

From salt carapace to secondary minibasin encasement—The Bolon Secondary Minibasin, Eastern External Betics, SE Iberia

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ABSTRACT

The Bolon Secondary Minibasin (BSM) evolved on the now eroded Elda Salt Sheet, which contains fragments of the diapiric roof and syn-contractual sediments. Our detailed analysis of the BSM reveals how diapir rejuvenation and salt sheet evolution can affect the structural and stratigraphic architecture of secondary minibasins. We present a comprehensive analysis of the BSM integrating detailed cartography with stratigraphic, paleontological, and structural data. The field data show that the BSM contains a carapace of latest Cretaceous marlstones overlain by an up to 800 m thick roof of terrigenous and outer platform deposits. These suprasalt deposits are characterized by tabular beds without diapir derived detritus and are cut by a series of extensional growth faults that sole into the salt. In the Oligocene, diapir rejuvenation due to contractional deformation resulted in the breaching and dismemberment of this diapiric roof and extrusion of allochthonous salt. Roof dismemberment is recorded as an unconformity truncating the roof strata and diapir derived detritus in the Oligocene-lower Miocene units. Throughout the middle Miocene composite halokinetic sequences, rapidly shifting depocenters, episodic unconformities, and stratigraphic onlaps record the rapid sinking and progressive northward rotation of the BSM. Debrite wedges, diapir derived detritus, and ramp-flat geometries along the upper salt-sediment contact show that in the middle Miocene the BSM was fully encased in allochthonous salt. Continued shortening and further extrusion of allochthonous salt rotated the BSM an additional 60° to the NW before its eventual grounding on subsalt strata.

1. Introduction

Allochthonous salt is commonly encountered in sedimentary basins and orogens worldwide and represents a key area of interest owing to its common association with oil and gas fields, and its potential for geothermal energy production, or for hydrogen or CO₂ storage (Hudec and Jackson, 2006; Jackson and Hudec, 2017; Pedrera et al., 2020; Duffy et al., 2023; Wei et al., 2023). It is defined as “sub horizontal or moderately dipping, sheet-like salt diapirs emplaced at stratigraphic levels above the autochthonous, or original, salt layer” (Hudec and Jackson, 2006). The evolution of allochthonous salt structures is principally controlled by the relative rates of salt rise versus sedimentation resulting in a complex stratigraphic and structural architectures (e.g.

Hudec and Jackson, 2006, 2009; Hudec et al., 2009; Hearon IV et al., 2015; Rowan, 2017; Jackson and Hudec, 2017; Fernandez et al., 2021). Research into allochthonous salt systems has constrained the modes of emplacement and breakout, styles and types of advance, and the post-emplacement evolutionary styles of allochthonous salt structures (e.g. Talbot, 1998; Hudec and Jackson, 2006; Hearon IV et al., 2015; Rowan, 2017; Williams et al., 2019; Rowan et al., 2020; Flinch and Soto, 2022).

Salt in orogenic settings can act as a regional detachment concentrating deformation along salt layers and into salt structures that become the locus of subsequent contractional deformation (Davis and Engelder, 1985; Rowan and Vendeville, 2006; Duffy et al., 2018, 2021; Santolaria et al., 2021). When a sedimentary basin contains well developed salt

Unit conventions: Dip direction and dip of bedding: dip direction-dip (###°, ##°), Lineations: plunge and bearing (##°, ###°).

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structures, shortening often results in the expulsion of salt sheets driven by the squeezing of salt diapirs and walls, or derived from source fed thrusts (Hudec and Jackson, 2006; Dooley et al., 2015; Hearon IV et al., 2015; Rowan, 2017; Kergaravat et al., 2017; Duffy et al., 2018; Célini et al., 2022).

Allochthonous salt breakout is associated with the rupture of the diapiric roof or carapace that overlies the salt structure (Fig. 1a). We use the term carapace as defined by Hart et al. (2004), meaning “condensed, sub-parallel/parallel strata originally deposited over bathymetrically raised salt diapirs, walls, canopies, and sheets”, whereas we use the term “roof” when discussing sediments deposited over salt structures without a condensed character (Hudec and Jackson, 2006; Dooley et al., 2015; Hearon IV et al., 2015; Rowan, 2017; Petrovska et al., 2021). Post breakout, allochthonous salt can carry carapace fragments and secondary minibasins for multiple kilometers as shown in the seismic line from the Gulf of Mexico (GOM) in Fig. 1d 1 and 2 (e.g. Fiduk et al., 2014; Dooley et al., 2015; Dooley and Hudec, 2020; Fernandez et al., 2021; Flinch and Soto, 2022). With continued salt advance these fragments can interact resulting in regions of contraction or extension (see upslope and downslope side of bucket minibasin, respectively and shown in Fig. 1d 2) based on their distribution and proximity to each other, their interactions with subsalt topography, and variations in salt flow caused by regional slopes, topographic barriers, coalescing salt sheets, developing secondary minibasins, or regional tectonics (Hudec and Jackson, 2006, 2009; Hudec et al., 2009; Jackson et al., 2010; Rowan, 2017; Fernandez et al., 2021; Ferrer et al., 2024).

While high-resolution seismic data have been instrumental in identifying and characterizing these structures at a hectometric-kilometric scale (i.e. Fiduk et al., 2014; Fernandez et al., 2017, 2021), a limited number of field studies have delved below the seismic scale to provide an outcrop scale analysis of mesoscopic structures, stratigraphic

architectures, and kinematic indicators in these complex settings (e.g. Giles and Rowan, 2012; Hearon IV et al., 2015; Saura et al., 2016; Kergaravat et al., 2017; Vidal-Royo et al., 2021; Gannaway Dalton et al., 2022). Leveraging field studies as analogues for seismic interpretations and analog/numerical modeling has proven crucial in ground truthing our understanding of salt tectonic processes.

This paper is based on the description and analyses of an outcropping secondary minibasin—the Bolon Secondary Minibasin (BSM)—that, unlike most of the studied cases, mainly developed in deepwater (>100 m water depth) environments. This is a minibasin that formed due to diapir rejuvenation and subsequent roof fragmentation during the latest Cretaceous-middle Miocene contractional building of the Betic orogen. Our study seeks to decipher the main features of the BSM, relating its evolution to the growth and kinematics of the Elda Salt Sheet and regional tectonic evolution of the Betic fold and thrust belt (FTB). The questions we address are: (1) how does carapace and diapiric roof evolve during shortening and after allochthonous salt breakout and roof fragmentation; (2) how do roof fragments affect the location and evolution of secondary minibasins and how does the roof fragment condition the shape of secondary minibasins and the development of halokinetic sequences; (3) what role does water depth and bathymetry play in the evolution of secondary minibasins; (4) what are the characteristics and timing of secondary minibasin rotations; (5) how are secondary minibasins encased and can we provide field evidence for the style of encasement; and (6) how does the BSM compare to other secondary minibasins in other salt basins?

2. Geologic setting

The Betic orogen is a Cenozoic fold and thrust belt that formed by the north-south (N-S) to north northwest-south southeast (NNW-SSE)

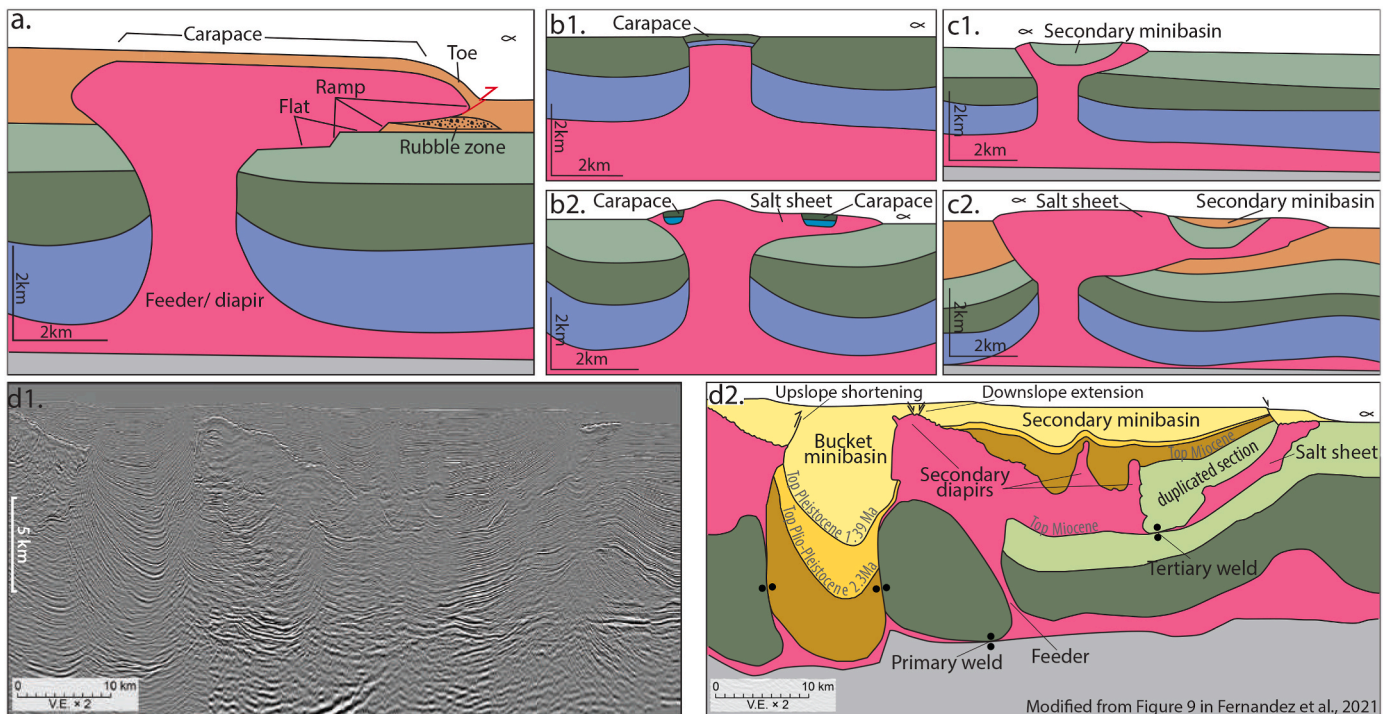


Fig. 1. Anatomy of a salt sheet, styles of carapace evolution, and seismic example of a salt canopy in the Gulf of Mexico (GOM). a. A schematic section of a salt sheet covered by carapace strata (notice the condensed nature of the carapace strata compared to regional). Common features associated with salt sheets are labeled (modified from: Jackson and Hudec, 2017). b. Schematic evolution of carapace strata in a salt sheet. b1. Carapace above a diapir before crestal extension and salt inflation. b2. Carapace fragments post-dismemberment (modified from: Jackson and Hudec, 2017). c. Schematic evolution of carapace strata incorporated into a secondary minibasin on a salt sheet (modified from Jackson and Hudec, 2017). (d) Seismic line from the GOM showing complex relationships between transported carapace strata, secondary minibasins, allochthonous salt advance, and downslope gliding. Notice the repeated top Miocene interval in the secondary minibasin (modified from Fig. 9 in Fernandez et al., 2021).

oriented collision of the African and Eurasian-Iberian plates (Dercourt et al., 1986; Dewey et al., 1989; Weijermars, 1991; Jabaloy Sánchez et al., 2019; van Hinsbergen et al., 2020). It is a fold and thrust belt that resulted from the moderate inversion of a portion of the passive margin that developed at the northern side of the Alpine Tethys during Middle Jurassic to early Santonian times (Bernoulli and Lemoine, 1980; Ziegler, 1990; Schettino and Turco, 2011; Frizon De Lamotte et al., 2011). Traditionally, the Betic orogen is subdivided into the Internal Zones to the south and the External Zones to the north (Fig. 2a; Blumenthal, 1927; Fallot, 1948; Vera, 2004). The Internal Zones correspond to an allochthonous stack of thrust sheets comprising Triassic and older metamorphic rocks (Torres-Roldan, 1979; De Jong, 1990; Jabaloy Sánchez et al., 2019). The External Zones consists of a broad orogenic wedge of folds and thrusts detached from the Iberian basement along the autochthonous Triassic salt (Fig. 2c; Roca et al., 2006; Escosa et al., 2018; Jabaloy Sánchez et al., 2019; Pedrera et al., 2020). The structural grain is dominated by northeast (NE)-to east northeast (ENE)-trending folds and thrusts, some pierced by diapirs of Triassic salt, which were emplaced from Mesozoic to Quaternary times (Moseley et al., 1981; De Ruig, 1992, 1995; Martínez del Olmo et al., 2015; Escosa et al., 2018).

The External Zones of the Betics, where the study area is located, resulted from the moderate inversion of the Eastern South Iberian passive margin (Fig. 2c). They are historically subdivided based on tectonic and Mesozoic stratigraphic criteria into the Prebetics, to the north, and

the Subbetics, to the south, (Fig. 2b). The Mesozoic stratigraphy of the Prebetics is characterized by southeastwards deepening carbonate platform deposits that are historically divided into the External and Internal Prebetics, located to the north and south respectively (Fig. 2c). The Subbetics correspond to a major thin-skinned thrust sheet formed by deep basinal and pelagic deposits (García-Hernández et al., 1980; Martínez del Olmo et al., 2015; Escosa et al., 2018; Pedrera et al., 2020). This facies distribution is a relic of the Mesozoic paleogeography and structure of the stepped and rifted passive continental margin of Southern Iberia (García-Hernández et al., 1980; De Ruig, 1995; Vera, 2004; Martínez del Olmo et al., 2015; Martín-Chivelet et al., 2019).

2.1. Tectono-stratigraphic evolution of the Eastern Prebetics

The remains of the Elda Salt Sheet are located in the Internal Prebetics (Fig. 2c) where the present-day outcrops preserve a Middle Triassic to Cenozoic tectonic evolution. Three main stages can be differentiated: (1) a Middle Triassic to Middle Jurassic pre-rift stage; (2) a Late Jurassic to early Santonian syn-rift to passive margin stage; and (3) a late Santonian to Quaternary syn-contractual stage (Fig. 3; De Ruig, 1992, 1995; Vera, 2004; Pedrera et al., 2014, 2020; Escosa et al., 2018; Jabaloy Sánchez et al., 2019; Vergés et al., 2019; Daudet et al., 2020).

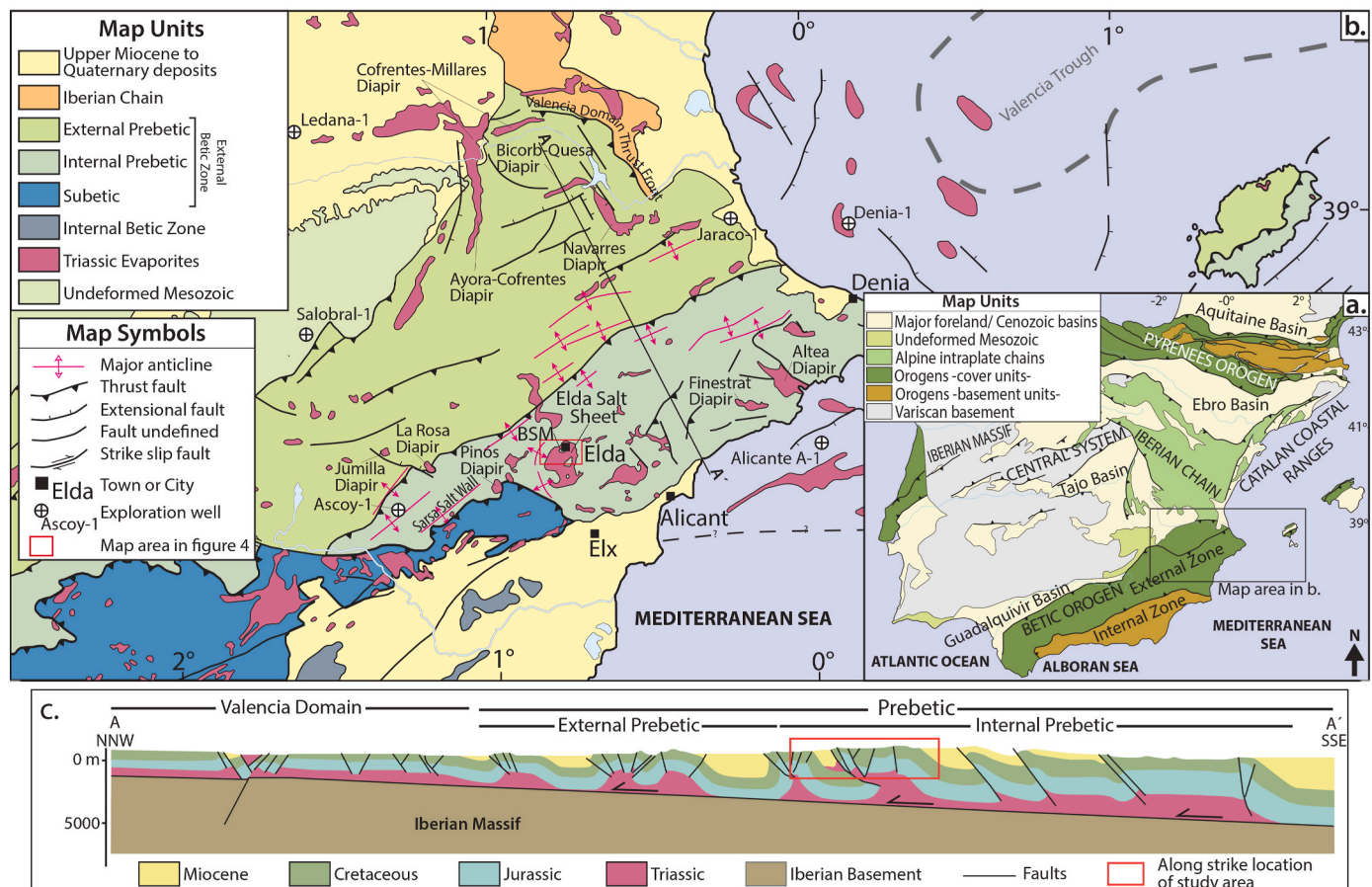


Fig. 2. a. Simplified tectonic map of the Iberian Peninsula showing the main inverted Mesozoic basins and Cenozoic mountain belts (shades of green), and foreland/cenozoic basins (tan). The black box highlights the map area in b. (modified from: Vergés et al. (2019)). b. Tectonic map showing the distribution of the major tectono-stratigraphic units at the eastern edge of Betic orogen. Notice the location of exposed Triassic salt (pink) and the location of the study area (red box in b.) in relation to the boundary of the Internal and External Prebetics (data compiled and modified from: Martínez del Olmo et al., 1974; Leret-Verdu et al., 1976; Gallego-Coiduras et al., 1981; De Ruig, 1995; Roca et al., 2006; Jackson and Hudec, 2017; Escosa et al., 2018; Pedrera et al., 2020). c. Cross section A-A' through the Eastern Prebetics showing the regional structural style, (red box shows the along strike location of the study area) modified from Roca et al. (2006). (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

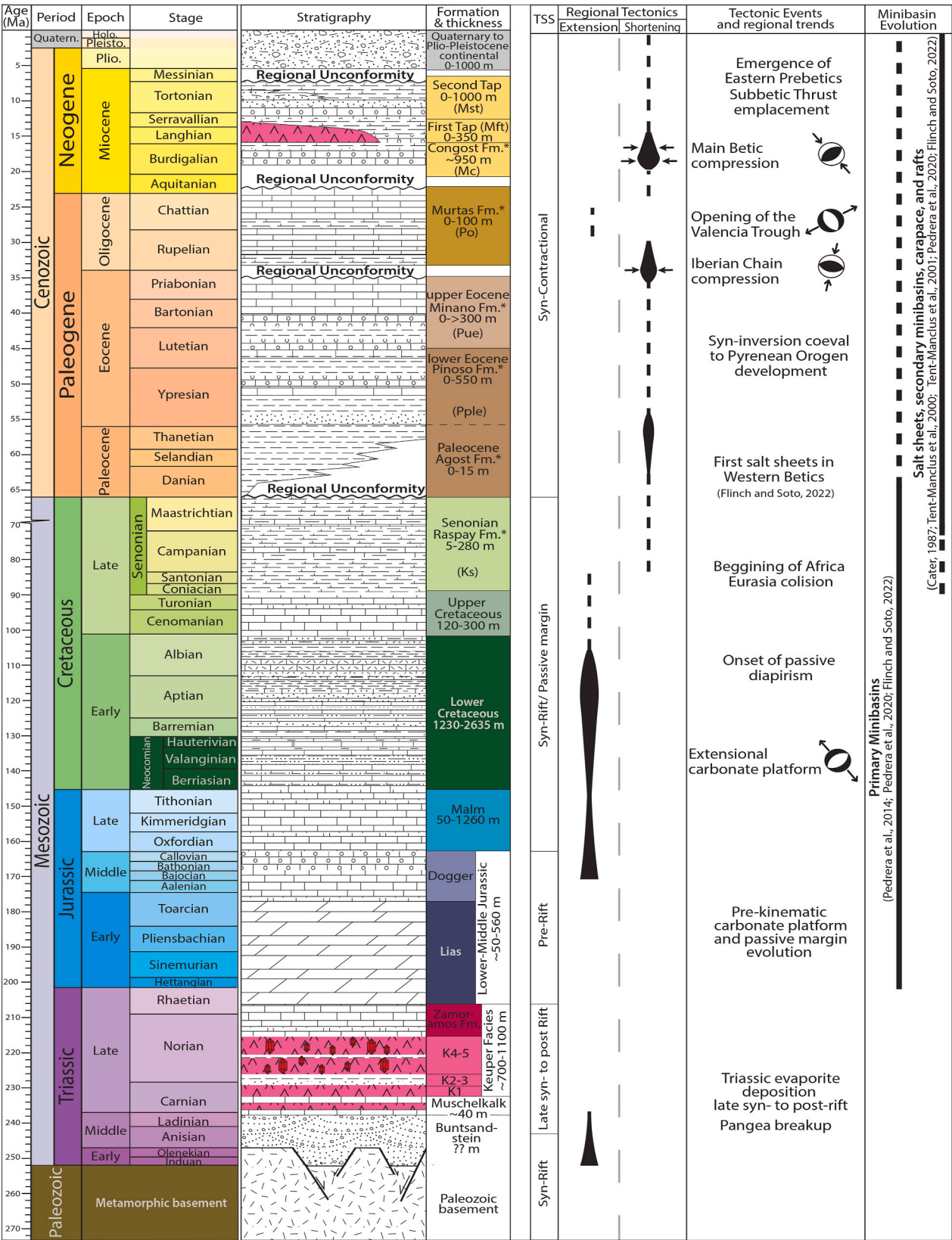


Fig. 3. Tectonostratigraphic panel of the Internal Prebetics. The left box shows ages, stratigraphy, lithology, and Formation names used throughout the text, the right box shows the tectonostratigraphic sequences (TSS), regional tectonic setting, important events, and the evolutionary timing of salt related structures (regional data compiled and modified from: Garcia-Hernandez et al., 1980; De Galdeano, 1990; Weijermars, 1991; Vera, 2004; Martín-Chivelet and Chacón, 2007; Flinch and Soto, 2017, 2022; Martín-Martin et al., 2018b, 2020b; Vergés et al., 2019; van Hinsbergen et al., 2020; Pedrera et al., 2020).

