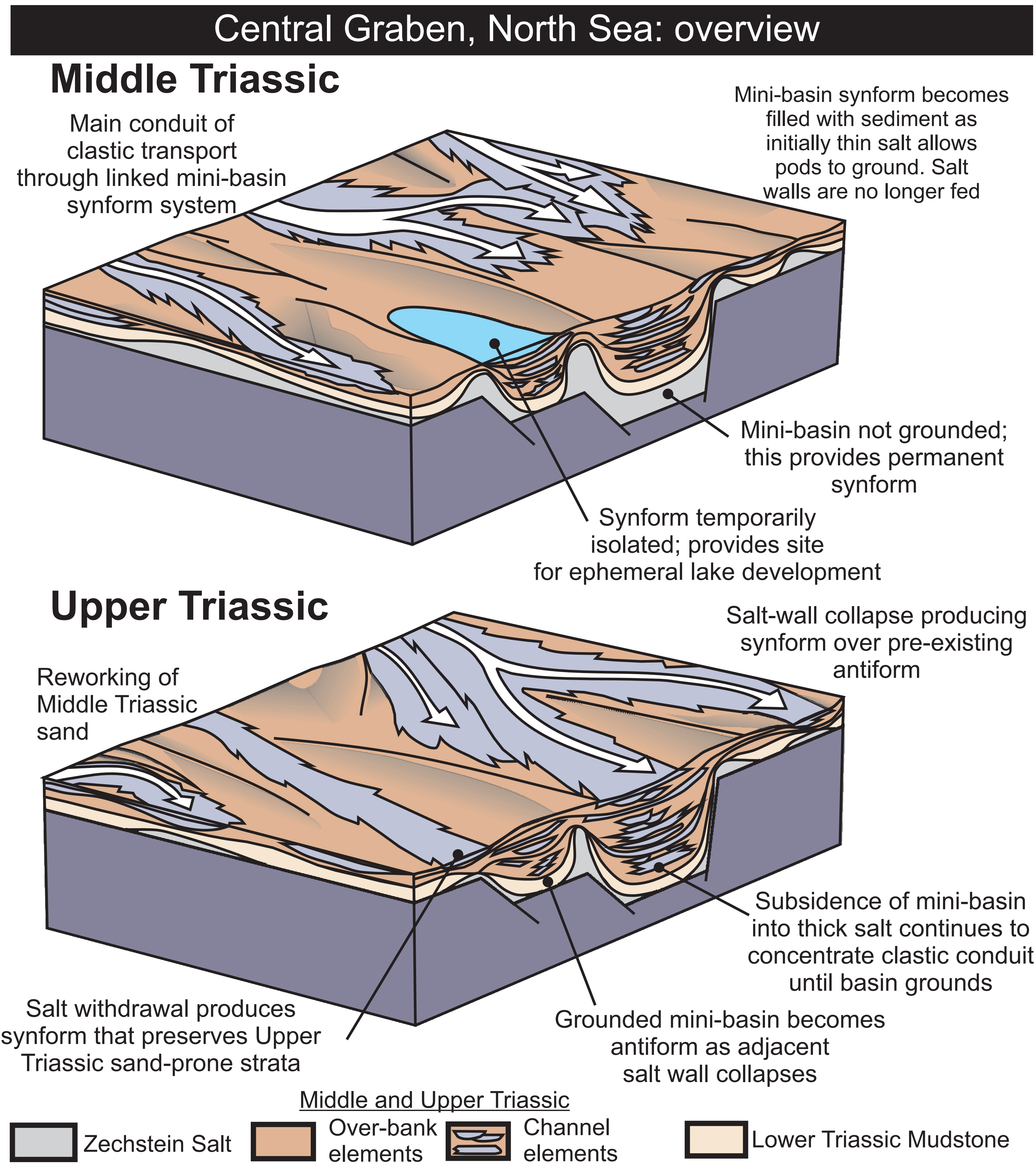


**Figure 12a La Popa Map**