2.1.1. Middle Triassic to Middle Jurassic: pre-rift stage

The Middle Triassic to Middle Jurassic pre-rift stage post-dates the breakup of Pangea that occurred during Permian-Early Triassic times. This stage is characterized by two well differentiated and initially isopachous units: a lower one, Middle to Late Triassic in age, made by a layered evaporite sequence (LES) that constitutes the major décollement of the region and feeds the diapiric structures; and an upper one also Late Triassic in age made by marine carbonate platform dolostones that grade upwards into limestones (Ortí et al., 2017; Gómez et al., 2019; Guerrero et al., 2021). The Middle-Late Triassic LES includes a basal layer of laminated to massive shallow-marine carbonates and evaporites (Mushelkalk facies) and above the detritic-evaporitic units of the Keuper facies that include: clays with thin beds of gypsum, halite, and sandstones in the lower portion; thick layers of sandstone and mudstone in the middle section; and a thick succession of halite with clay and gypsum interbeds and abundant diagenetic red and white bipyramidal quartz crystals—the so-called “Jacintos de Compostela” in the upper part (Pérez-Valera and Pérez-López, 2008; Ortí et al., 2017; Herrero et al., 2020). Note that in the remainder of the paper we use the term Triassic salt to refer to the entire deformed LES regardless of caprock formation or the current proportions of halite versus other lithologies.

2.1.2. Late Jurassic to early Santonian: syn-rift to passive margin stage

This stage is characterized by a major Late Jurassic to early Santonian rift event that affected the Eastern External Betic Zones and is related to both the continued opening of the Central Atlantic and synchronous development of extensional basins in the western Tethyan realm (Ziegler, 1990; Hanne et al., 2003). This rifting allowed for the deposition of the Late Jurassic to early Santonian syn-extensional sediments that are up to 3850 m thick (shown by the Ascoy-1 well ~56 km along strike, location shown in Fig. 2) but show large thickness variations in the Eastern Prebetics (Escosa et al., 2018; Martín-Chivelet et al., 2019). These sediments are mostly marine carbonates with an important siliciclastic input at the beginning of the late Albian marking the end of rifting and the onset passive margin development (Fig. 3; Granier and Perthuisot, 2009; Escosa et al., 2018; Martín-Chivelet et al., 2019; Pedrera et al., 2020). The deposition of these sediments took place in southward deepening depositional environments and was accompanied by extensionally triggered salt diapirism (Rodríguez-Estrella, 1977b, 1977a; Rondeel and Gaag, 1986; De Ruig, 1995; Granier and Perthuisot, 2009; Martínez del Olmo et al., 2015; Escosa et al., 2018; Martín-Chivelet et al., 2019). Throughout the Cretaceous, both reactive and passive diapirism played a pivotal role in the margins evolution, influencing the structural and stratigraphic architecture of the Prebetics (Rondeel and Gaag, 1986; De Ruig, 1995; Granier and Perthuisot, 2009; Martínez del Olmo et al., 2015).

2.1.3. Late Cretaceous (late Santonian) to Tortonian: syn-contractional stage

This stage begins with the onset of north (N) to north-northwest (NNW) oriented convergence of Africa-Eurasia at late Santonian times (around 85 Ma) that is recorded by the development of local unconformities and the initial squeezing of pre-existing diapirs (Chacón and Martín-Chivelet, 2001, 2003; Martínez del Olmo et al., 2015; Jabaloy Sánchez et al., 2019; Martín-Chivelet et al., 2019; Vergés et al., 2019; Daudet et al., 2020). It was followed with a Campanian-late Eocene period of relative tectonic quiescence in which a succession of pelagic carbonates (Raspay Formation), detrital carbonates, turbidites (Pinoso Formation), and platform limestones (Minano Formation) were deposited (Fig. 3). These sediments thicken and represent deeper water domains towards the south-southeast (SSE; Martín-Chivelet and Chacón, 2007; Guerrero and Martín-Martín, 2014; Martín-Martín et al., 2018b, 2020a, 2020b; Guerrero et al., 2021).

Subsequently, at the latest Eocene-Oligocene times (from 45 to 25 Ma), there was a second major contractional event. This was coeval to the Iberian intraplate contractional deformation that was responsible for

the formation of the Iberian Ranges (Fig. 3). The contractional episode is marked by a regional angular unconformity and was followed by a short latest Oligocene-Aquitania period of relative tectonic quiescence in which continental to shallow marine sediments were deposited (Fig. 3; Hillebrandt, 1974; Ott D'Estevou et al., 1988; Geel, 1995; Guerrero et al., 2014; Martínez del Olmo et al., 2015; Martín-Martín et al., 2018b; Jabaloy Sánchez et al., 2019; Martín-Chivelet et al., 2019).

Finally, a third and main contractional deformation took place in the early to middle Miocene (from 23 to 11 Ma) when the External Betics fold and thrust belt was formed (Fig. 3; Azema et al., 1979; García-Hernández et al., 1980; Geel, 1995; Martín-Martín et al., 2018b, 2020b; Jabaloy Sánchez et al., 2019). This contractional deformation is characterized by the development of thrusts and folds detached at the autochthonous Triassic salt, pronounced squeezing and rejuvenation of pre-existing diapirs, and extrusion of allochthonous salt sheets (Martínez del Olmo et al., 2015; Escosa et al., 2018; Jabaloy Sánchez et al., 2019; Pedrera et al., 2020; Flinch and Soto, 2022). From the Serravallian to the present day, the region remains under the influence of regional shortening. However, contractional deformation has diminished significantly and is manifested solely through localized folding and reverse to strike-slip motion of basement faults (Van Balen et al., 2015; Martín-Martín et al., 2018b, 2020b; Jabaloy Sánchez et al., 2019).

2.2. The Elda Salt Sheet

The Elda Salt Sheet is composed of Triassic salt and covers an area of more than 110 km² encompassing the towns of Elda and Petrer in the north, Monovar in the east, and Novelda in the south (Fig. 4). The true extent of the Triassic salt is unknown but was likely much larger than the current outcrop extent. In general, it occupies the lowest topography in the Vinalopo River Valley. At the surface, the halite is dissolved so that the upper Keuper unit crops out as a caprock made of laminated, massive, and brecciated gypsum and mudstones with abundant diagenetic red and white bipyramidal quartz crystals (Gil Marco, 2013; Poe et al., 2018; Herrero et al., 2020; Závada et al., 2021). Numerous hyper-saline springs located in the Vinalopo River Valley indicate significant ongoing subsurface halite dissolution within the LES (Fig. 4; Andreu et al., 2010). To the northwest (NW) and southeast (SE), the salt is bound by large-wavelength (5–10's of km) N to northeast (NE) trending antiforms comprised of Mesozoic to middle Miocene stratigraphy. To the NW, the Elda Salt Sheet overlies the Langhian strata of the l'Ombria Anticline, to the north they overlie the Langhian strata of the Camara Anticline, and to the east they are below and adjacent to the Serra del Cid antiformal structure (Vera, 2004; Martínez del Olmo et al., 2015).

The allochthonous nature of the Elda Salt Sheet is best expressed along its NW contact where the LES lies above the SE dipping Langhian stratigraphy on the SE limb of the l'Ombria Anticline (Vera, 2004; Martínez del Olmo et al., 2015). Along the SE most boundary of the l'Ombria anticline, subsalt stratal cutoffs indicate the progressive advance of an allochthonous sheet throughout the early to middle Miocene (Fig. 4). The Elda Salt Sheet is underlain by thick, gently folded sedimentary units of the latest Triassic to earliest Miocene successions of the Internal Prebetics as shown by the Ascoy-1 well to the west (Fig. 2) and along strike in the Sarsa Salt Wall area (see Fig. 7 in Escosa et al., 2018). Currently, the Elda Salt Sheet is partially covered by Campanian to Serravallian (83–11 Ma) deposits that exist as isolated outcrops surrounded by Triassic salt. Our study focuses on one of these non-evaporite bodies, El Bolon Mountain, that exposes Campanian (Senonian) - Langhian (83–13 Ma) aged deposits completely surrounded by Triassic salt (Fig. 4).

3. Data/observations

The Bolon structure is located in the north-central part of the Elda Salt Sheet, NE of the city of Elda covering the El Bolon Mountain and

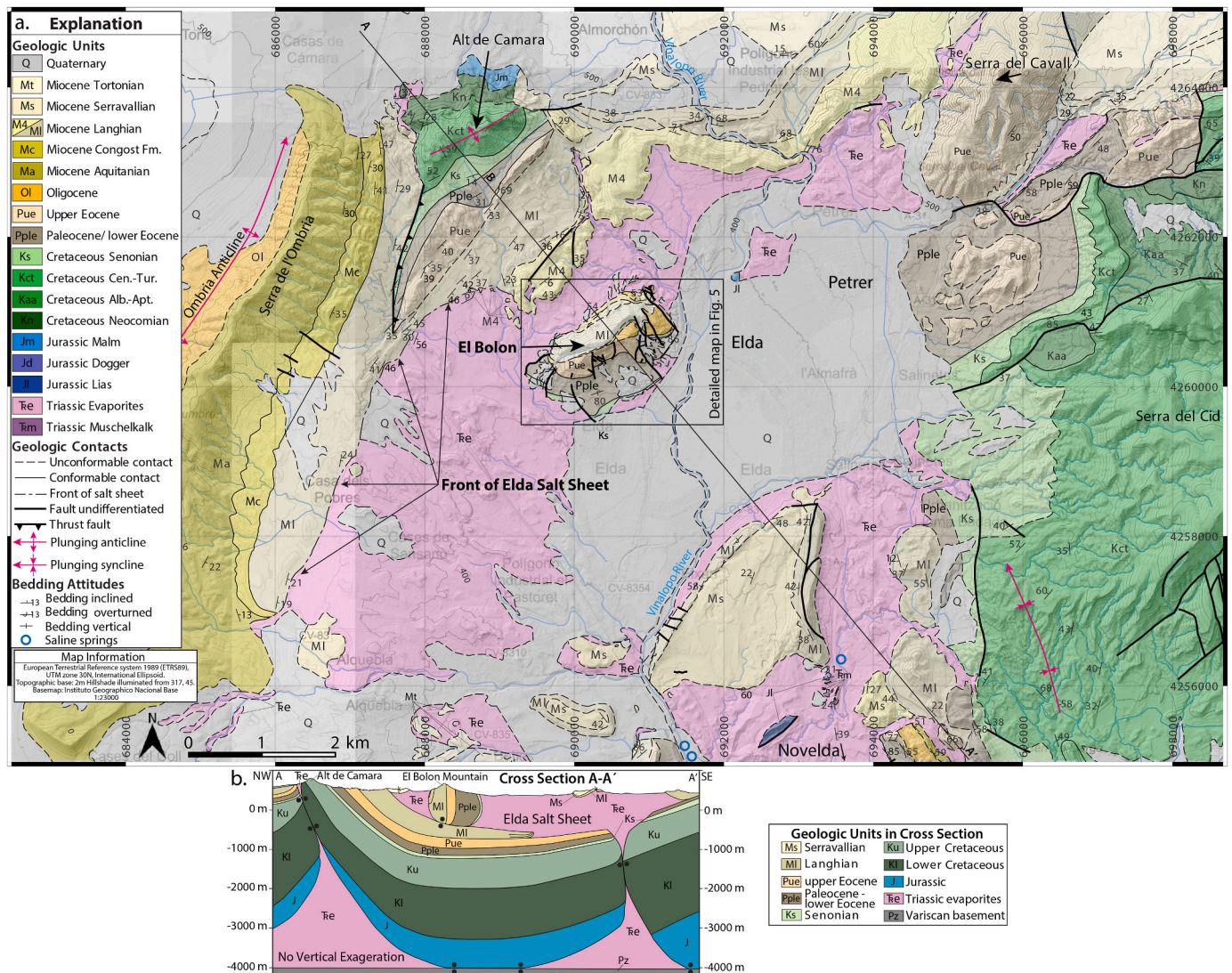


Fig. 4. a. Geologic map of the Elda region showing the location of the Bolon structure and the front of the Elda Salt Sheet (modified from: [Martinez del Olmo et al., 1974](#); [Leret-Verdu et al., 1976](#); [Gallego-Coiduras et al., 1981](#); supplemented by our own mapping). b. Simplified cross section through the Elda Salt Sheet and El Bolon Mountain.

surrounding foothills encompassing an area of $\sim 2.2 \text{ km}^2$ (Fig. 4). It is a $\sim 1200\text{-m}$ -thick succession of strata that crop out with rather continuous exposures. In general, the strata strike NE-SW with a northward stratigraphic polarity although multiple unconformities, faults, and folds occur near the salt-sediment contact and throughout the central portion of the structure resulting in complex three-dimensional structural relationships (Fig. 5).

3.1. Stratigraphy

We conducted detailed geologic mapping ($<5 \text{ m}^2$ resolution) collecting more than 1500 structural data points and measured nine stratigraphic sections throughout the Bolon structure to resolve the complex stratigraphic and structural relationships. Fieldwork was supplemented by a high-resolution 3-D model compiled from drone imagery (Supplementary Material 1) and thin section petrography/foraminiferal analysis. The key planktonic and benthic foraminifera observed in the samples collected from the Bolon structure are provided in Appendix 1. The key species used to constrain the age of the samples are shown in Fig. 6. From these data we differentiate eight lithostratigraphic units (Fig. 5). Here we describe the general stratigraphy within these units saving the detailed analysis of the salt-sediment contacts in the SE,

southwestern (SW), NE, and northwestern (NW) terminations for the following subsections.

3.1.1. Senonian—Ks (83–66 Ma)

The oldest unit in the Bolon structure are white marlstones that were deposited unconformably above the Triassic salt (Figs. 5 and 7a). It is composed of white and sometimes pink marls, silty marlstones, and minor marly limestones-dolostones, deposited in tabular centimetric to decimetric beds with no diapir derived detritus observed (Fig. 8). Across the field area this unit varies in thickness ranging from 0 to 110 m (Fig. 5). In general, the unit is thicker in the west (location 1 in Fig. 5; 35–100 m), thins towards the Trinitary Fault (location 2 in Fig. 5; 25 m), is much thinner in the footwall of the Trinitary Fault (location 3 in Fig. 5; 12 m), and is absent beneath the lower Eocene unit to the northeast of the East Bolon Fault (location 4 in Fig. 5; Figs. 7 and 8).

The age of this unit is constrained by planktonic foraminifera in sample Bol-5-03 (sample location shown by black dots in Fig. 5), specifically *Globotruncanella stuarti* (Lapparent, 1918), *Radotruncana subspinosus* (Pessagno, 1960), and *Rugoglobigerina* spp. (Brönnimann, 1952, Fig. 5 and Appendix Id). These foraminifera indicate a late Campanian-early Maastrichtian age with deposition taking place in hemipelagic outer ramp to upper bathyal settings (Martin-Chivelet et al.,

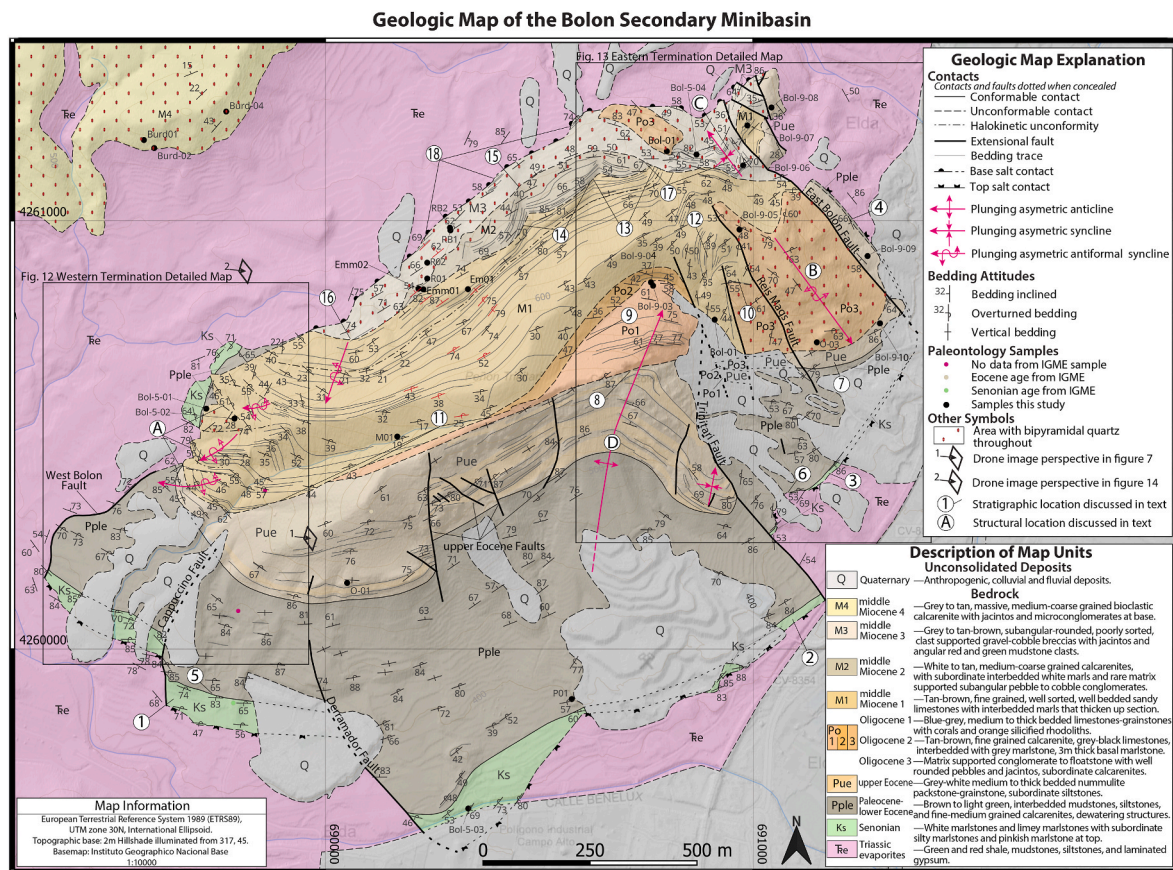


Fig. 5. Geologic map of the Bolon Secondary Minibasin. Sample locations for foraminiferal analysis are shown as black dots, bedding data used in the rotational analysis of Fig. 16a are shown in red, locations discussed in the text are labeled with numbers (stratigraphic locations) and letters (structural locations), locations and orientation of drone images are shown as numbered rectangles with arrows. Previous foraminiferal ages from the *El Instituto Geológico y Minero del Estado* (IGME) are also plotted. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

2019). This unit correlates with the Raspay Formation and the Capas Blancas Formation that attain a regional thickness of 280 m (Chacón and Martín-Chivelet, 2005; Martín-Chivelet and Chacón, 2007; Guerrero et al., 2014). Therefore, the Senonian unit in the Bolon structure shows a reduced thickness compared to the regional.

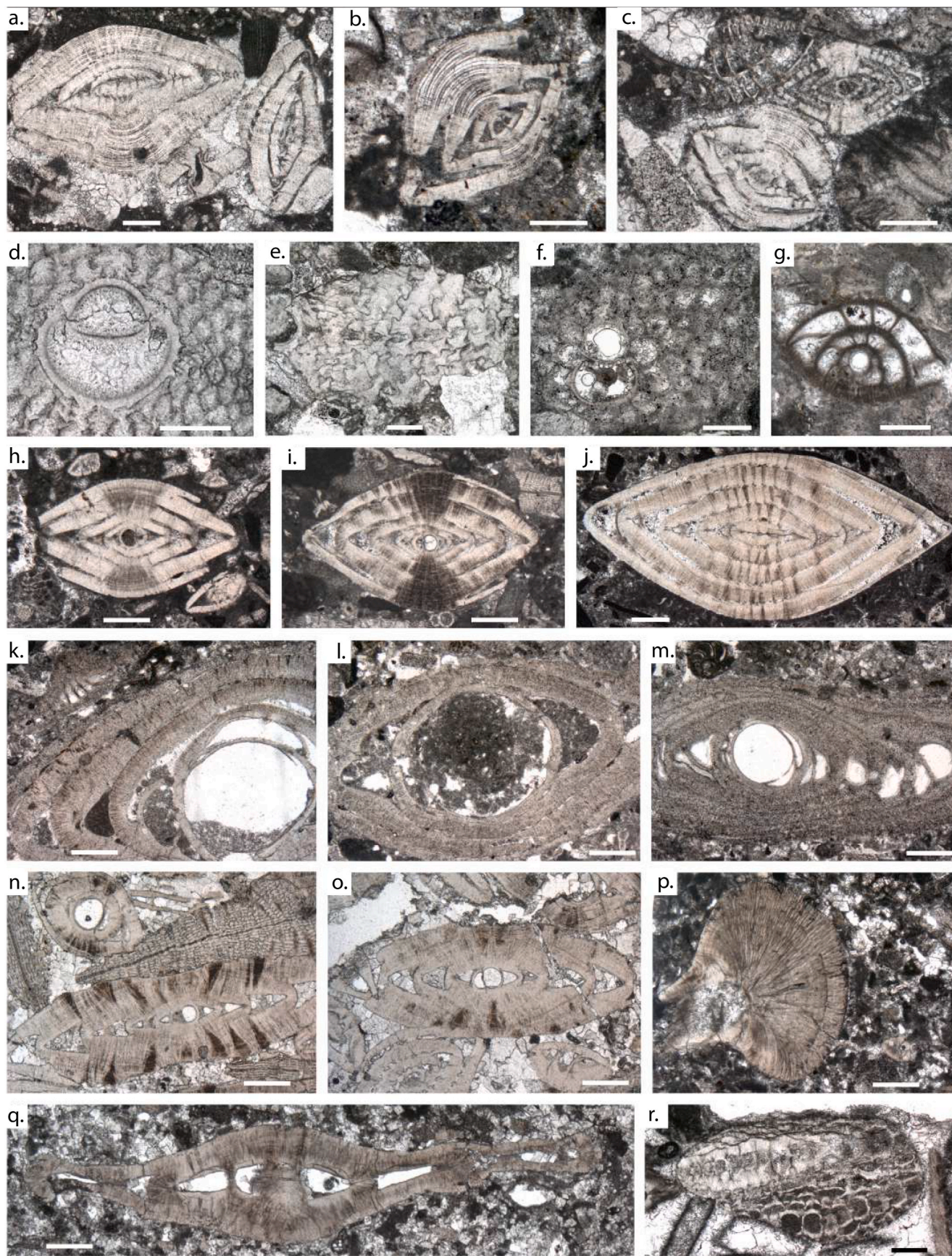
3.1.2. Paleocene – early Eocene—Pple (66–48 Ma)

The Paleocene – early Eocene unit is the thickest unit in the Bolon structure (Figs. 5 and 8) but shows significant lateral variations. It is composed of pale green to ochre-colored marlstones and mudstones interbedded with fine-to medium-grained sandstones and calcarenites, siltstones, and rare grainstones (Fig. 9b). The basal contact to the SW of the Trinitari Fault is gradational with the underlying Senonian unit transitioning from white marlstones to interbedded marlstones and fine-medium grained sandstones with rip up clasts (location 5 in Fig. 5; Fig. 7b). East of the Trinitari Fault, channelized sandstones erode the Senonian and place the Pple unit in direct contact with the Triassic salt (location 6 in Fig. 5), evidenced by reworked bipyramidal quartz in the basal section of the Pple unit. On the NE side of the East Bolon Fault, this unit is deposited directly on top of the Triassic salt (location 4 in Fig. 5).

The thickness of this unit varies laterally, but is thickest in the hanging wall of the Derramador Fault (520 m) in the center of the structure and thins towards both the east and west terminations (250 m east of the Trinitari Fault thinning to only 5 m east of the East Bolon Fault and 260 m west of the Derramador Fault). In the center of the structure, where it is thickest, the bottom 70 m are dominated by fine-to medium-grained sandstones and calcarenites with trough cross bedding, dewatering structures, and ripple laminations interbedded with tan to ochre marlstones with rare grainstones of imbricated nummulitids at the

base. The middle 380 m is composed of mostly green to ochre mudstones with a prominent package of interbedded calcarenites, sandstones in the middle, and the upper 70 m is a gradational change to limestones, calcarenites, sandstones and siltstones with minor interbedded mudstones and sandstones with dewatering structures and iron concretions making a thickening and coarsening upward sequence.

Field relationships suggest that the Pple unit is Paleocene – early Eocene in age. Based on sample P-01 collected near the base of the unit we refine this to early Ypresian (Shallow Benthic Zone (SBZ) 7–8), indicated by the presence of *Assilina* of the groups of *A. spira* and *A. exponens* with species characteristic of the middle-late Ilerdian: *A. arenensis*, *A. pustulosa/pomeroli*, (Appendix 1a; Schaub, 1981; Tosquella and Serra-Kiel, 1998). Therefore, Paleocene aged strata are inferred to be represented by a relatively thin interval of interbedded siltstones and green marlstones that are in continuity with the Senonian deposits and are comparable to the Agost Formation described by Chacón and Martín-Chivelet (2005) in the Agost Basin (~16 km to the east). The upper section is comparable to the latest Paleocene-early Eocene (Alberquilla and Garapacha Formations) described by Guerrero et al. (2014), and the Pinos-Rasa Formation described by Chacón and Martín-Chivelet (2005) near the Pinos diapir (~20 km to the west; Fig. 2a). According to these correlations and the sedimentological features, this unit is interpreted to be deposited as intercalations of hemipelagic and turbiditic beds deposited in a mud rich submarine ramp located below the euphotic zone with water depths ranging from 200 to 500 m (Chacón and Martín-Chivelet, 2005; Martín-Chivelet and Chacón, 2007; García et al., 2015; Martín-Chivelet et al., 2019; Martín-Martín et al., 2020b).



(caption on next page)

Fig. 6. Thin section images in plane light of the main larger foraminifera taxa used for biostratigraphical interpretation. a. *Amphistegina mammilla*, sample Bat-01. b. *Amphistegina bohdanowiczi*, sample O-03. c. *Elphidium* cf. *crispum* (top) and *Amphistegina* sp. (bottom), sample Bol-01. d. *Nephrolepidina* sp., sample R-01. e. *Miogypsina* cf. *M. globulina*, axial section, sample R-02. f. *Miogypsina* cf. *M. mediterranea/cushmani*, equatorial section, sample BURD-04. g. *Heterolepa dutemplei*, sample O-03. h. - j. *Nummulites vascus*, megalospheric forms (h., j.), sample Bol-9-03, and microspheric form (i.), sample Bol-9-08. k. - l. *Nummulites maximus*, sample O-01. m. *Assilina exponens*, sample O-01. n. *Assilina* cf. *pustulosa*, sample P-01. o. *Assilina* cf. *arenensis*, sample P-01. p. and q. *Assilina praespira*, sample Prim-01. r. Reworked clast with miogypsinids, sample Emm-02. Scale bars: a. - g. 200 μ m, h. - r. 500 μ m.

3.1.3. Late Eocene—Pue (48–34 Ma)

The late Eocene unit conformably overlies the Paleocene-early Eocene strata. In the center of the structure, the bottom contact is characterized by a gradational change from calcarenites and mudstones to metric packages of white to light tan packstone-grainstones rich in benthic foraminifera (Fig. 7c). The bottom 115 m of the late Eocene unit is composed of metric beds of nummulite grainstones-packstones interbedded with silty limestones and siltstones, while the upper 120 m is dominated by decametric beds of nummulite packstones and grainstones, with subordinate wackestones and carbonate siltstones. Overall this unit shows a thickening upwards sequence (Fig. 8). The true depositional thickness of this unit is hard to determine due to the erosional nature of the upper contact, nonetheless, as a minimum it attains a thickness of 235 m between the Derramador Fault and the Trinitari Fault and a thickness of only 35 m east of the Trinitari Fault (Fig. 8). The reduced thickness to the NE of the Trinitari Fault (Location 7 in Fig. 5) represents a condensed section compared to the equivalent strata observed to the west of the Trinitari Fault.

The age of this unit is late middle Eocene (Bartonian), indicated by the co-occurrence of *Nummulites maximus*, *Assilina exponens*, *Orbitocypus varians varians*, and *Alveolina* sp. observed in samples O-01, Bol-9-08, and Bol-9-10 (Appendix I; Schaub, 1981; Less et al., 2011; Özcan et al., 2019). The late Eocene unit correlates with the Minano Formation and is interpreted to be deposited in a marine carbonate platform (Guerrera et al., 2014; Martín-Martín et al., 2020a).

3.1.4. Oligocene to earliest Aquitanian—Po1 Po2 and Po3 (~32–~22 Ma)

This unit has three distinct subunits with distinct facies deposited above a minor unconformity with low angle truncations of the underlying upper Eocene unit (Location 8 in Fig. 5). The three subunits are: Po1 limestones; Po2 limestones and marlstones; and Po3 matrix supported conglomerates (Fig. 7d and e). Subunit Po1, located west of the Trinitari Fault, is characterized by decametric tabular bedded packstones and grainstones dominated by benthic foraminiferal communities with corals heads near the top of the section (Figs. 8 and 9). The age of this unit is constrained by the presence of *Nummulites* cf. *vascus*, *Asterigerina* sp., *Heterostegina* sp., and victoriellids. The presence of *N. vascus* indicates an Oligocene age (Rupellian-Chattian: SBZ 21–23) (Fig. 6; Less et al., 2011; Ferrández-Cañadell and Bover-Arnal, 2017).

Subunit Po2 is characterized by a 5-m-thick grey to white marlstone layer at its base and 20 m of decametric packages of thin bedded, dark grey, micritic limestone separated by thin metric intervals of grey massive to laminated marlstones (Figs. 7e and 8). This subunit is only exposed in the central portion of the Bolon structure. Its base is an unconformity that truncates the Po1 subunit down section to the SW whereas this subunit onlaps the unconformity surface at a low angle ($\leq 10^\circ$; Location 9 and 10 in Fig. 5; Fig. 7e). The upper contact is also an unconformity that joins the basal unconformity to the southwest near the peak of El Bolon Mountain. The true depositional thickness of this subunit is hard to determine, but field measurements range from 0 to 25 m in thickness. The age of the Po2 subunit is poorly constrained due to the lack of paleontological data but its age is younger than the sedimentation of the Oligocene Po1 unit and older than the late Burdigalian to early Langhian M1 unit. However, if we consider that it is a low angle onlap with the Po1 subunit, and it is situated below the major late Paleogene-early Miocene unconformity, we can assume a middle-late Oligocene age.

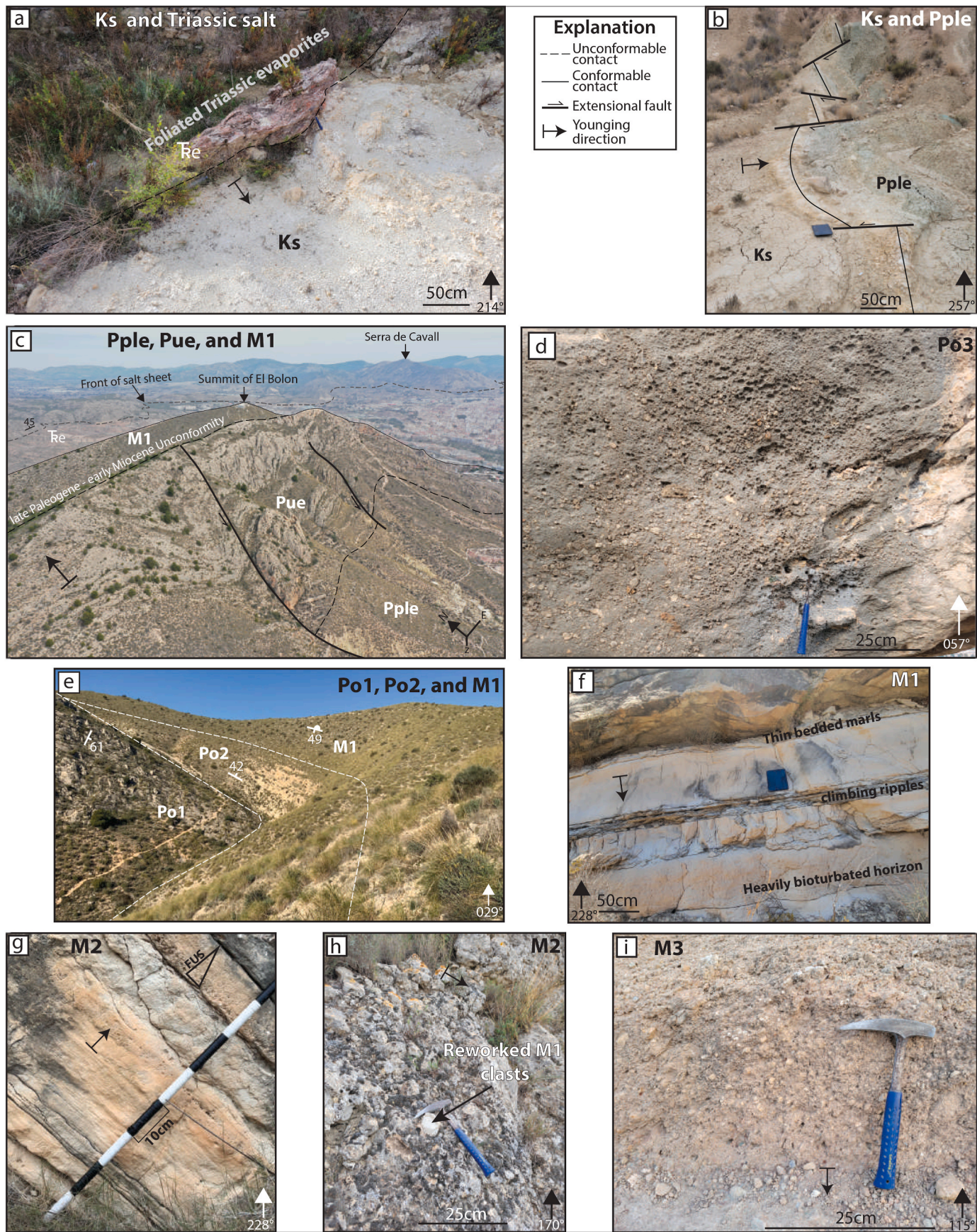
East of the Trinitari Fault, subunit Po3 rests unconformably above the late Eocene (Pue) limestones and is comprised of conglomerates with sub-rounded to well-rounded granule to pebble-size clasts in an angular to subangular, medium-to very coarse-grained bioclastic calcarenite matrix, rich in reworked bipyramidal quartz. The clasts are dominantly composed of micritic limestones and dolostones with subordinate sandstone clasts. The conglomerates sometimes pass to fine-grained and coarse-grained calcarenites where doubly terminated quartz crystals are present but not prevalent (Fig. 7d). The presence of *Amphistegina bohdanowiczi* and *Elphidium* cf. *crispum* with reworked *A. mammilla* indicate a latest Oligocene-early Miocene age (SBZ 23–24) (Ferrández-Cañadell and Bover-Arnal, 2017). This unit correlates laterally with the Oligocene-Aquitania units (Murtas Formation and lateral equivalents) of Guerrero and Martín-Martín (2014), and Martín-Martín et al. (2020a). In the Bolon structure these units are characterized by shallow water benthic assemblages along with corals, shell fragments, and shark teeth that allow us to interpret an inner ramp to external platform environment likely in a barrier reef transitioning upwards into a deep marine-outer ramp to slope type setting and to the east of this fault the sedimentological features suggest a deepwater fan delta depositional environment.

3.1.5. Latest Burdigalian – early Langhian—M1 (20–15 Ma)

The M1 unit rests on a major erosional unconformity (i.e. the late Paleogene-early Miocene unconformity) that truncates bedding across the structure, but is most pronounced SW of the Trinitari Fault where truncations of the Po1, Po2, and Pue units indicate significantly more erosion on the western side (Location 11 in Fig. 5). It is incorporated into, and onlaps the Po3 subunit in the antiformal syncline on the NE side of the structure before truncating against the East Bolon Fault. This unit is at least 260 m thick and consists of a thinning upwards sequence of 10–40 cm thick tabular beds of very fine-grained calcarenites to silty limestones interbedded with subordinate white to grey marlstones that thicken up section (Fig. 7e and f and 10). In thin section, this unit is a planktonic foraminifera packstone with rare small benthic forms (Figs. 9e and 11a).

In map view, the M1 unit onlaps the late Paleogene-early Miocene unconformity from west to east, but also onlaps into the map view plane. The multidirectional onlaps suggest a remnant paleo-topography present along this unconformity surface. Apart from the progressive thickening of the unit towards the west (Fig. 10), a local thickened depocenter (~100 m) is observed between the Trinitari and Reis Mags Faults where strata show a wedge-shaped, drape fold type geometry (Location 12 in Fig. 5). The upper contact is another erosional unconformity that cuts down section to the NE (Location 13 in Figs. 5 and 10).