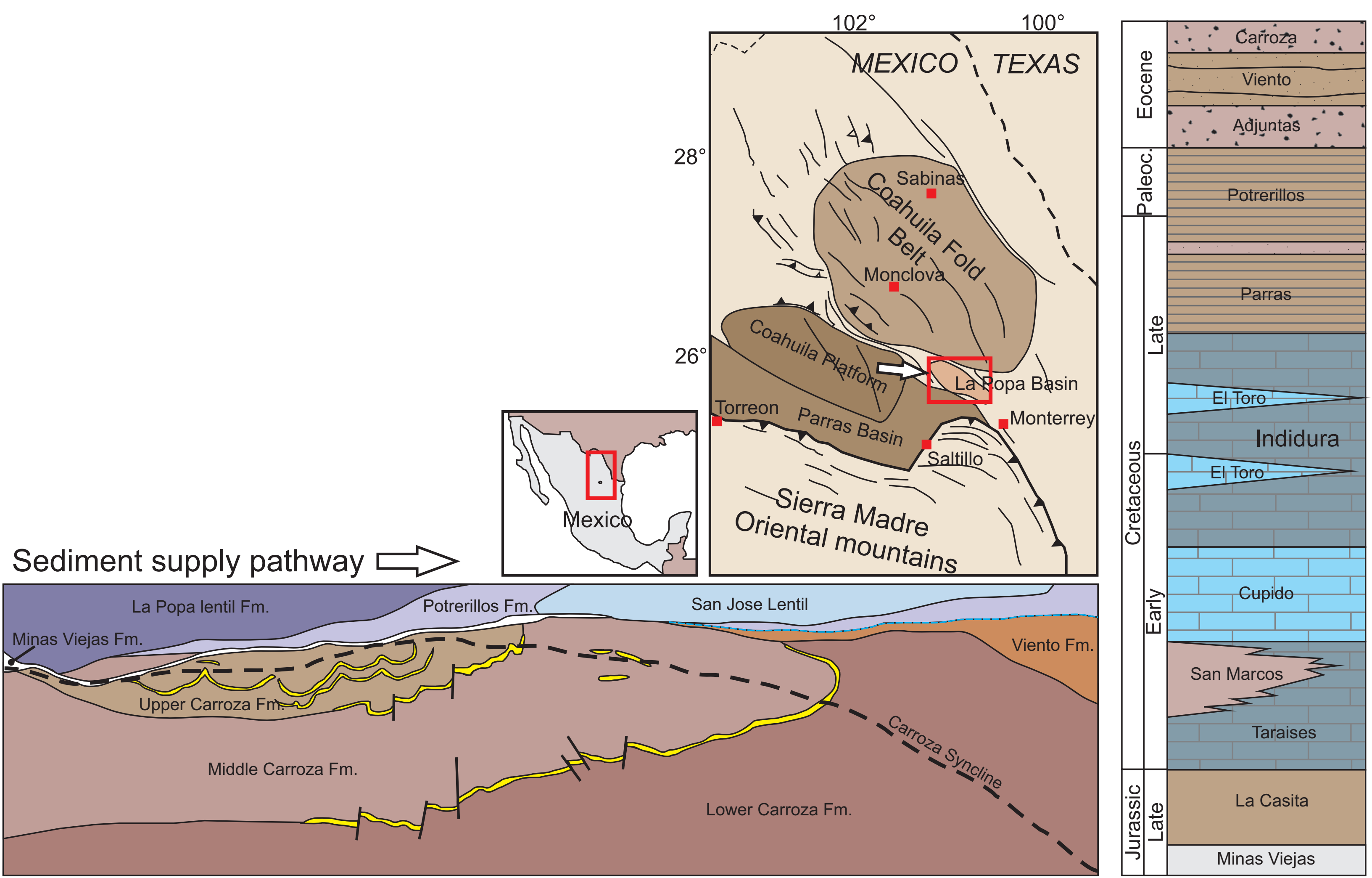


Figure 12b La Popa model

# La Popa Basin: Carroza Fm. evolution