The age of this unit is constrained by *Fohsella peripheroacuta* that indicates a Langhian to early Serravallian age (Kennett and Srinivasan, 1983; Wade et al., 2011) and *Globigerinoides diminutus* giving a Burdigalian-early Langhian range (Appendix Id, samples M01 and Bol-01 locations in Fig. 5). The assemblage is dominated by non-keeled planktonic foraminifera with only rare small calcareous benthic forms suggesting deposition in deep outer ramp or shelf to upper bathyal environments. This age indicates that in the southwest, the basal section is time-equivalent to the Congost Formation described by Martín-Martín et al. (2018b) and Martín-Martín et al. (2020b) that is interpreted as shallow marine deposits representing beach to marine platform to external platform realms, whereas the upper part of the M1 unit is correlated with the lower member of the First Tap Formation (Jabaloy



(caption on next page)

Fig. 7. Field images from the Bolon Secondary Minibasin. a. Senonian marlstones unconformably deposited above the Triassic evaporites. b. Small extensional faults at the Senonian – Paleocene lower Eocene contact. c. Drone image showing the conformable transition between the lower Eocene and upper Eocene units with the thickening upward sequence of the upper Eocene units well represented in the left side of the image. Notice the extensional faults are truncated by the late Paleogene early Miocene unconformity. d. Outcrop photo of matrix supported conglomerates of the Po3 unit. e. Field image showing the onlapping relationships of the Po1, Po2, and M1 units just west of the Trinitari Fault. f. Overturned M1 stratigraphy with interbedded marlstones. g. Fining upwards sequences in the M2 unit with bi-pyramidal quartz in the coarse-grained horizons. h. Conglomerates composed predominantly of clasts of M1 stratigraphy within the M2 unit. i. Outcrop of poorly lithified M3 breccias 1.5 m from the upper salt sediment interface.

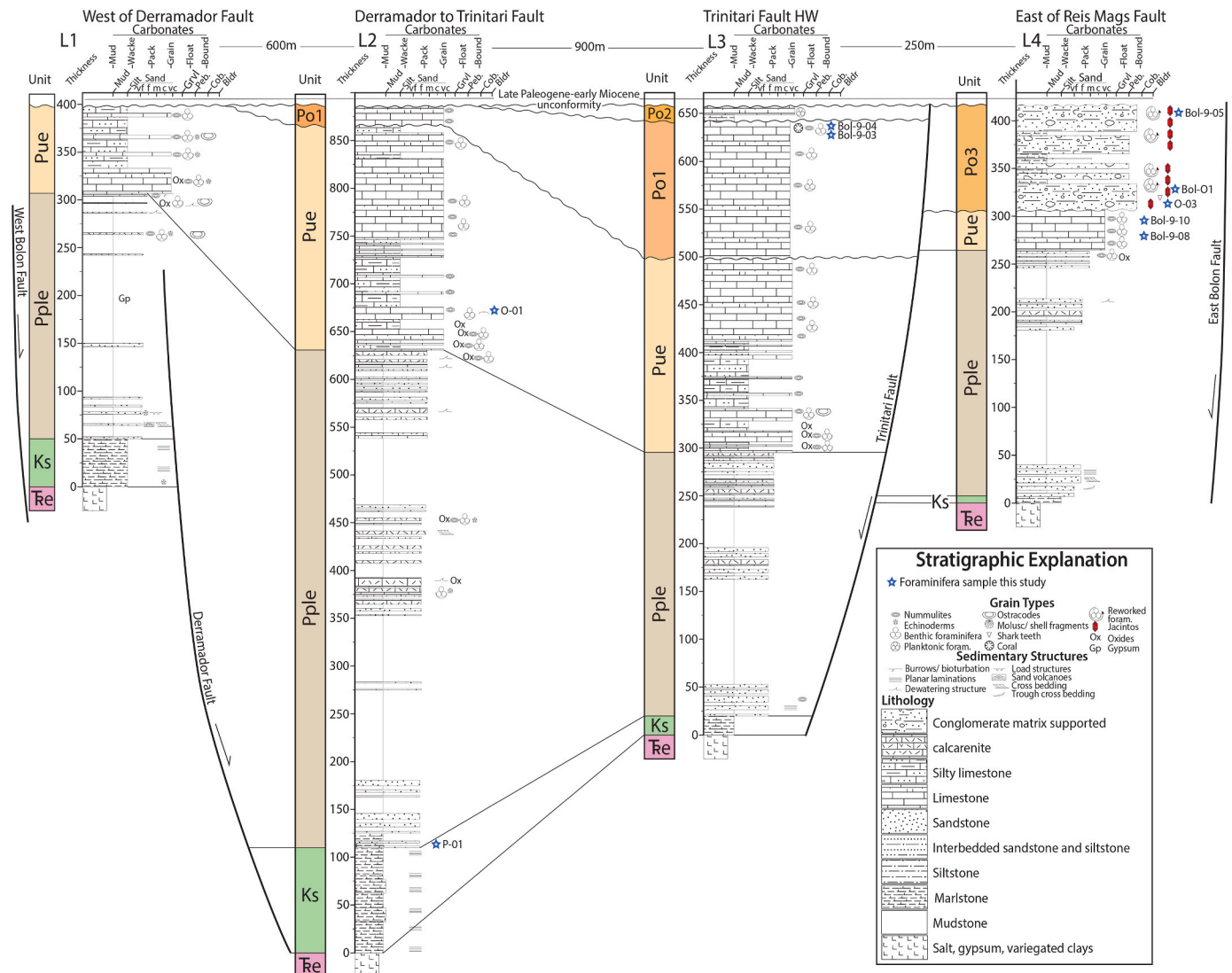


Fig. 8. Stratigraphic section of the Oligocene and older units exposed in the Bolon structure. The locations of measured sections are shown in Fig. 13. Sections are correlated along the late Paleogene – early Miocene unconformity.

Sánchez et al., 2019). In the Bolon structure these deposits represent a deep marine, outer ramp setting as evidenced by the marl dominated facies and planktonic foraminifera packstones.

3.1.6. Langhian to early Serravallian—M2 (20–15 Ma)

The Langhian to early Serravallian M2 unit onlaps the underlying erosional unconformity at the top of the M1 unit (Location 14 in Fig. 5). The onlapping beds are tabular, ranging in thickness from 5 to 20 cm comprising medium-to coarse-grained calcarenites, microconglomerates, and subordinate matrix supported conglomerates with well-rounded pebbles. Most beds comprise fining upward sequences grading from microconglomerates rich in diapir derived detritus (i.e. bi-pyramidal quartz and red and green mudstone clasts) at the bottom to fine-to medium-grained calcarenites at the top (Figs. 6r, 7g and 10, and

11b). On the NW side of the structure, there is a distinct monomictic, clast supported conglomerate horizon that is characterized by sub-angular to moderately rounded M1 cobbles indicating the erosional stripping and reworking of an adjoining topographic high composed of M1 stratigraphy (Figs. 7h and 15a). The M2 unit is thickest (48 m) near the center of the structure and pinches out both to the SW and to the NE (Fig. 10).

The presence of *Fohsella peripheroacuta* indicates a Langhian to early Serravallian age correlating with the lower Member of the First Tap Formation (Fig. 3; Appendix 1a sample EMM-01 and EMM-02; Kennett and Srinivasan, 1983; Vera, 2004; Wade et al., 2011). The foraminiferal-bioclastic packstone facies with common planktonic foraminifera and bioclastic debris of macrofossils suggest deposition in outer ramp/shelf with repetitive turbidity currents sourced from

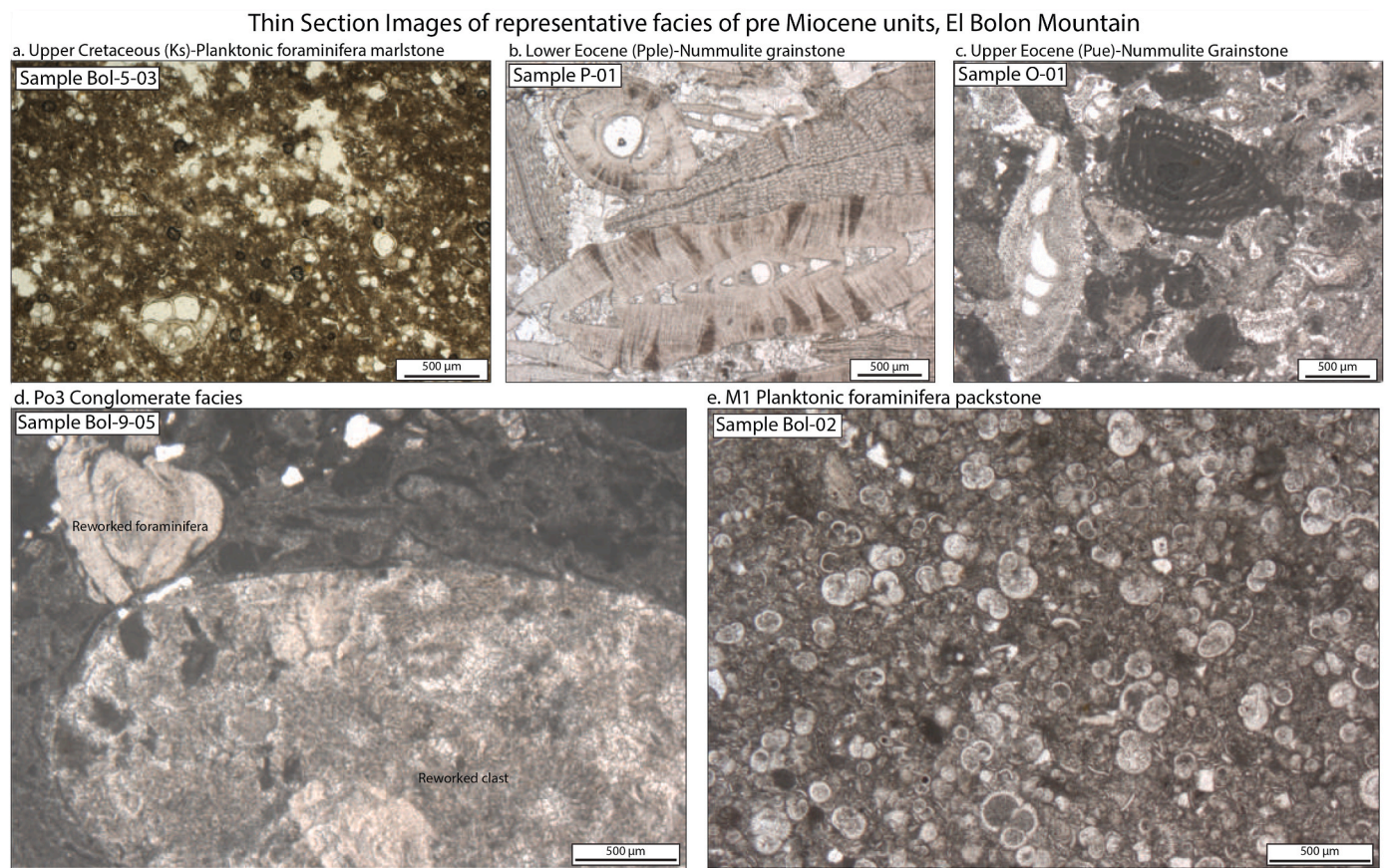


Fig. 9. Thin section images in plane light from the Pre-Miocene units exposed in the Bolon Secondary Minibasin. (a) Sample Bol-5-03 uppermost Cretaceous (Senonian) marlstones with many planktonic foraminifera. (b) Sample P-01, lower Eocene Nummulite grainstone. Notice the imbrication of the nummulites and relatively well-preserved nature of all specimens, see appendix 1 for an analysis of the benthic foraminifera. (c) Sample O-01, upper Eocene grainstone dominated by different species of Nummulites. (d) Sample Bol-9-05 Po3 Oligocene conglomerates with well-rounded clasts and a very dark calcite matrix. (e) Sample Bol-02 planktonic foraminifera grainstone from the M1 unit. See [appendix 1](#) for a complete description of the foraminiferal analysis.

bathymetrically elevated Triassic salt structures.

3.1.7. Langhian to early Serravallian—M3 (20–15 Ma)

The M3 unit is the youngest unit exposed in the Bolon structure and semi-conformably overlies the M2 unit in the center of the basin (i.e. low angle $<10^\circ$ onlaps with the underlying unconformity; **Location 15 in Fig. 5**). To the SW it onlaps the M1 unit (**Location 16 in Fig. 5**) and in the NE it onlaps the M1 and M2 units and presents a growth wedge geometry at the footwall of the East Bolon Fault with associated talus breccias (**Location 17 in Fig. 5; Fig. 10**). Overall, this unit shows a coarsening upwards trend along with higher concentrations of Triassic salt clasts near the salt-sediment contact, and generally coarser grained deposits towards the NE (**Fig. 10**). In the west central section, the bottom of this unit is composed of well indurated coarse-grained calcarenites to microconglomerates, overlain by layers of grey-purple mudstones with interbedded and reworked Triassic salt blocks. At the top of the section are isolated bodies of breccias rich in reworked diapir derived detritus. The top contact is unconformable with the Triassic salt and the strike of bedding is sub-parallel to the salt contact with consistent truncations into the Triassic salt towards the SW. Along the upper salt-sediment contact, three distinct wedges of breccias rich in diapir derived detritus crop out (**Fig. 7i**). These wedges pinch out towards the NE and truncate against the salt to the SW (**Location 18 in Figs. 5 and 16b**).

The presence of reworked *Miogypsina* cf. *globulina*, *M. intermedia/mediterranea*, and *Nephrolepidina* indicates a Burdigalian or younger age (**Appendix I-Sample R-01, 02; Drooger, 1993; Cahuzac and Poignant, 1997; Verrubbi and Schiavinotto, 2005; Özcan et al., 2009, 2019; Fer-randini et al., 2010**). Based on the stratigraphic position, the M3 unit

should be placed in the late Langhian-early Serravallian and would correlate with the upper part of the lower member of the First Tap Formation (**Fig. 3; Vera, 2004; Martín-Martín et al., 2020b**). We interpret the M3 unit to be deposited in an outer ramp/shelf with the breccias sourced from the encroaching and bathymetrically elevated Triassic salt and deposited as mass-transport deposits.

3.2. Structure

The Bolon structure, in map view, is formed by predominantly NE-striking beds with a northwestward's stratigraphic polarity. Throughout the structure, bedding maintains high dip values, ranging from overturned beds dipping $45\text{--}80^\circ$ towards the WSW to SE in the southeast, to normal polarity strata dipping $50\text{--}60^\circ$ to the N-NW in the northwest (**Fig. 5**). Therefore, the map view exposure can be considered as an oblique cross section. To accurately describe the small scale (100's of meters) structures within the kilometeric Bolon structure, we divide the area into seven structural domains based on stereographic analysis, salt-related folding, and stratigraphic architecture (**Fig. 12b**). Structural domains 1 and 2 delineate the central minibasin stratigraphy and exclude any secondary folds (i.e. the antiform of SD5). Structural domain 3 and 4 are Miocene growth wedges and structural domain 5 comprises the fault related folding next to the Trinitari Fault. The western detail and eastern detail are described separately because of the intricate salt-sediment relationships along the flanks of the structure. These last two domains denote different relationships with the salt structure that represent previously developed diapiric contacts. After the description of the major faults, we describe the different salt-

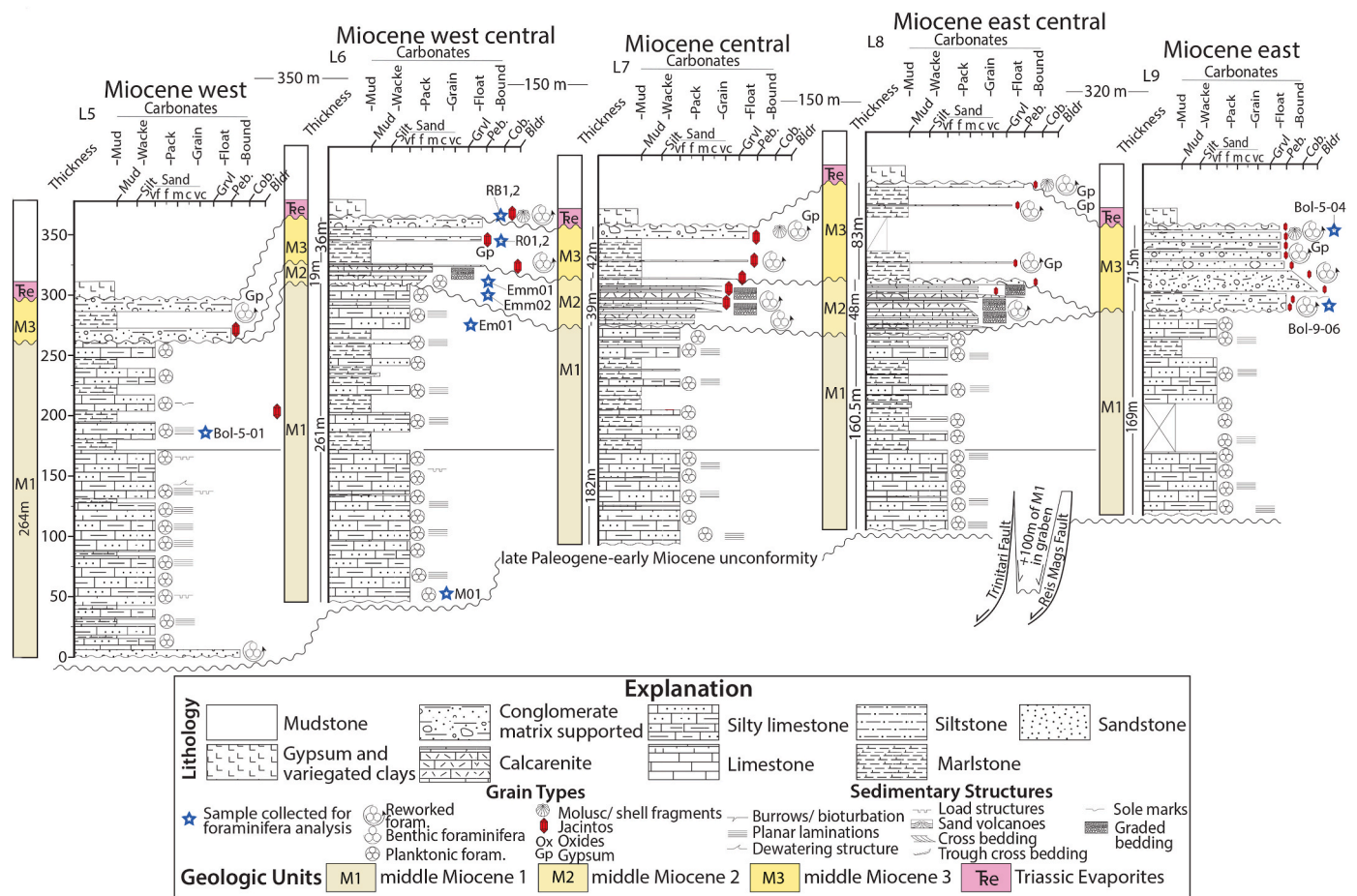


Fig. 10. Stratigraphic sections of the Miocene stratigraphy of the Bolon Secondary Minibasin. The location of each measured section is shown in Fig. 13. The sections are correlated along the first laterally continuous marlstone bed in the M1 unit. Notice the M1 unit infilling the previously developed paleotopography on top of the late Paleogene – early Miocene unconformity.

sediment interfaces that are: the top salt contact—SE, lateral salt contacts—SW and NE, and a base salt contact—NW, which contain smaller structures that are vital for a comprehensive interpretation of the Bolon structure.

Stereographic projections of the Paleogene bedding data (Fig. 12, SD1) indicate that the Bolon structure should be considered as an overturned basin with a broad (1.5 km) hinge zone i.e. a doubly plunging, downward facing, gentle antiformal syncline, verging towards the northwest. The limbs of the fold are not planar structures but are short wavelength folds (200–300 m) that bend the stratigraphy subparallel to the salt-sediment interface (Location A in Fig. 5). Therefore, **the structure is not a simple concentric fold**, in fact most units are separated by unconformities that juxtapose very different bedding attitudes that often mark changes in the position of the axial plane of the kilometer fold.

Strata in the Bolon structure truncate into, or are overlain by salt. Specifically, on the SW and NE sides of the structure, juxtaposed with the salt are small scale folds that we call “terminal folds.” On the SW side, the terminal folds are a series of stacked asymmetric antiformal synclines characterized by moderately SW plunging fold axis with close-gentle interlimb angles. Each fold has a short western limb where strata truncate at an unconformity, and a long eastern limb where strata become conformable away from the salt contact (Area A in Fig. 5). The terminal folds in the NE side of the structure are separated by the late Paleogene-early Miocene unconformity (Fig. 5, and Fig. 15). These terminal folds, and the salt-sediment contacts are described in detail in section 3.2.2. Below the late Paleogene-early Miocene unconformity is a

steeply south plunging, antiformal syncline incorporating the M1, Po3, and late Eocene units with a gentle interlimb angle (Area B in Figs. 5, Figs. 12 and 15). Above this unconformity there is a moderately north plunging asymmetric syncline in the M3 unit with an open interlimb angle and a short eastern limb that forms a stratigraphic wedge against the East Bolon Fault (Area C in Fig. 5). The structures and terminal folds near the salt sediment interface are described in detail in the salt-sediment interface sections.

Other prominent folds include: (1) the hectometric asymmetric anticlinal fold in SD5 the center of the Bolon structure that is characterized by a north to northeast steeply plunging fold axis with a long western limb and a short eastern limb that terminates in a smaller, synclinal drag fold adjacent to the Trinitari Fault (Area D in Figs. 5 and 12, SD5); and (2) an antiformal syncline in SD4 with a gentle interlimb angle and a moderate plunge to the W-SW with a thickened section of lower M1 unit between the Trinitari and the Reis Mags Faults (Location 12 in Figs. 5 and 12, SD4).

3.2.1. Major faults

In the present map view, the strata of the BSM are affected by multiple faults with a hectometric to decametric strike separation (Fig. 5). Based on the bedding cutoffs and the map view geometries, these faults are all close to vertical ($>70^\circ$) and have inconsistent orientations that vary across the structure; they are mostly NNW-SSE at the northeastern end, NNE-SSW to NE-SW at the southwestern end, and both NNE-SSW and NW-SE in the central area of the structure (Fig. 13).

The West Bolon Fault, Cappuccino Fault, Derramador Fault, upper

Thin Section Images of Miocene units of the Bolon Structure

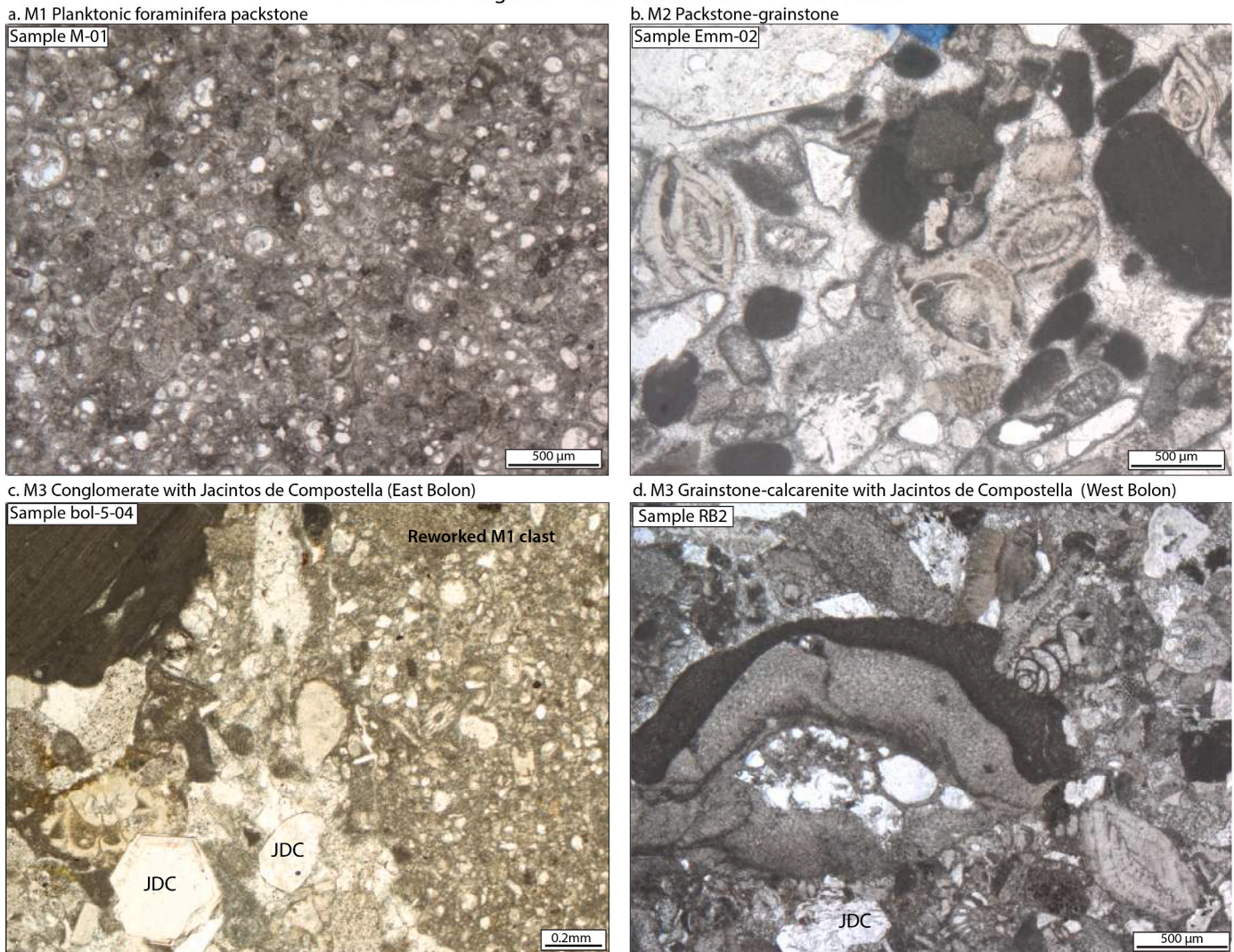


Fig. 11. Thin section images in plane light showing the lithological characteristics of the Miocene units in the Bolon structure. a. Planktonic foraminiferal grainstone from the M1 unit. b. Packstone-grainstone from the M2 unit with many opaque clasts derived from the Triassic evaporites. c. Sample bol-5-04 from the M3 conglomerates with Jacintos de Compostella (JDC) i.e. bipyramidal quartz, reworked clasts of the M1 unit, reworked foraminifera, and large shell fragments. d. Sample RB2 grainstones-conglomerates from the M3 unit.

Eocene Faults, Trinitari Fault, Reis Mags Faults, and the East Bolon Fault have an extensional sense of slip that presents a significant strike separation on the map, which is sinistral where the faults dip towards the west and dextral where the faults dip towards the east (Figs. 5 and 13). The West Bolon Fault, Cappuccino Fault, Derramador Fault, and the Trinitari Fault affect the Senonian and Paleocene-early Eocene units. Based on the growth geometries and sealing layers the activation of these faults become progressively younger towards the NE with the Trinitari Fault being the youngest of this set. The oldest fault, the Derramador Fault, affects the Senonian and lower part of the early Eocene section and is characterized by hectometric displacements where there is a marked thickening of the Senonian and Paleocene-early Eocene succession in the hanging wall block.

The Trinitari Fault, located in the NE portion of the Bolon structure has an apparent map-view offset of 225–275 m with the west block moving south. The Trinitari Fault cuts the Senonian through Oligocene 2 units but only the upper Eocene and Oligocene 1 units show a growth wedge geometry (Fig. 5 near location D). Above the Trinitari Fault two conjugate faults form a graben in the basal M1 unit. Here the basal M1 section, when restored to its original orientation, would assume a drape

fold geometry with the Reis Mags Faults indicating their syntectonic character. The Trinitari and Reis Mags Faults, and their linked synthetic and antithetic counterparts, are sealed by the middle M1 unit and superjacent units. These tectono-stratigraphic relationships indicate that the Trinitari Fault and the Reis Mags Faults were active at least during the Senonian to the Oligocene 1 and they stopped functioning during the deposition of the middle M1 unit (Fig. 15a and b). Faults in the central part of the Bolon structure cutting the late Eocene succession, have smaller offsets (10–75 m) and are truncated by the late Paleogene-early Miocene unconformity (**upper Eocene Faults** in Figs. 5 and 13). These faults sole out into the middle-upper mudstone interval in the Paleocene-early Eocene unit. The faults have a consistent orientation with apparent sinistral movement in faults that dip to the west and apparent dextral movement in the faults that dip to the east (Figs. 5 and 13). These faults are related to the listric movement of the Trinitari Fault and accommodate its rotational deformation (Ellis and McClay, 1988; Brun and Mauduit, 2008).

The East Bolon Fault has an apparent offset of 100–130 m and consists of a master fault with synthetic and antithetic faults developed in the footwall block (Figs. 5 and 13). In the footwall the sediments show a

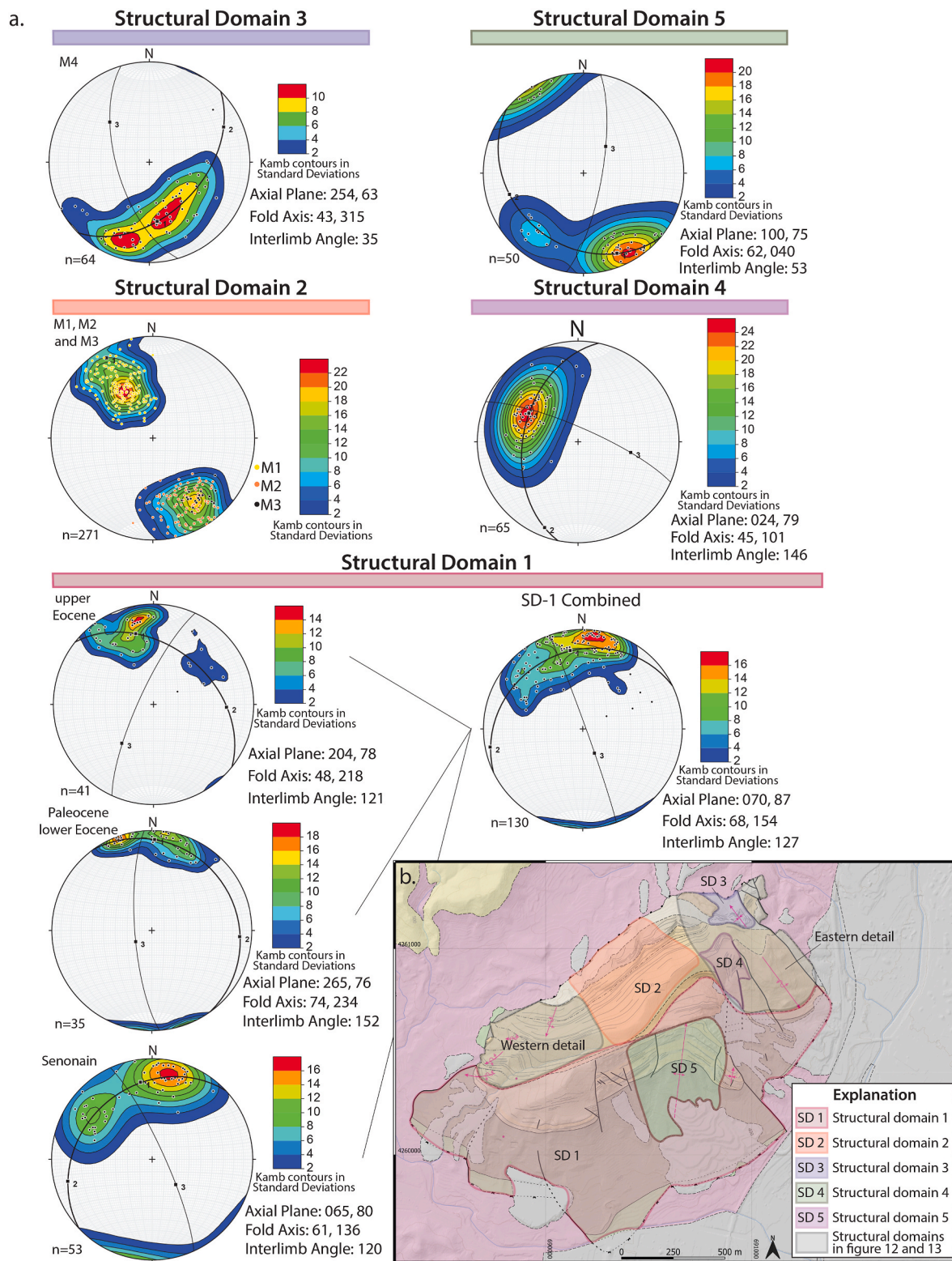


Fig. 12. a. Lower hemisphere equal area stereoplots showing the poles to bedding data for the different structural domains within the Bolon structure. Calculated fold axis and axial plane are plotted where appropriate. Structural Domain 3 (SD3): fault bounded synclinal growth wedge plotted with Kamb contours, contour interval (c.i.) 2, significance level (s) 2. Structural Domain 2 (SD2): Miocene units through the center of the structure with Kamb contours, c.i. 2, s. 3. Structural Domain 5 (SD5): Eocene units to the west of the Trinitary Fault with Kamb contours, c.i. 2, s. 1. Structural Domain 4 (SD4): Miocene units in small synclinal growth wedge between Trinitari and Reis Mags Faults with Kamb contours, c.i. 2, s. 1. Structural Domain 1 (SD1): left, stereonet with bedding plotted for the upper Eocene (c.i. 2, s. 1), lower Eocene (c.i. 2, s. 1), and Senonian (c.i. 2, s. 3) units without near fault or near salt bedding variations. Right is the combined stereonet with all data from the Senonian through the upper Eocene units, c.i. 2, s. 2. b. Structural domains of the Bolon structure used for stereonet analysis.

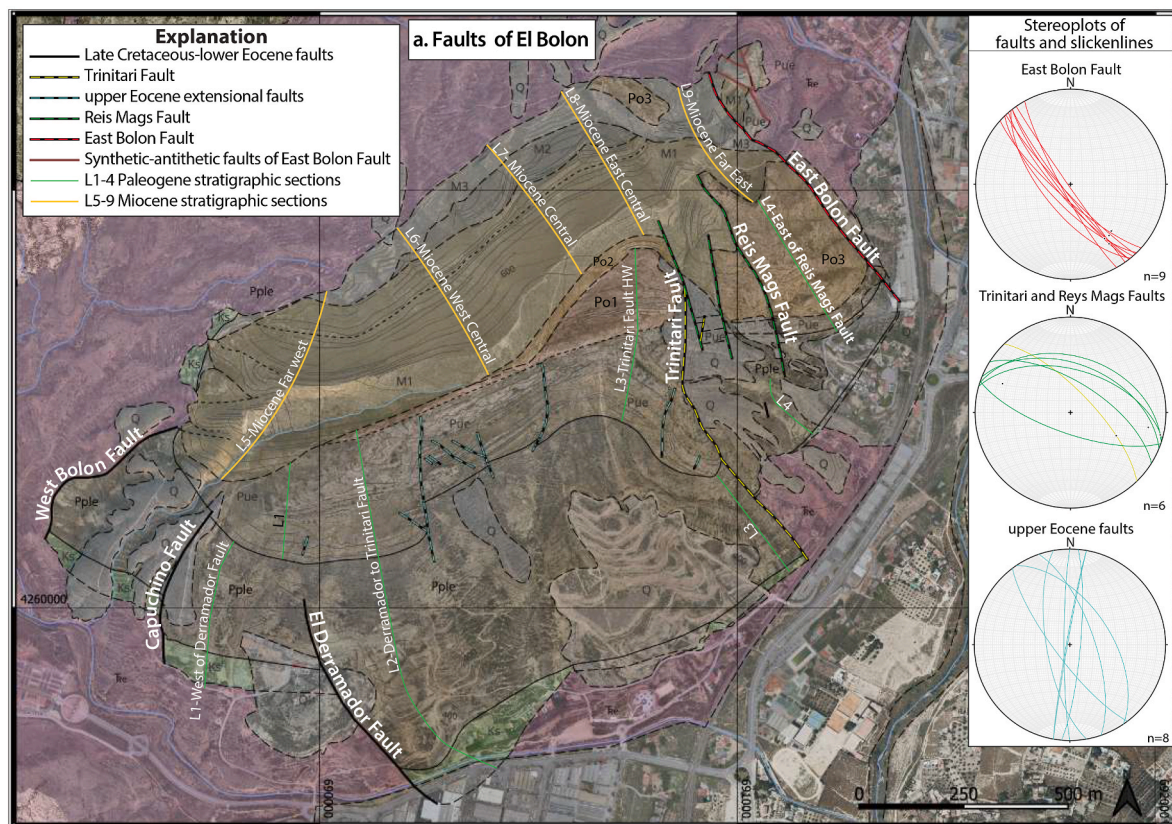


Fig. 13. Orthophoto from 3-D model of the Bolon Secondary Minibasin with the geologic units shown and the main faults labeled. Locations of measured sections are shown as green lines for the pre-Miocene stratigraphic sections and yellow lines for the Miocene stratigraphic sections. Lower hemisphere equal area stereoplots of fault planes (great circles) and slickenfibers (lines) are shown in the box on the right with the colors correlating to the different fault sets identified on the map. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)

growth wedge geometry with the East Bolon Fault but only minorly displace the NW salt-sediment contact. Therefore, these faults were active during the M3 deposition and they stopped functioning shortly after M3 deposition (Figs. 5 and 13).

Finally, it should be noted that all the major faults appear synchronous with sedimentation of the Senonian-Cenozoic units. Therefore, they formed prior to folding. When bedding is restored to horizontal all the observed faults show extensional geometries with dips between 50° and 90° (Figs. 5 and 13).

3.2.2. Southeastern salt-sediment interface-top salt contact

The southeastern salt-sediment contact coincides with the base of the sedimentary successions of the Bolon structure and is characterized by an undulating salt-sediment contact with smooth ridges and troughs onto which the Senonian strata are concordantly deposited (Fig. 5). There is a pronounced thickening of the Senonian unit in the hanging wall of the Derramador Fault indicating its syn-depositional activity (Fig. 5). In the footwall of the East Bolon Fault the Senonian unit is absent and the Paleocene-early Eocene turbidites are concordantly deposited above the salt-sediment interface. Additionally, from the Derramador Fault to the NE the Senonian unit becomes progressively thinner and is completely absent in a small section in the footwall of the Trinitari Fault (Location 6 in Fig. 5).

3.2.3. Southwestern salt-sediment interface-lateral salt contact

The southwestern salt sediment interface is characterized by overturned stratigraphy and five decametric to hectometric folds (HK 1–5). These folds have axial traces that are displaced along angular unconformities that truncate the short limb (western) of the folds but become conformable in the longer limb further into the basin (Figs. 5

and 14a). These characteristics suggest a halokinetic origin for these folds, therefore, we use the methodology from Hearon IV et al. (2015) to constrain the dip of the salt sediment interface and the direction of salt advance during halokinetic folding. This methodology relies on basic halokinetic fold theory where drape folds develop around the elevated diapir, therefore the fold axis is perpendicular to the direction of salt advance, or halokinetic breakthrough (Hearon IV et al., 2015). Fig. 14b provides a table of the fold axial planes, fold axis, interlimb angles, halokinetic truncation angle, type of halokinetic sequence, and type of minibasin tectonostratigraphic sequence (MTS; Giles and Rowan, 2012; Rowan and Giles, 2023).

Folds HK1 and HK2 are comprised of the basal M1 strata and are capped by unconformities with small debrite wedges rich in diapir derived detritus just above the unconformity surface (Fig. 14c step 2 and 3). In HK3 there are multiple debrite intervals within the very fine-grained sandy limestone horizons. They are characterized by angular pebble-to-cobble-size clasts of nummulite grainstones and M1 strata in a coarse matrix and arranged in lobate tongues. These debrites become larger up section culminating in a decametric mass transport deposit (MTD) that truncates the western limb of the fold (Location 1 in Fig. 14a and c step 3). The MTD is characterized by breccias with angular to subangular clasts of the M1 and late Eocene carbonates in a very coarse-grained matrix rich in diapir derived detritus. The western contact of this MTD is a decametric outcrop of Senonian marlstones and early Eocene mudstones with bedding attitudes subparallel to the salt sediment contact. Samples from the bioclastic breccia of the MTD show a mixed assemblage of Eocene and Langhian foraminifera suggesting an origin due to mass transport processes sourced from nearby reliefs of Triassic salt and Eocene rocks.