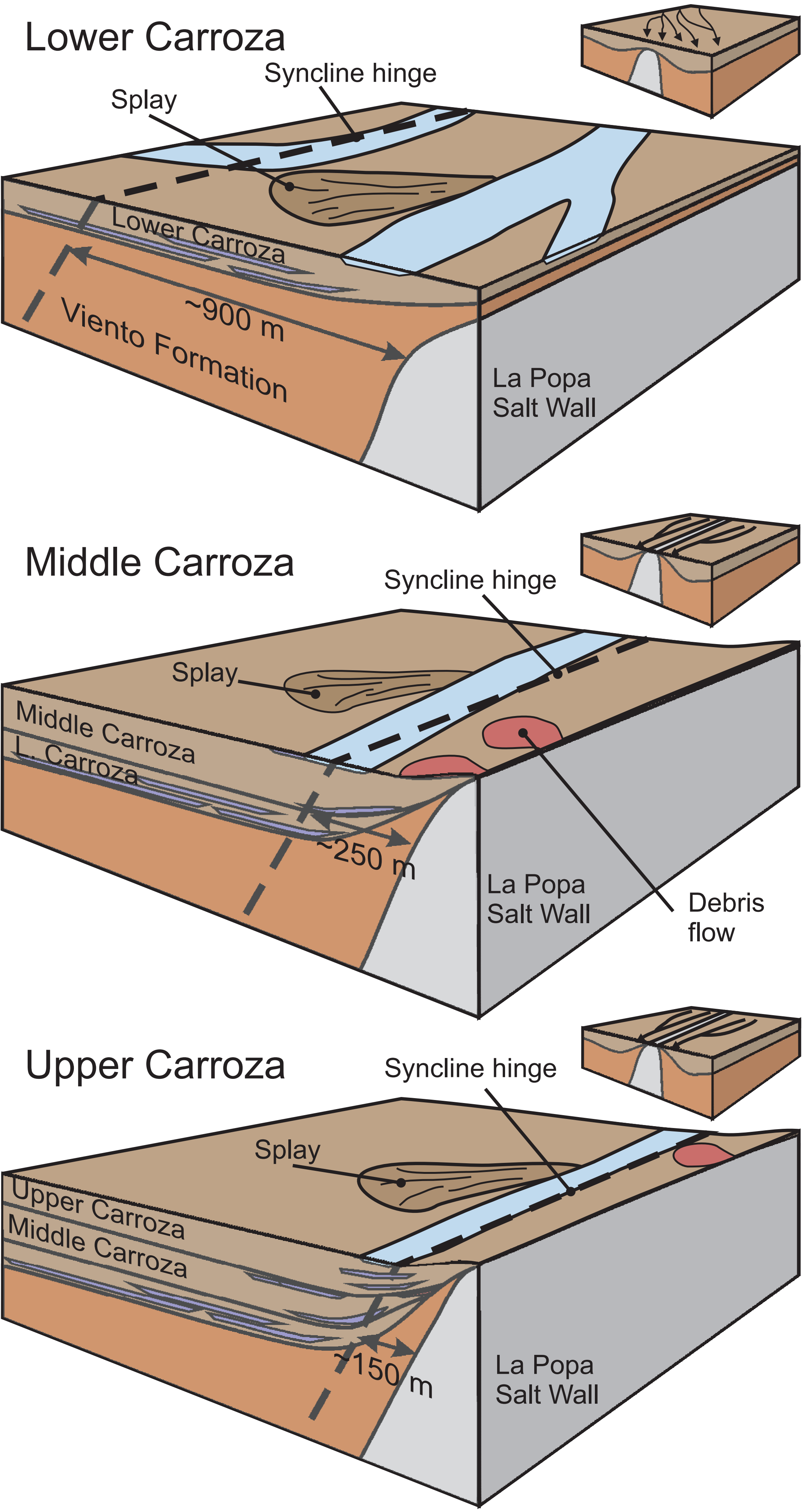
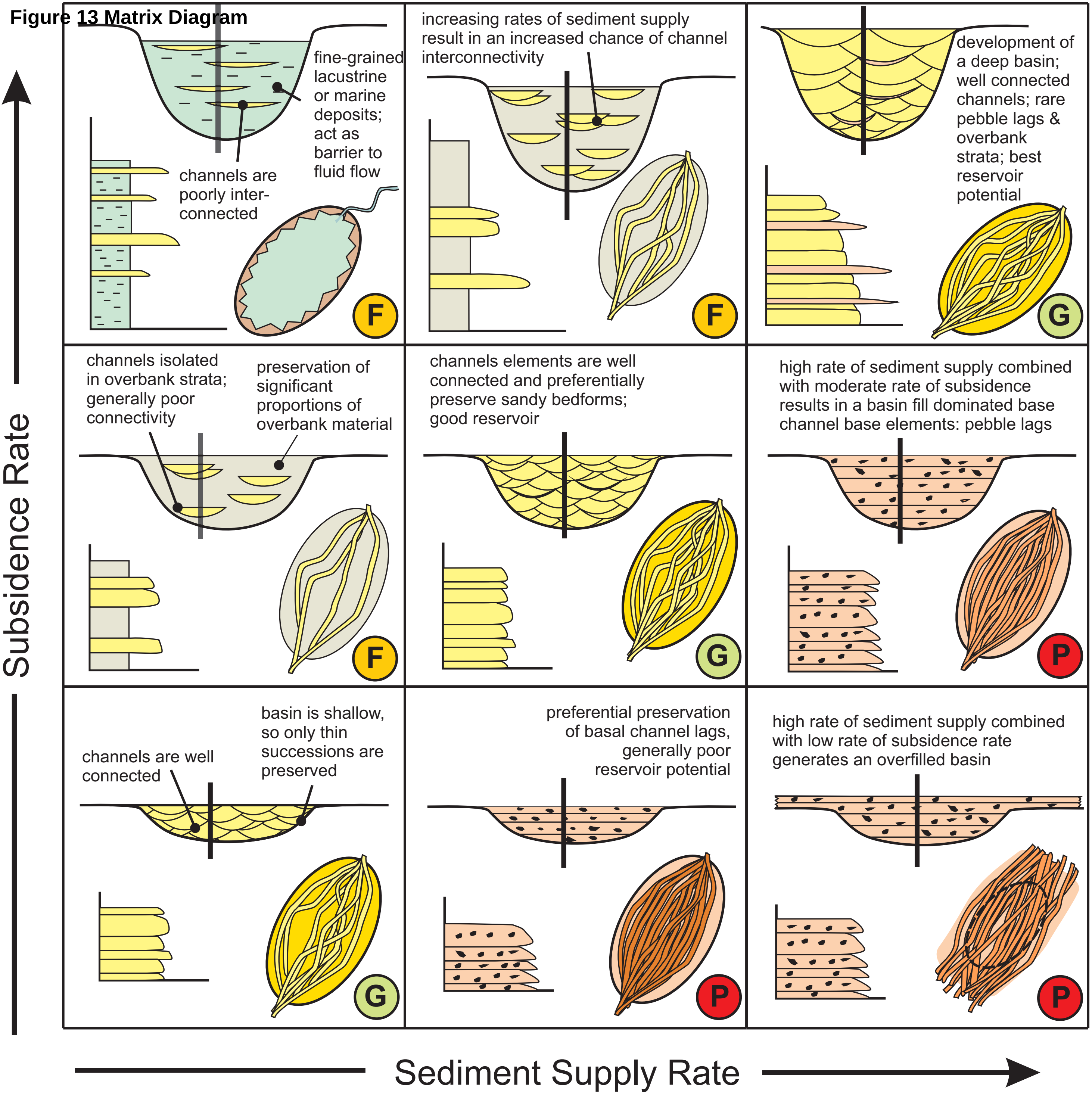


**Figure 13 Matrix Diagram**




very poorly sorted; clay-pebble

moderately to well sorted; sand

poor to moderately sorted; clay-silt

moderately sorted; silt-fine sand

G good reservoir potential

F fair reservoir potential

P poor reservoir potential

Matrix displays likely basin-fill architectures associated with various rates of sediment supply and subsidence.

Overfilled basins tend to occur when rates of sedimentation outpace rates of subsidence

Underfilled basins tend to occur when rates of subsidence outpace rates of sediment supply and the available accommodation space remains partly unfilled.

Filled Basins occur when the rates of sediment supply and subsidence are balanced.

**Figure 14ab Conclusion L1 L2**

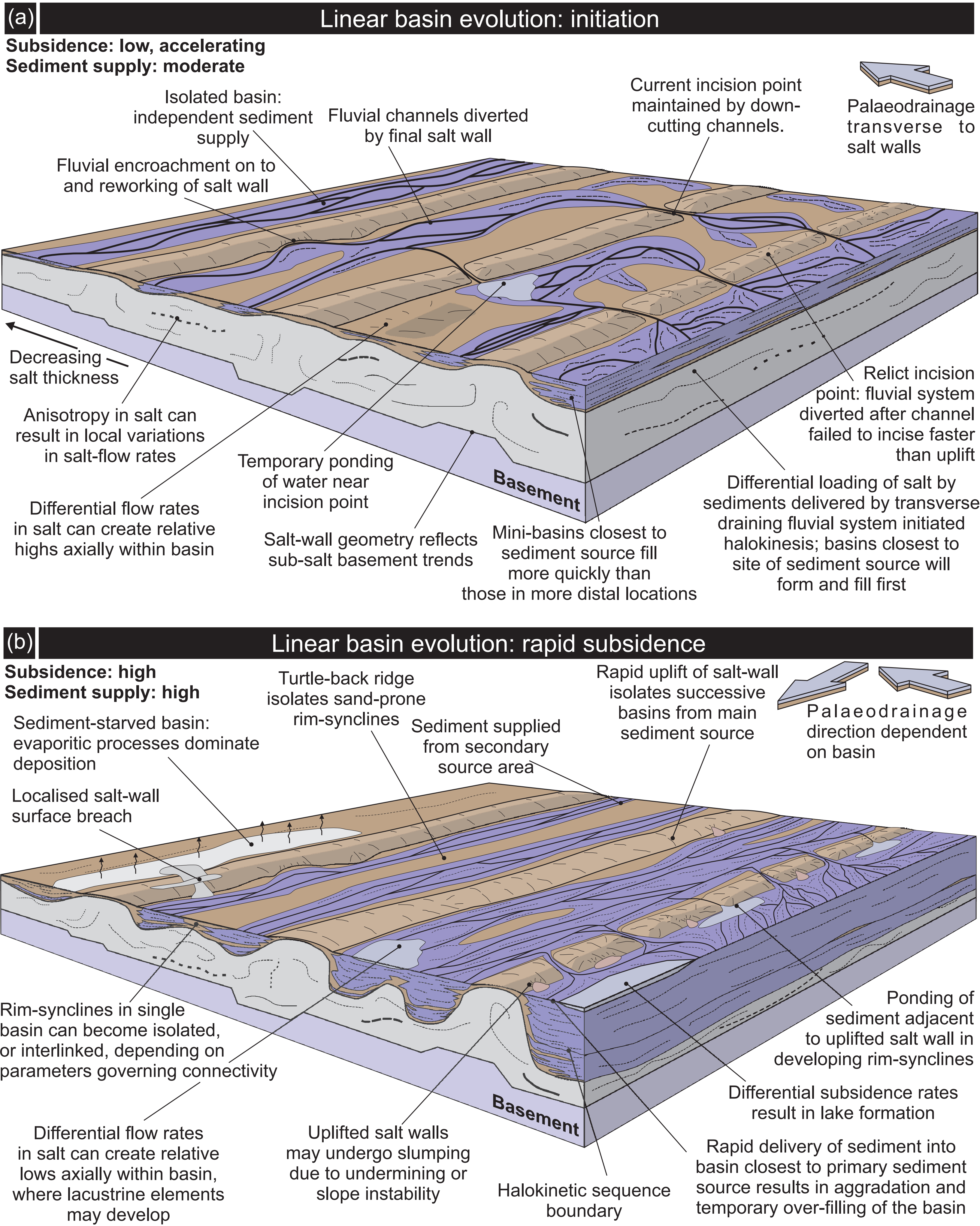


**Figure 14cd Conclusion L3 L4**

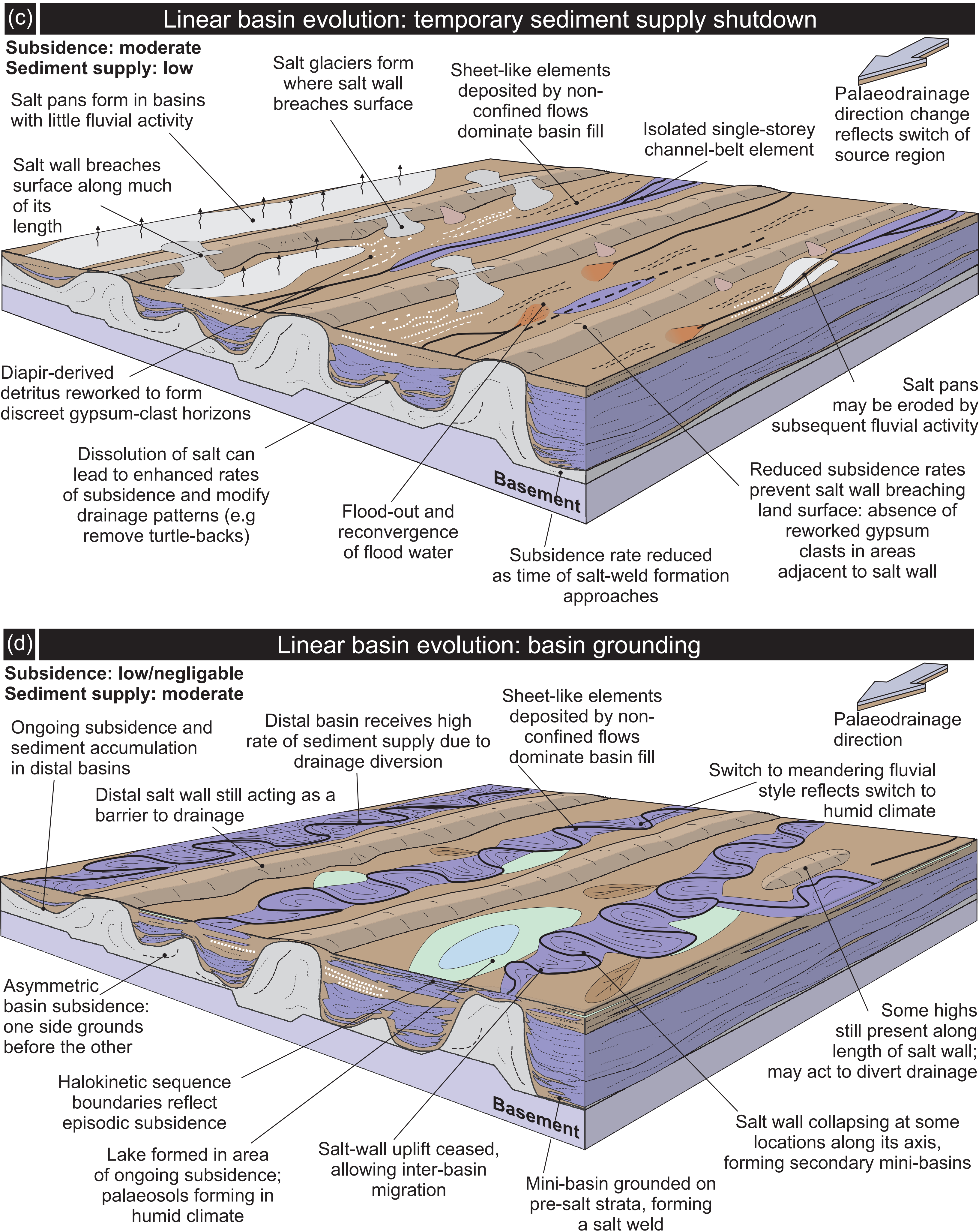


Figure 15ab Conclusion P1 P2

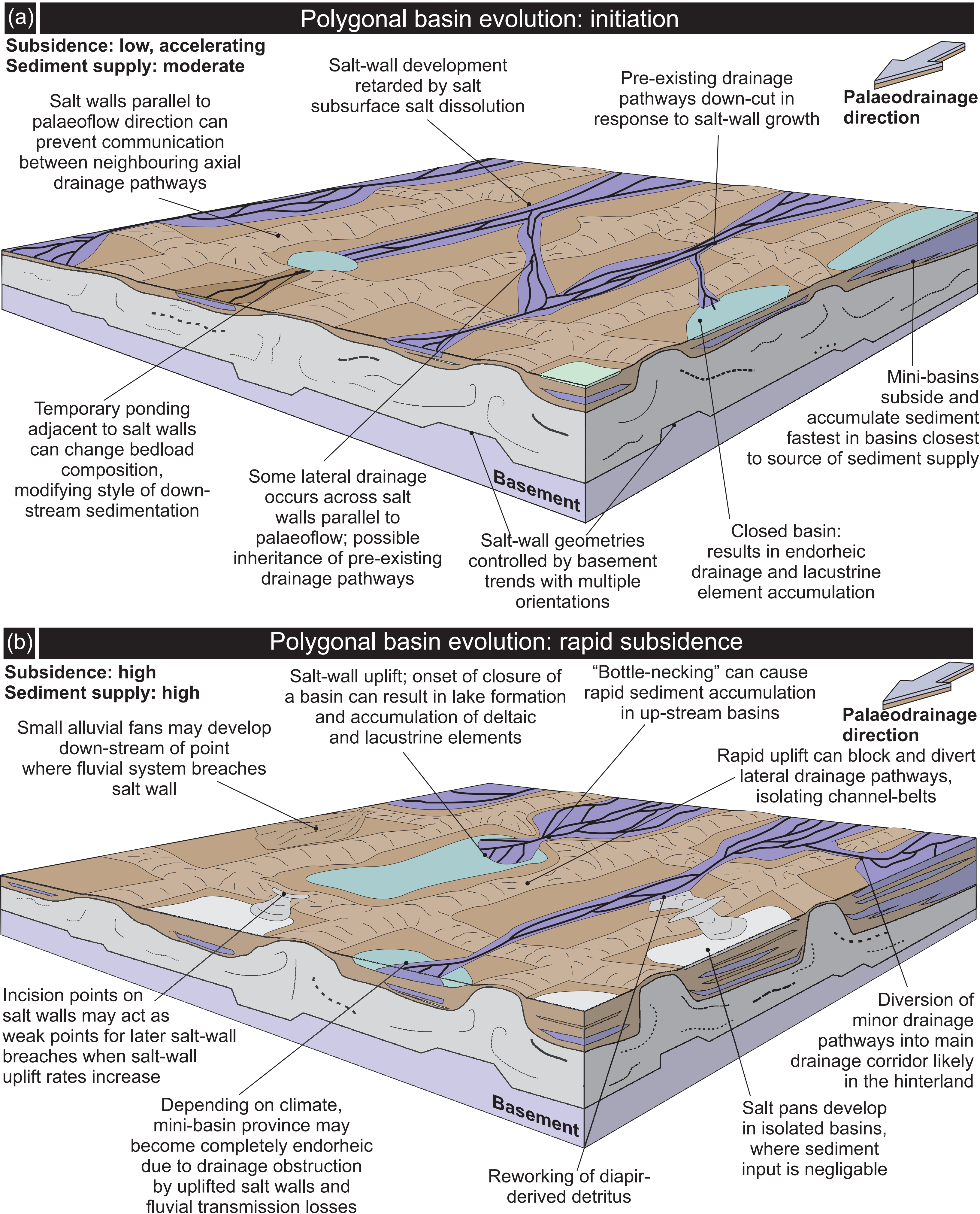
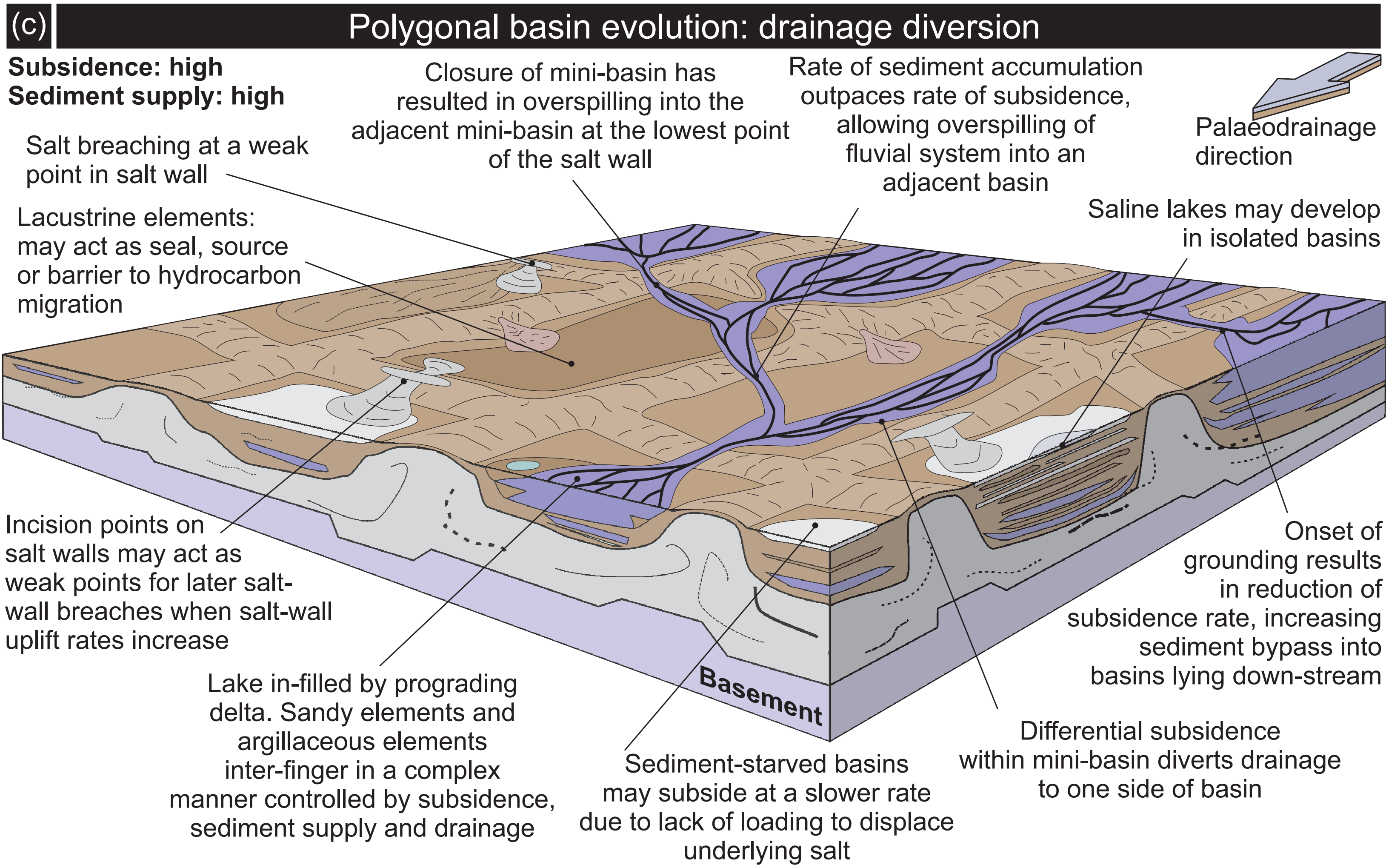