The uppermost part of the M1 unit is characterized by another

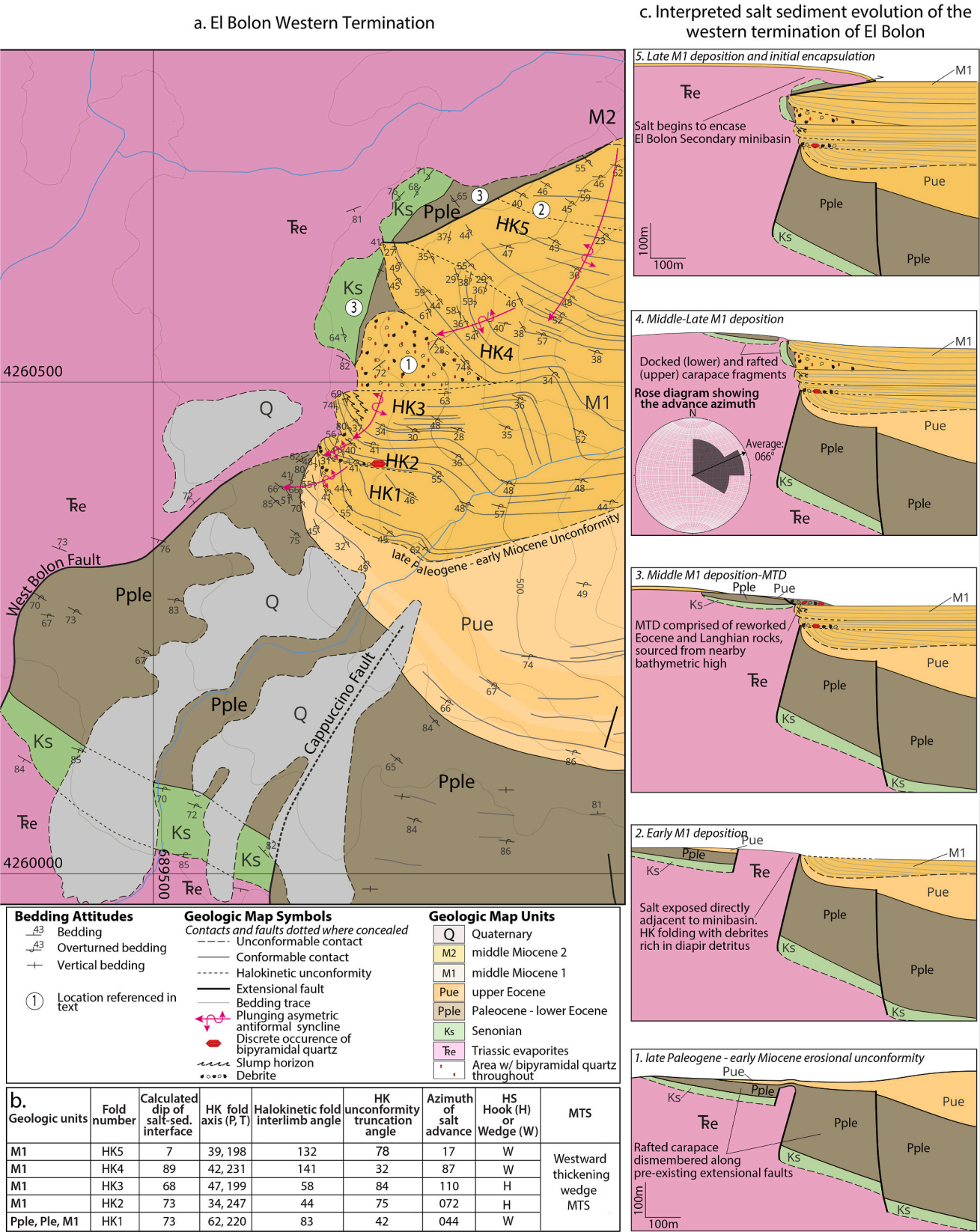
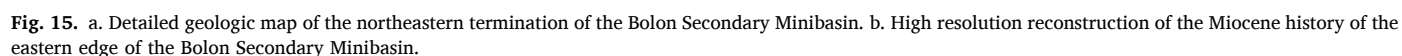


Fig. 14. a. Detailed geologic map of the southwestern termination of the Bolon Secondary Minibasin. Colors and units are the same as in Fig. 5. b. Table of calculated halokinetic fold geometries and interpreted style of halokinetic folding. c. Schematic high-resolution reconstruction of the western salt sediment interface. The rose diagram in the frame of stage 4 shows the average calculated advance direction for the different halokinetic folds. (For interpretation of the references to color in this figure legend, the reader is referred to the Web version of this article.)



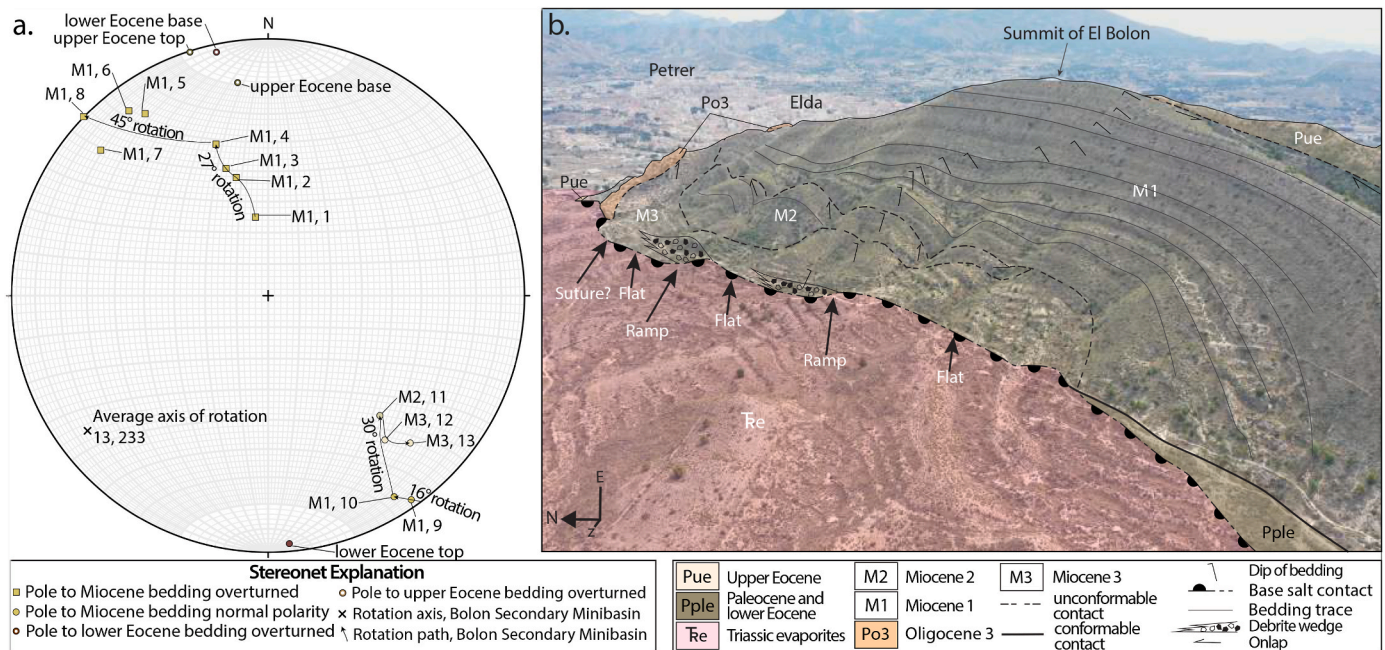


Fig. 16. a. Lower hemisphere equal area Stereonet plot showing poles to bedding data collected through the center of the Miocene section. Calculated rotation axis is shown by the x in the stereonet plot with the degrees of rotation shown between points. b. Drone image with the Bolon geologic map and northern salt-sediment contact depicted in detail. Notice the wedges of debrites associated with the ramping sections and how they pinch out towards the NW.

antiformal syncline HK4 (Fig. 14c step 4) and then HK5 (Fig. 14c step 5) that is a larger wavelength hectometric antiformal syncline. These gentle folds extend ~450 m from the salt sediment interface. Halokinetic fold analysis indicates that the salt sediment contact was dipping seven degrees towards the salt, thus constituting a ramp-type geometry that cuts up through the stratigraphic section towards the northeast before intersecting the M1-M2 unconformity. This contact truncates bedding at 82° with bed lengths decreasing up section (location 2 in Fig. 14). Halokinetic fold 4 and HK 5 are unconformably mantled by the early Eocene and Senonian units that strike subparallel to the salt-sediment contact (Location A, rose diagram of the advance directions are shown in Fig. 14c step 4 to visualize the range of advance directions calculated from the different halokinetic folds.

3.2.4. Northeastern salt-sediment interface-lateral salt contact

The northeastern salt sediment interface is characterized by overturned stratigraphy in the early Eocene to M1 unit with a rapid change in bedding attitudes located across the late Paleogene-early Miocene unconformity (Fig. 15a). The early Eocene-M1 unit forms an asymmetric antiformal syncline with a long eastern limb and a short western limb. Most bedding attitudes are overturned and dip to the south between 60 and 80° in the early and late Eocene units while dip values become progressively more overturned towards the north where the M1 unit is characterized by 40–45° overturned strata (Fig. 15a). The eastern edge is truncated by the East Bolon Fault with an orientation of (241, 82), approximately parallel to the salt sediment interface (Fig. 15a). This fault cuts all the stratigraphy and minorly displaces the upper salt sediment contact in the northeast.

The stratigraphy above the basal M3 unconformity dips towards the north ranging from steeply dipping to near vertical attitudes near the late Paleogene-early Miocene unconformity and passing to moderately north dipping 45–65° normal polarity strata towards the northern salt sediment contact. Below the basal M3 unconformity in the M2 unit there is a distinct monomictic conglomerate of M1 strata (Location 1 in Fig. 15a). Stratigraphically above the basal M3 unconformity, the M3 unit has an asymmetric synclinal geometry with a short eastern limb that is vertical to overturned with strata becoming subparallel to the East

Bolon Fault. These strata show a folded-wedge type geometry indicating that the East Bolon Fault was active during M3 deposition. At the base of the wedge there are angular boulders of M1 strata (Location 2 in Fig. 15a). Additionally, there are numerous decametric asymmetric folds at the base of the M3 unit with a consistent vergence towards the east-northeast that are interpreted as slumps (25 m east of Location 2 in Fig. 15a).

In Fig. 15b we provide a high-resolution schematic reconstruction of the northeastern salt-sediment interface. The reconstruction is constrained by sedimentological and structural field relationships. Specifically, we highlight the repeated episodes of inflation and subsidence that occurred throughout Miocene deposition and evidenced by unconformities, the reworked M1 strata, and debrites.

3.2.5. Northwestern salt-sediment interface-base salt contact

The northwestern, and youngest salt-sediment contact is subparallel to slightly angular in relation to bedding attitudes within the M3 unit, which have a normal polarity and dip towards the north between 55° and 70° (Figs. 5 and 16b). The contact trends parallel to and sometimes slightly oblique to the bedding trends within the M3 unit. Throughout this section there are three distinct ramping zones where the contact moves up section separated by regions where the salt contact and bedding dips are parallel (ramps and flats in Fig. 16b). The ramps extend between 20 and 50 m laterally, whereas the flats extend 170 m laterally. The ramping zones are spatially correlated with debrite wedges in the M3 unit that pinch out towards the northeast (Fig. 16b). The debrite horizons are characterized by poorly sorted pebble to cobble size angular-subrounded clasts with a high proportion of diapir derived detritus. The concentration of diapir derived detritus within the conglomerate horizons generally increases up section and is highest adjacent to the salt sediment contact.

4. Interpretation

From the above described stratigraphic and structural data, we interpret the evolution of the Bolon structure to be an encased secondary minibasin that evolved through nine major stages starting at the end of

Table 1

Summary of the key observations used in the structural interpretation of salt tectonic phases in The Bolon Secondary Minibasin. Information is summarized from section 3.2 and 3.3.

Unit	Key Observations	Salt Tectonics Interpretation
Senonian	- Rugose contact with underlying Triassic salt (top salt) - Tabular, thin to medium bedded marlstones - Condensed stratigraphic sequence compared to regional - Thin footwall thick in hanging wall (Derramador and Capuccino) - Gradual thinning towards the NE - No diapir derived detritus	- Low energy, hemipelagic basin environment. - Pelagic rain depositing on salt top topography - Local extensional faulting - Early salt exposure followed by salt burial - Burry the pre-existing salt structure
Paleocene-early Eocene	- Generally conformable and gradational bottom contact - Unconformable over the Triassic salt in the footwall of the East Bolon Fault - Tabular bedded, hemipelagic, deepwater turbidites with channelized sandstones - Non-condensed stratigraphic sequence compared with regional thickness - Thicker in hanging wall of Derramador Fault - Overall thinning towards the NE - SW salt-sediment contact is a fault surface grading into diapiric contact at the top. - Diapir derived detritus present in turbidite channels contacting the Triassic salts in the Trinitari fault footwall	- Mud rich submarine ramp environment - Synsedimentary extensional faulting - Only local and ephemeral minor salt exposure in the footwall of the Trinitari Fault - Generally, no salt exposed except for east of the Trinitari fault at the base of the unit and east of the East Bolon Fault where the Triassic salts are buried by a thin interval of mudstones and sandstones - Migration of salt towards the NE - Salt was exposed, then buried in the NE domains
Late Eocene	- Gradational bottom contact - Tabular, wedge, and lenticular nummulite limestones - Thickening and coarsening upwards sequence - Thick in hanging wall of Trinitari Fault thin in footwall - No diapir derived detritus	- Carbonate platform, marine internal to external platform realm - Extensional deformation moves to the east Trinitari Fault - Salt-sediment contact is vertical with units truncating at fault surface that are now juxtaposed to salt. - No Salt Exposed
Oligocene 1	- Largely conformable bottom contact - Tabular bedded massive limestone - Benthic foraminifera and corals	- Shallow water, inner ramp carbonate deposition developing reef facies - General uplift of the basin before deposition - Regressive trend from Senonian-Oligocene (deepwater to shallow)
Oligocene 2	- Minor unconformity at the bottom contact major unconformity at the top - Deepwater pelagic marlstones and planktonic foraminifera packstones - only exposed in hanging wall of Trinitari Fault	- Low energy hemipelagic basin environment - Minor uplift and rotation of Bolon towards the north - Trinitari fault still active - Transgressive trend through the upper Oligocene
Oligocene 3	- Matrix supported well rounded conglomerates with a high proportion of diapir derived detritus and reworked Eocene clasts	- Shallow water fan delta sourced from nearby growing relief - no evidence for synsedimentary fault activity - First strong evidence of significant salt exposure in the BSM - Surface present across entire basin representing the uplifting and isolation of the secondary minibasin - Greater salt inflation in the SW as compared to the NE - Development of significant paleotopography
late Paleogene-early Miocene unconformity	- Erosional unconformity with moderate to high angle truncations - Cuts down section towards the SW through the Oligocene and late Eocene units - More erosion in the southwest	
M1	- Onlapping late Paleogene-early Miocene unconformity - Onlaps from SW to NE and into the map plane - Tabular bedded sandy planktonic foraminifera packstones with periodic interbedded marlstones - Halokinetic sequences in western salt-sediment contact - Fanning dips, minor angular unconformities, and wedge-shaped deposition into the map plane (center of SMB) - Reworked upper Eocene clasts in MTDs in SW, some reworked M1 clasts - Growth strata geometries in hanging wall of Reis Mags fault, faulting cuts younger units towards the NE - Diapir derived detritus only in MTD horizons near the SW salt-sediment contact	- Onlapping turbidites deposited in outer shelf to upper bathyal environments - Transgressive trend to deeper depositional environments - Halokinetic sequences in SW indicate salt supply > sed. accumulation rate - MTD's record the unroofing of nearby carapace fragments in SW - HK 5 is a decapitated halokinetic fold due to allochthonous salt breakout over the BSM along a low angle ramp - Extensional faulting migrates further to NE - Begin horizontal axis rotation of the BSM - Salt exposed periodically throughout deposition (related to HKS development) - Isolated minibasin in an evolving allochthonous salt sheet - SW thickening MTS
M1-M2 unconformity	- Erosion of M1 unit in the east - onlapping M2 on the unconformity surface	- Erosive episode related to salt inflation that was greater in the east. - Expose M1 unit eroding most of the upper part of the section in east - ~33° rotation towards the NW
M2	- Onlaps M1 unconformity - Tabular beds rich in diapir derived detritus - Fining upwards sequences, climbing ripples, cross bedding - Reworked M2 conglomerates in eastern part of M2 unit	- Deepwater, proximal turbidites deposited in outer shelf to upper bathyal environments shed from encroaching salt relief - Infilling of minibasin surrounded by allochthonous salt relief - Syn-depositional erosion of M1 relief in the east (salt inflation in Trinitari Fault footwall) - Encroaching salt relief confining basin
M3	- Onlaps the M1-M2 unconformity at a low angle - Debrites rich in diapir derived detritus - Slumps and debrites with angular M1 boulder sized clasts in the NE - Ramp and flat type geometries along the northern contact with the Triassic salt	- Debrites shed into a confined basin in a deepwater outer shelf to upper bathyal environment - Suture at the thickest point of M3 deposition? - Encapsulation of the BSM by allochthonous salt - infilling of small paleo-valley and migration of fault activity to the East Bolon Fault and its synthetic and antithetic counterparts^a

^a HKS: halokinetic sequence, MTD: mass transport deposits, SMB: secondary minibasin, MTS: minibasin tectonostratigraphic sequence, BSM: Bolon Secondary Minibasin.

the Cretaceous (Senonian) and finishing with its present-day re-emergence from the encasing Triassic salt. The key evidence supporting the interpretation of each stage is listed in Table 1.

4.1. Senonian stage (83–66 Ma)

The Senonian strata were deposited in a deepwater basinal setting with hemipelagic sediments covering a previously developed salt structure (Fig. 17a). The condensed nature of the Senonian deposits compared to the regional thickness allows us to classify them as carapace strata based on the definition of Hart et al. (2004). The homogeneous facies, tabular beds, and lack of diapir derived detritus indicates that the flow of salt into the diapir or sheet must have been diminished during Senonian times. Therefore, during Senonian times sedimentation rates were slightly greater than the rate of diapiric rise resulting in a condensed stratigraphic interval to be deposited above the exposed salt. In map view, the thinning of the Senonian unit towards the NE could be related to both previously developed salt top topography and/or early salt withdrawal with migration of the salt from SW to NE. Thickness changes across the extensional faults denote that some of these faults were active, the main one being the Derramador Fault.

4.2. Paleocene – early Eocene stage (66–48 Ma)

The Paleocene to early Eocene unit was deposited in a hemipelagic environment with turbidity currents concordantly above the Senonian strata, except very locally at the footwall of the Trinitari Fault and NE of the East Bolon Fault where they rest directly above the Triassic salt (Fig. 17b). Deposition is characterized by tabular beds that are cut by the Derramador Fault, the Cappuccino Fault, and the West Bolon Fault. Thickness changes and growth stratal geometries denote that these faults were synsedimentary, with the Derramador Fault being the most active during this stage. Apart from these thickness changes, as in the Senonian, an overall thinning is observed towards the NE. This indicates that the Triassic salt continued migrating towards the NE during this time. Despite the continued migration of salt towards the NE, the lack of diapir derived detritus throughout the middle and upper part of the unit indicates that salt was buried by the hemipelagic and turbiditic sediments.

4.3. Late Eocene stage (48–34 Ma)

During this stage subsidence of the BSM continues with the development of the Trinitari listric growth fault in the center accompanied by relative salt inflation along the flanks (Fig. 17b). During the late Eocene water depths decreased and internal platform carbonates were deposited. These carbonates consist of lenticular to tabular limestones deposited concordantly on the underlying Paleocene-lower Eocene unit. Locally, in the NE domains of the structure, upper Eocene rocks are found in direct contact with the Triassic salt. This contact is interpreted as a later, fault related feature and not a record of late Eocene carbonate deposition on exposed salt. Our interpretation is supported by the lack of diapir derived detritus throughout this section. Thickness changes and local growth strata geometries indicates that during the late Eocene stage, extensional deformation migrated to the NE to the Trinitari Fault where growth wedge geometries occur in the hanging wall and a condensed section of late Eocene limestones is found in the footwall of this fault (Fig. 17b). Importantly, the Derramador Fault and the Cappuccino Fault become inactive in the middle-late Eocene stage. The late Eocene Faults are associated with the listric geometry of the Trinitari Fault and accommodate its rotational deformation (Ellis and McClay, 1988; Imber et al., 2003; Brun and Mauduit, 2008). The Trinitari fault was the major active structure inside the secondary minibasin during this stage. Additionally, the NE and SW margins of the minibasin (i.e. the footwalls of East and West Bolon Faults) continued to inflate (Fig. 17b).