**Figure 15cd Conclusion P3 P4**



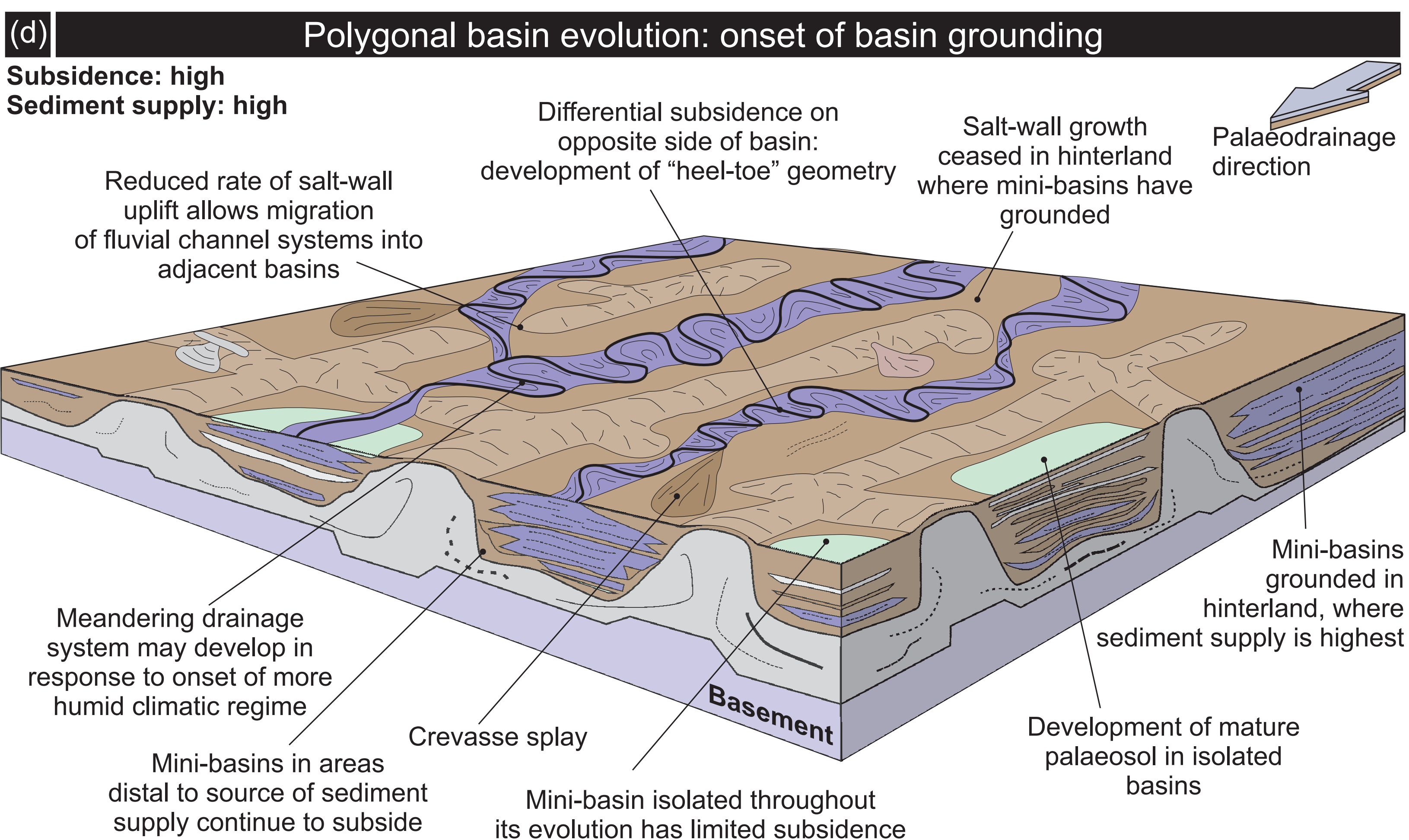



**Table 1**

Click here to download Table: 2013 08 13 Table 1 Full Case Study Table.xls

| Case Study | Age | Drainage Orientation | Climate | Type | Sed Supply Rate | Subsidence Rate | Fill Style |
|---|---|---|---|---|---|---|---|
| Paradox Basin | | | | | | | |
| **Cutler Group** | Permian | Transverse | Arid | Braided | High | High | Gravel Prone |
| **Moenkopi Fm** | Triassic (Lower) | Axial | Hyperarid | Braided/Nonconfined | Low | Moderate | Basin Dependent (Silt Prone) |
| **Chinle Fm** | Triassic (Upper) | Axial | Subhumid - Arid | L. Meandering U. Braided | High | Low | Sand Prone |
| **Kayenta Fm** | Jurassic | Tangental | Arid | Braided | High | Low | Sand Prone |
| Precaspian Basin | | | | | | | |
| **Tatarian** | Permian | Transverse | Arid | Braided, Evaporitic | | | Basin Dependent |
| **Triassic** | Triassic | Transverse | Semiarid | Braided, Lacustrine, Detaic | High | Moderate | Basin Dependent |
| Central Graben, CNS | | | | | | | |
| **Skaggerak Fm** | Triassic | Axial | Arid | Braided/NonConfined Terminal splay | Low-Mod. | Moderate | Basin Dependent (Sand Prone) |
| La Popa Basin | | | | | | | |
| **Carroza Fm** | Eocene | Axial | Arid | Braided | Moderate | ? | |
| Germany | | | | | | | |
| **Weisselster Basin** | Eocene | | Temperate | Meandering | | Low | |
| **River Weser & Aller** | Recent | Axial | Temperate | Meandering | | Low | |
| Canada | | | | | | | |
| **New Brunswick** | Carboniferous | - | - | - | - | - | - |
| **Severdrup Basin** | Juras. - Cret. | - | ?Temperate | - | - | - | - |