4.4. Oligocene stages Po1, Po2, and Po3 (~32–~22 Ma)

The Oligocene stage represents a time of changing depositional environments and the initial transition towards a new overall tectonic environment. This is a period where the influence of extensional growth faults decreases and is only manifested through the salt adjacent extensional structures in the NE domains (Fig. 17c). This stage is divided into substages based on the tectonostratigraphic relationships observed in the field in the Po1, Po2, and Po3 subunits.

4.4.1. Early – middle Oligocene substage Po1

The Po1 substage is essentially a continuation of the late Eocene stage where progressively shallower carbonate platform sediments are deposited (Figs. 15b and 17c). Our interpretation is supported by the presence of corals and silicified rhodoliths indicating a shallowing depositional environment compared to the Senonian-Oligocene units. Tabular bedded, benthic foraminifera grainstones were deposited in an inner platform environment above a very minor unconformity and show a regressive trend towards the upper bounding unconformity.

4.4.2. Late Oligocene stage substage Po2

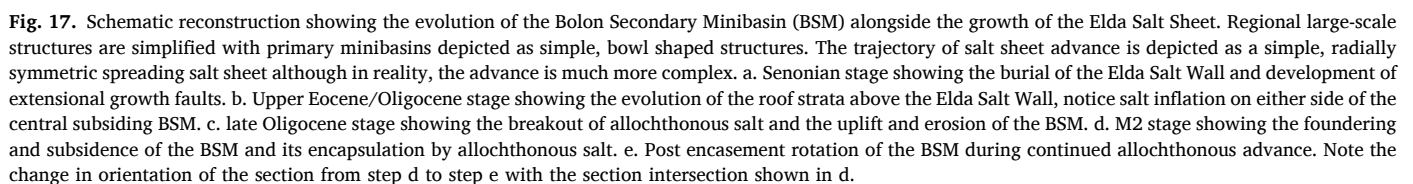
During this substage the water depth rapidly increased with basinal hemipelagic marlstones and then interbedded marlstones and deep-water planktonic foraminifera packstones deposited in an outer ramp setting. These deposits are only exposed in the center of the minibasin above a minor unconformity with the Po1 subunit (Figs. 15b and 17c). The generally low angle of onlap with the underlying Po1 subunit indicates fairly conformable deposition with only minor rotation of the minibasin towards the north. These onlaps are the first record of horizontal axis rotations in the BSM.

4.4.3. Latest Oligocene – early Aquitanian substage Po3

During the final Oligocene substage, more than 110 m of fan delta conglomerates were deposited in the northeastern domains i.e. east of the Trinitari Fault (Figs. 15b and 17c). These conglomerates onlap, at a low angle, the unconformity developed above the late Eocene unit but do not crop out in the hanging wall of the Trinitari Fault, although they are likely present in the subsurface. Therefore, it's reasonable to assume that the Po3 subunit seals the Trinitari Fault which became inactive at this time. The fan delta conglomerates rich in diapir derived detritus suggest a relatively high relief developed adjacent to the northeastern boundary of the minibasin comprised predominantly of Eocene rocks and Triassic salt. From a broader perspective, the Po3 subunit is the first true indication of salt exposure in the BSM as evidenced by the high proportion of diapir derived detritus in the conglomerates.

4.5. Early Aquitanian – late Burdigalian stage (~22–~20 Ma)

The major late Paleogene-early Miocene unconformity that developed in the BSM correlates with the regional unconformity proposed by Geel (1995), Guerrero et al. (2014), Martín-Martín et al. (2018b), and Martín-Martín et al. (2020a). Therefore, this unconformity in the BSM records both a regional tectonic event combined with localized allochthonous salt movement (Fig. 17c). In the BSM this unconformity has a complex 3-dimensional geometry (Fig. 5). In the east, The M1 unit onlaps the unconformity with a younging direction towards the northeast, whereas in the center of the minibasin, near Location 11 in Fig. 5, this unit onlaps into the map plane. These different onlapping directions converge SW of the summit of El Bolon Mountain (adjacent to Location 14 in Fig. 5). This suggests the development of at least 100 m of paleotopography at the top of the Paleogene which is confirmed in the stratigraphic correlation panel in Fig. 10. Contrastingly, in the footwall of the Trinitari Fault, a few hundred meters to the east, the onlaps and truncations are of much lower angle suggesting a relatively muted topography. Importantly, in the sub-unconformity strata, truncations indicate that the SW domains of the minibasin were subject to more



erosion and thus greater salt inflation. Therefore, the paleotopography was likely developed synchronous to or slightly after the deposition of the Po3 unit which contributed to the confinement of the Po3 unit to the eastern most portions of the minibasin (Figs. 15b and 17c).

4.6. M1 Stage—latest Burdigalian – early Langhian (20–15 Ma)

This stage is characterized by a general marine deepening in the BSM where sandy limestones, marls, and fine grained calcarenites were deposited in a deep outer ramp/shelf-bathyal environment. In map view in Fig. 5, these deposits progressively onlap the Paleogene unconformity from SW to NE, but also onlap into the map plane near location 11 in Fig. 5. This indicates that the M1 unit was infilling and onlapping the paleotopography generated during the Aquitanian-late Burdigalian erosion. The Aquitanian – early Burdigalian erosion was likely driven by both regional events and salt inflation uplifting the BSM.

In general, the style of subsidence and sediment accommodation changed from fault-controlled growth wedges that characterize the Paleogene unit, to overall subsidence of the whole BSM where halokinetic sequences and salt-sediment interactions develop around the periphery of the structure (Fig. 17d). Halokinetic sequences 1–5 (Fig. 14a) evidence rapid and increasing subsidence of this side of the minibasin where the rate of salt rise becomes progressively greater than the rate of sediment accumulation, supported by the hook halokinetic sequences and decreasing bed lengths up section in HK4 and HK 5. Concurrently, a series of extensional faults developed (Reis Mags Faults and antithetic counterpart) generating growth wedges in their hanging walls and condensed sections in their footwalls (See Fig. 15b steps 2 and 3). As shown by the growth strata geometries, these faults become younger towards the NE (Fig. 15a and b).

Based on the bedding dip values and discrete dip changes across minor unconformities in the M1 unit, we can deduce that the BSM began to rotate towards the NW, most prominently in the middle-upper part of the M1 unit (Fig. 16 and inter M1 unconformity in Fig. 5). The stereoplot in Fig. 16 shows the poles to bedding of the Miocene units moving up section (bedding data shown in red on Fig. 5), the first four data points are strongly overturned strata near the peak of El Bolon Mountain and are likely associated with halokinetic folding occurring at the edge of the BSM (Fig. 16b). Up section, unconformities and fanning dip data indicate a rotation of about 61° during the middle-late M1 deposition (see red bedding attitudes in Fig. 5 near the center of the BSM). Interestingly, the rotation of the Bolon Secondary Minibasin occurred almost perpendicular to the orientations of the previously developed extensional growth faults. A detailed analysis of the rotation data is provided in the discussion section.

4.7. M2 Stage—Langhian – early Serravallian (20–15 Ma)

Prior to the sedimentation of the M2 unit, the BSM was again rotated 30° towards the NW allowing the M2 unit to onlap the erosional unconformity capping the M1 unit (Figs. 16b and 17d). The M2 deposits continue to record a deep marine depositional environment (outer ramp/shelf) and denote a sedimentary basin bounded by bathymetric highs with adequate topography to drive the observed small scale turbiditic flows into the confined basin. The eastern portion of BSM, in the Trinitari Fault footwall, salt inflation resulted in the erosion of the M1 unit as evidenced by the reworked M1 monomict conglomerates in the M2 and M3 units (Fig. 15a image 1).

4.8. M3 Stage—Langhian – early Serravallian (20–15 Ma)

The M3 unit is a continuation of the M2 unit style of deposition with an important increase in the angularity and distribution of breccias. Notably, the upper contact with the Triassic salt, which overlies this unit and part of the M2 unit, displays ramp-flat geometries that suggest encasement of the BSM by allochthonous salt occurred from SW to NE

(although this is undoubtedly more complex in 3D). This stage shows a marked increase in the ratio of salt extrusion versus sediment accumulation rates. This event is a continuation of the vigorous growth of the diapir that developed along the SW boundary of the BSM during the previous M1 stage. According to the ramp-flat geometries, the stratal truncations, and the geometries of the debrite wedges, the minibasin was mostly encased by salt from the SW with only a small salt sheet emanating from the NE margin (Fig. 17d). Thus, the suture point evidenced by the stratigraphic pinnacle in the M3 unit is located near the third debrite wedge in the M3 unit (**furthest NE reference of Location 18 in Fig. 5**), which could be an autosuture or allosuture (see Dooley et al. (2012) for a detailed discussion and explanation of sutures in allochthonous salt).

4.9. Post-early Serravallian stage (15–0 Ma)

After being fully encased, the BSM was rotated an additional 50–60° to its present position (Fig. 17e). This final rotation is evidenced by the youngest units consistently dipping to the NW at around 50–60°. This rotation was likely driven by the continued influx of salt into the sheet from the feeder uplifting the diapir side of the BSM (the SW) and rotating the whole minibasin towards the NW. The final stage of salt sheet evolution includes the erosional exposure of the BSM and overall uplift and erosion of the Eastern Betics throughout the middle to late Miocene to the present day.

4.10. From Carapace to Secondary Minibasin—Summary of the Bolon Structure

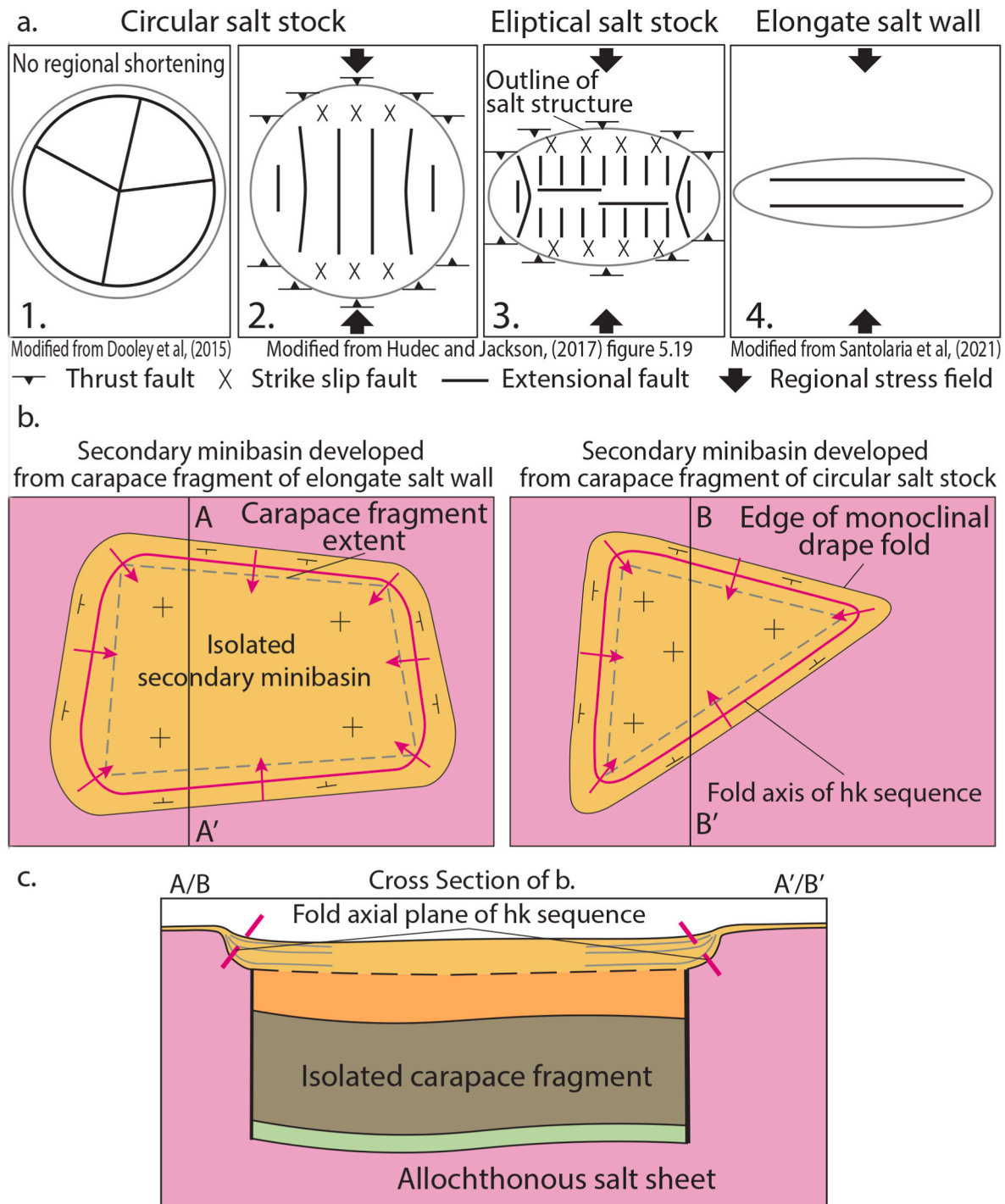
We interpret the Bolon structure as an encased secondary minibasin that evolved from a rafted carapace fragment on an allochthonous salt sheet based on multiple lines of evidence including: (1) parallel bedded, condensed stratigraphic units that were initially deposited on a salt inflated area and subsequently thickened stratigraphic sequences—i.e. carapace and roof development respectively (Senonian-Oligocene strata); (2) erosional truncations formed during roof fragmentation, isolation, and translation (e.g. the late Paleogene-early Miocene unconformity); (3) halokinetic sequences along the salt sediment interface during the early-middle Miocene with attendant unconformities and debrites; and (4) stratigraphic thickening, and rapidly migrating depocenters within the minibasin (see depocenters in Figs. 5, Figure 8, and Fig. 10). The Miocene units of the BSM show northeastward migrating depocenters, a variety of high- and low-angle onlaps, and significant erosional truncations suggesting a complex subsidence history driven by inflating and migrating salt and departing from the standard geometries associated with simple density driven subsidence (Hudec et al., 2009; Jackson and Hudec, 2017). Therefore, we interpret these geometries as evidence that salt inflation related to regional shortening is the primary driver affecting accommodation space and erosion in the BSM (similar to minibasins in Fig. 14 of Hudec et al., 2009). Given this relationship we use the stratigraphic architecture of the BSM to reconstruct part of the salt sheet history delineating periods of salt inflation versus minibasin subsidence.

The evolution of the BSM is represented in a schematic reconstruction shown in Fig. 17. The BSM has previously been interpreted as diapir carapace by Martínez del Olmo et al. (2015) and Naim, (2020). Our detailed field analysis refines this interpretation providing a comprehensive analysis defining the carapace stage (Fig. 17a), the roof development stage (Fig. 17b) and the salt breakout and roof fragmentation stage (Fig. 17c). This was followed by rafting (Fig. 17d) and then foundering and encasement by allochthonous salt and subsequent minibasin rotation (Fig. 17d and e). These regional restoration steps (Fig. 17) correlate with the higher-resolution restorations given in Figs. 14b and 15b.

The carapace stage begins in the Senonian with the development of a condensed stratigraphic sequence above the Elda Diapir (Fig. 17a).

Based on the original definition of carapace by Hart et al. (2004), only the Senonian marlstones should be considered as true carapace strata. The roof development stage encompasses the Pple, Pue, and Po units (Fig. 17b). These units record extensional depocenters thicker than regional stratigraphy and are therefore classified as roof strata (Jackson and Hudec, 2017). Fault growth during this period is related to crestal extension above a rejuvenated diapir caused by the squeezing of diapiric stems, as they are the first areas to accommodate contractional

deformation (Fig. 17b; e.g. Rowan and Vendeville, 2006; Dooley et al., 2015; Santolaria et al., 2021). The salt breakout and roof fragmentation stage begins in late Oligocene times with shortening triggered salt inflation responsible for fully rupturing the roof strata and culminating in the isolation of carapace fragments on the allochthonous Elda Salt Sheet (Fig. 17c; e.g. Dooley et al., 2015). This stage includes: (1) the initial breakout of salt from the Elda Diapir, (2) the isolation of a carapace fragment in the elevated salt sheet (Fig. 17c), (3) the subsequent



Monoclinial drape fold develops at the boundary of the roof fragment.

Fig. 18. a. Schematic map view representations of diapir roofs subjected to regional contractional deformation with the resultant fault orientations shown. b. Map view of minibasins developed from the different shaped carapace fragments post dismemberment. c. Cross section showing the location of halokinetic folds with respect to the carapace bounding faults.

erosion due to salt inflation, and (4) its initial foundering in the early Miocene. Foundering of the carapace fragment is likely associated with the transport of the carapace fragment away from the feeder and over the primary minibasin and therefore away from the upwelling salt (a similar location to what is shown in Fig. 17d). Finally, the encasement and post encasement rotation and grounding of the minibasin places the BSM into its current position. The BSM represents an exquisite field exposure of an encapsulated and transported secondary minibasin developed in a deepwater setting that is directly comparable with other secondary minibasins for example: Fernandez et al. (2021) Fig. 9; Fernandez et al. (2017) Figs. 12 and 15; Alshammasi et al. (2023); and Hudec and Jackson (2022).

4.11. Carapace fragment shape and subsequent secondary minibasin evolution

From the data collected at the BSM, specifically the Paleogene extensional structures, we can hypothesize different endmember scenarios for the fragmentation of carapace strata during regional shortening, which are shown in Fig. 18a and build from the ideas of Dooley et al. (2015), Jackson and Hudec (2017) and Santolaria et al. (2021). These studies show that the shape of the diapir or wall is the main controlling factor that determines the orientation of roof faults and thus the shape of the carapace fragment (Dooley et al., 2015; Jackson and Hudec, 2017; Santolaria et al., 2021). A rising circular salt diapir with no regional shortening results in circumferential extension and triangular shaped roof fragments (Figs. 18a–1). A circular diapir under regional shortening forms horsts and grabens parallel to the regional shortening direction (Figs. 18a–2; Jackson and Hudec, 2017). An elliptical salt stock forms extensional structures both parallel to and perpendicular to the regional shortening direction (Figs. 18a–3). A salt wall forms a graben perpendicular to the regional shortening direction (Figs. 18a–4; Santolaria et al., 2021). Thus, the shape of the diapir controls the shape of the roof fragments that in turn will control the shape of the subsequent secondary minibasin if the fragment is isolated and eventually subsides (Fig. 18b; Dooley et al., 2015; Jackson and Hudec, 2017; Santolaria et al., 2021). In the BSM, the location of the salt-sediment interface during the Miocene was inherited from the shape of the previously developed roof bounding faults (i.e. West Bolon Fault) and the locations of secondary diapirs (Fig. 17b *inflated salt in Eastern Termination*). In other words, the roof bounding faults conditioned the location of later halokinetic folding (Fig. 18b and c). The roof fragment served as an anchor point for the Miocene sediments, with future halokinetic folds developing just outboard of the roof fragment boundary (Fig. 18c). These assumptions and interpretations are most valid in submarine allochthonous salt sheets as opposed to their continental equivalents where erosion and dissolution are prevalent.

5. Discussion

5.1. Bolon 3-D structure: encased secondary minibasin vs. grounded and overlapped roof fragment vs. extruded burial wedge

In the BSM the variation in the stratigraphic architecture and the complex structural relationships make the 3-D interpretation dubious. Specifically, the base salt-minibasin contact is the largest uncertainty in the subsurface interpretation. This contact could feasibly be a tertiary weld, a tertiary thrust-weld, or continuous uninterrupted stratigraphy. To better address these possibilities and their implications we present three cross sections depicting different plausible interpretations of the Bolon structure at depth. Sections are drawn using regional data from previously published geologic maps and cross sections, (e.g. Roca et al., 2006; Guerrero and Martín-Martín, 2014; Martínez del Olmo et al., 2015; Martín-Martín et al., 2018b, 2020a, 2020b; Escosa et al., 2018), reasonable assumptions about salt tectonic systems, comparisons with analogue models (i.e. Dooley et al., 2009, 2012, 2015; Rowan, 2017;

Jackson and Hudec, 2017), comparison with other allochthonous salt systems well imaged in seismic data (Fiduk et al., 2014; Fernandez et al., 2017, 2020a, 2021, 2024), and our field work. Based on the above criteria, we favor the scenario presented in Fig. 19a, an encased secondary minibasin developed from a rafted roof fragment that subsided into an allochthonous salt sheet. In the following paragraphs we present the arguments that support the interpretation in Fig. 19a and refute the others.

The first two scenarios, Fig. 19a and b, are broadly similar in terms of the regional interpretations of the subsalt structure. The Camara Anticline (Fig. 4a and b), to the NW of the Bolon Structure, would represent a primary minibasin developed above autochthonous salt with an associated diapir that was shortened forming a secondary weld/thrust weld or decapitated diapir during the contractional deformation in the middle-late Miocene (Vidal-Royo et al., 2021). In the first scenario, the Miocene units to the NW would correlate to subsalt sediments developed in front of an advancing allochthonous salt sheet, and the BSM-subsalt contact would represent a tertiary weld. In this scenario the initial subsidence of the rafted roof fragment begins once the fragment is no longer located over the feeder. This suggests that the BSM was uplifted more than 2000 m out of the diapir throat, whereas if it were an already developed allochthonous salt sheet this value would be lower (see section 5.2 for a discussion on the salt sheet versus diapir origins for the BSM).

In the second scenario, the subsalt stratigraphy would be continuous with the Miocene stratigraphy of the Camara Anticline, thickening towards the south towards the El Bolon structure. Instead of a secondary minibasin subsiding into allochthonous salt, the structure would represent a piece of the diapiric roof that was rafted to the front of the allochthonous salt sheet before being encircled by allochthonous salt (Fig. 19b inset, note there is no post-grounding sedimentation in this model, see other models from Dooley et al. (2012, 2015)). However, this scenario is unlikely for a few reasons. First, the halokinetic sequences and laterally migrating depocenters are difficult to explain using growth wedge geometries, lateral halokinetic sequence evolution, and progressive unconformities that are necessary in order to obey the geometrical relationships observed in the field. Second, the overturned nature of the El Bolon structure necessitates post-grounding folding of the rafted carapace fragment, but models from Dooley et al. (2015) show that these fragments are more likely to be overridden instead of continually uplifted and folded. In the analogue models, small, thin and flexible rafts (<500 m thick) were folded at the sheet front while the thick and stable rafts (>1000 m thick), as is the case in the Bolon structure, were simply grounded at the sheet front and later encircled by allochthonous salt.

The third scenario presented in Fig. 19c is an extruded burial wedge that began to develop in the Senonian stage. These structures are well developed in the Moab Basin in association with diapir burial (Langford et al., 2022). The base salt structure in this scenario is significantly different, it requires a secondarily welded diapir to be placed directly beneath the El Bolon structure. The burial wedge model requires a significant (>2000 m) and persistent uplift and folding of the thick (>1000 m) Paleogene sediments. However, the development of halokinetic sequences and growth wedge geometries in the NE-SW direction, perpendicular to the folding direction would require a significant subsidence into a large >2.5 km radius diapir with secondary diapirs developing at the periphery. Additionally, the timing of contractional deformation would have to be post Miocene deposition in order for the halokinetic sequences to form prior to folding of the burial wedge. This timing contradicts the established timing of contractional deformation in the Eastern Betics which began before the M1 deposition (Guerrero and Martín-Martín, 2014; Martín-Martín et al., 2018b; Escosa et al., 2018; Jabaloy Sánchez et al., 2019; van Hinsbergen et al., 2020; Pedrera et al., 2020). Last, the isolated blocks of Senonian-lower Eocene strata on the SW minibasin edge cannot form in this scenario. Therefore, we can reject scenario three. Based on the above arguments the most likely

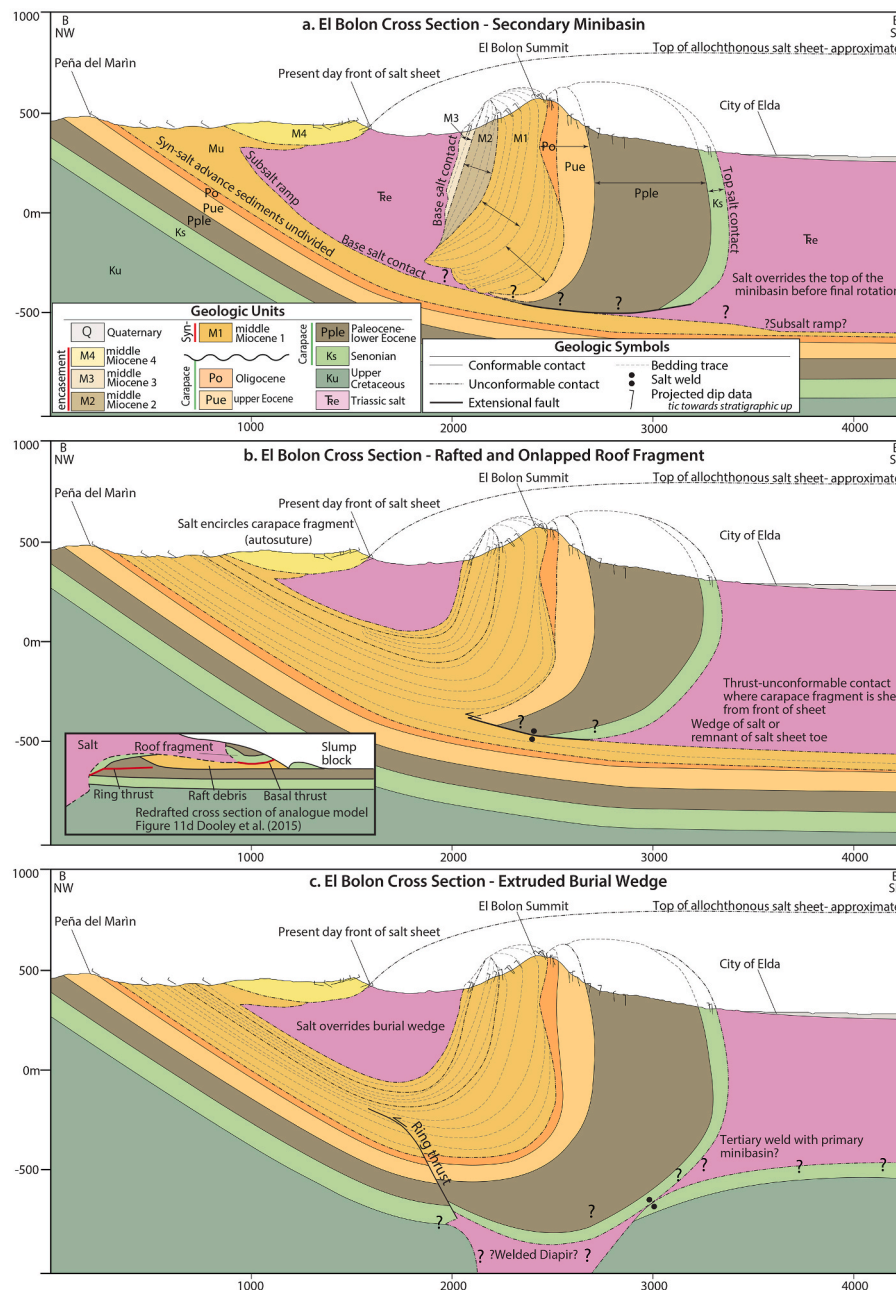


Fig. 19. Three plausible cross sections of the Bolon Secondary Minibasin. a. Isolated and tertiary welded secondary minibasin transported above the subsalt minibasin. b. a rafted carapace fragment that was transported to the front of the allochthonous salt sheet and partially buried by syn-salt advance sediments. c. an extruded burial wedge developed above a secondarily welded diapir. See text for a discussion on the three interpretations and which one fits the field data best.

scenario is one of a rafted roof fragment that formed a secondary minibasin in an allochthonous salt sheet, as shown in Fig. 19a.

5.2. The Bolon Secondary Minibasin Origin Story—Dismembered carapace of an allochthonous salt sheet or dismembered diapir carapace

We interpret the Senonian unit of the Bolon Secondary Minibasin as a carapace deposited above a small buried allochthonous salt sheet or above a passive salt diapir (Fig. 20a and b respectively). The development of carapace strata often indicates a decline in diapiric activity, or atleast the relative rate of sediment accumulation was greater than salt rise (Hart et al., 2004; Rowan and Inman, 2011; Graham et al., 2012; Fiduk et al., 2014; Rowan, 2017).

We constrain the evolution of the carapace and roof strata in the BSM by analyzing the orientation and distribution of faults and comparing

them to known orientations from analog models and other salt basins. The roof of a salt stock would produce wedge-shaped rafted fragments of carapace strata due to radial faulting of a circular roof (Fig. 18a; i.e., (Carruthers et al., 2013; Dooley et al., 2015; Coleman et al., 2018; Gutiérrez et al., 2023)). In contrast, the Senonian unit was broken by numerous sub-parallel extensional growth faults that sometimes affect the Paleocene-Oligocene units. Therefore, we suggest that these roof faults developed above an advancing allochthonous salt sheet (Fig. 20a) or elongated salt wall (Fig. 20b), producing the observed subparallel orientations. On the other hand, carapace developed above salt sheets often exhibit extensional growth faults with listric geometries, reflecting ongoing allochthonous advance with strikingly similar geometries to those observed in the BSM (Hudec and Jackson, 2009). In the case of El Bolon, due to the nature of an isolated minibasin, we cannot directly constrain the original orientation of these faults with respect to the

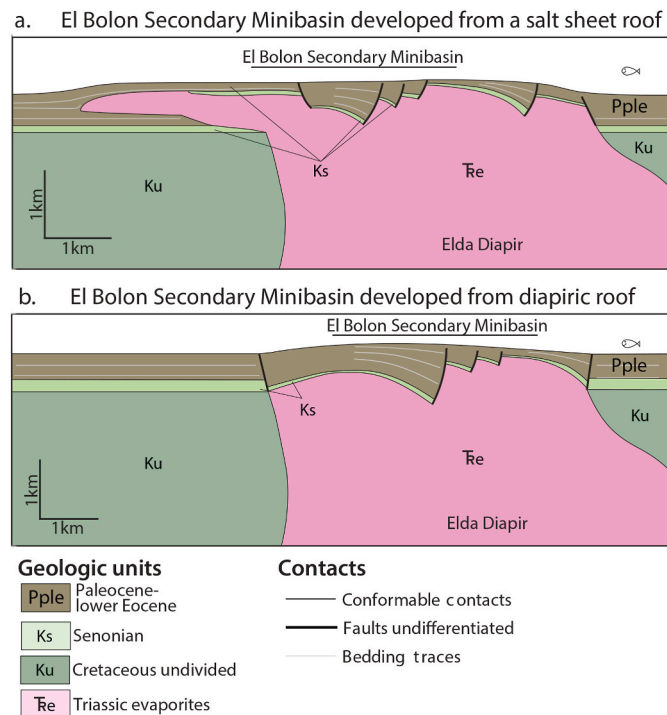


Fig. 20. Two end member scenarios showing the initial stage of the Bolon Secondary Minibasin. a. Shows the evolution of a salt sheet roof with extensional growth faults developed due to gravity spreading. b. diapiric roof with extensional growth faults developed due to crestal extension. Notice the suprasalt Pple unit is thinner than the regional in a but thicker than regional in b.

previously developed salt structure or with the regional shortening direction due to unconstrained vertical axis rotations. Despite this limitation, the sandy character of the turbidites of the Pple unit and its thicker section compared to regional thicknesses suggests subsidence into a salt structure in order to provide adequate accommodation space. Likewise, sandy turbidites are unlikely to be deposited above inflated salt because the turbidity currents would be diverted around the growing bathymetry (see bathymetry shown in Fig. 20a; Fernandez et al., 2024). Additionally, the extensional faults within the BSM indicate formation within a fairly uniform extensional stress field during a period of regional contraction. This is due to crestal extension of the diapiric roof caused by contractional deformation and diapir squeezing, as opposed to dissolution, which would produce more chaotic and disorganized faults (Gutiérrez et al., 2023).

For the case of a small allochthonous salt overhang or sheet that developed in the Late Cretaceous and was buried by the hemipelagic marlstones of the Senonian, the subsequent development of the early salt sheet would have followed a similar sequence of events. We favor the hypothesis of diapiric carapace because it is nearly impossible to account for the thicker than regional character of the Pue and Pple units if they were deposited above allochthonous salt because strata deposited above allochthonous salt should record a condensed stratigraphic sequence when compared to the regional (Hudec et al., 2009; Rowan, 2017).

5.3. Minibasin horizontal axis rotations

Secondary minibasins commonly show wedge-shaped stratigraphic packages associated with the progressive rotation of the basins through time but they never rotate more than 90° due to geometrical constraints imposed by the basal detachment (Jackson and Hudec, 2017; Jackson et al., 2020). For the case of BSM some of the rotation appears to be

accommodated before the final encasement (Fig. 16a and b). By calculating the angle between bedding attitudes in the basal M1 unit and the top of the M3 unit, an angle of 90–110°, it becomes apparent that wedging stratigraphy can't explain the drastic change in bedding attitudes and therefore a different mechanism must be invoked to account for a portion of the folding observed in the BSM. The most likely scenario is one in which the M1 sediments at the peak of El Bolon were onlapping the paleotopography of the late Eocene unit while in contact with salt in a region that has now been removed by erosion (Fig. 19a – cusp near the summit of El Bolon). Therefore, the strongly overturned nature of the M1 unit near the summit of El Bolon is likely the result of halokinetic folding, similar to what is observed in the SW domains of the BSM (Fig. 14a HK 1-5).

The generally planar contact and the consistent dip values along the northern salt-sediment interface suggest that the minibasin was fully encased in salt prior to the final 60° NW directed horizontal axis rotation. Therefore, we hypothesize that the further rotation of the secondary minibasin was driven by additional allochthonous salt flow and/or rolling along a tertiary weld. A similar situation of post-encapsulation rotation is seen in many secondary minibasins in the northern GOM mostly driven by continued allochthonous salt flow (Neely et al., 2017). Admittedly, further rotation could have been driven by subsalt thrusting associated with the later contractional evolution of the Betic fold and thrust belt. Without subsurface data to support this interpretation we believe that the salt driven rotation after encasement and before grounding is the most plausible explanation (Fig. 17e).

5.4. Modes of minibasin encasement and implications for the Bolon Secondary Minibasin

Three styles of minibasin or roof encasement by allochthonous salt have been postulated: (1) foundering or encasement by density driven subsidence into the allochthonous salt, (2) encasement by override in place, where the minibasin is encased by allochthonous salt after welding with the base of salt, and (3) encasement by overriding a salt suture (Rowan and Inman, 2011; Edwards et al., 2014; Rowan, 2017; Jackson and Hudec, 2017; Fernandez et al., 2017). For the case of the Bolon Secondary Minibasin, we show that it was encased by a second level of allochthonous salt. An overriding salt suture is unlikely due to the stratal cutoffs of minibasin flank strata in both the NE and SW domains that indicate the isolated nature of the minibasin. Similarly, the stratal cutoffs indicate progressively younger truncations towards the NE culminating in a pinnacle comprising the youngest strata deposited before complete encasement. We suggest that foundering into an allochthonous salt sheet is the most likely scenario because: (1) the final, 60° northwestwards rotation of the minibasin is mechanically easiest to explain by sinking and concomitant rotation into an extensive and thick, northward advancing salt sheet. Therefore, we can demonstrate that minibasins encased by a second level of allochthonous salt should display: (1) ramp-flat geometries and debrite wedges along the upper contact of the minibasin (fold ramps are also possible), (2) diapir-derived detritus within the uppermost stratigraphy of the minibasin, (3) migrating depocenters not correlated with the overall synclinal minibasin fill, due to accommodation space created by the encroaching salt relief (e.g., M3 unit in Fig. 13b), and (4) A pinnacle in the uppermost minibasin stratigraphy, accompanied by a suture zone within the salt (e.g. Dooley et al., 2012; Edwards et al., 2014).

In contrast, these features will be absent when minibasins or roof fragments are rotated into contact with salt. The salt-sediment contact will be relatively planar due to the lack of syn-encapsulation sediments possibly accompanied by a significant erosional unconformity along the top contact. Care should be taken with sediment-starved minibasins, as the encasement sediments might be relatively thin and dominated by reworked evaporitic material or completely absent.

The encasement of secondary minibasins also depends on the salt budget, proximity to a diapir or another secondary basin, and the

relative sizes of the minibasin compared to the salt structure into which it sinks (Hudec and Jackson, 2006, 2009; Edwards et al., 2014; Dooley et al., 2015; Rowan, 2017; Alshammasi et al., 2023). In the case of the BSM, the autochthonous salt budget was relatively small, with ~700 m of autochthonous evaporites (Ortí et al., 2017; Gómez et al., 2019), whereas examples from the Gulf of Mexico (GOM), the Flinders Ranges, and the Sivas Basin show salt sheets that are often >1 km thick and sometimes reach more than 5 km in thickness (Hudec and Jackson, 2006; Hearon IV et al., 2015; Kergaravat et al., 2016; Rowan, 2017; Fernandez et al., 2020b; Duffy et al., 2020). Because the Bolon Secondary Minibasin was encased by salt before its final rotation, the allochthonous sheet into which it sank must have been greater than the total thickness of the secondary minibasin. Therefore, the Elda Salt Sheet must have been at least 1.4 km thick in order to completely encapsulate the BSM, which has a maximum thickness of 1.3 km. This large salt sheet was likely developed due to the regional contractional evolution of the Betic FTB during the middle Miocene, squeezing primary diapirs and extruding significant amounts of salt (Cater, 1987; Rowan and Vendeville, 2006; Rowan, 2017).

5.5. Comparison with other natural examples

Salt sheets, and rotated and encased secondary minibasins have been identified in seismic lines in the Gulf of Mexico (Hart et al., 2004; Edwards et al., 2014; Alshammasi et al., 2023) the Caspian Basin (Fernandez et al., 2017; Duffy et al., 2017) and in field studies from the Western Betics FTB (Flinch and Soto, 2022), the Adelaide FTB in South Australia (Hearon IV et al., 2015), and the sub-alpine FTB in France (Célini et al., 2022). Here we highlight the key differences and similarities between the BSM and some examples from these other salt basins.

Salt sheets in the GOM tend to be much larger and thicker than allochthonous salt sheets in orogenic settings (Hudec and Jackson, 2009). Despite these differences, the BSM and minibasins in the Gulf of Mexico show many similarities. The basal stratigraphy often comprises tabular bedded carapace strata (like minibasins in Hart et al., 2004; Fiduk et al., 2014; Fernandez et al., 2021) with later deposition of halokinetic units. The sedimentation of the syn-halokinetic units is predominantly controlled by the overall dynamics of the evolving sheet rather than density-driven subsidence (Dooley et al., 2012). In the GOM secondary minibasins are rotated and deformed after encapsulation (Neely et al., 2017). For the case of the BSM, there are no nearby (within 2 km) minibasins that would have obstructed or competed with the BSM (e.g. Fernandez et al., 2021). Instead, the BSM was likely rotated post encapsulation and obstructed by subsalt relief, which was a relic from the advancing salt sheet (i.e., abutting against base salt ramps).

Allochthonous salt and secondary minibasins are well exposed in South Australia (i.e. Hearon IV et al., 2015; Rowan et al., 2020). The main differences between allochthonous salt in the Eastern Betics and South Australia are the presence of diapir-derived detritus beneath allochthonous salt sheets—the so-called “rubble” zones commonly observed in the GOM (Rowan et al., 2020). In South Australia, the lack of subsalt rubble zones is attributed to thin diapiric roofs, smaller and lower-relief sheets/canopies developed in a shelf setting, and a lack of significant suprasalt gravity spreading or gliding (Rowan et al., 2020). Additionally, subsalt deformation appears to be mostly confined to ramps associated with allochthonous salt advance (Williams et al., 2019). In contrast, the syn-encapsulation strata at the BSM are predominantly comprised of reworked Triassic debris. The more prevalent diapir derived detritus in the Elda Salt Sheet could be associated with the layered evaporite sequence of the Triassic evaporites, producing high-relief allochthonous salt structures susceptible to slope failure. Like observations made in South Australia, the strata directly below (within 1–3 m) the allochthonous salt at the BSM are effectively undisturbed and show no evidence of shear deformation, faulting, fluid trapping (no matrix cements or veins), or significant fracturing. Therefore, we agree

with previous interpretations of subsalt rubble zones as being sourced from the erosion and slumping of salt sheet roofs in front of the advancing sheet (e.g. Hudec and Jackson, 2006; Hearon IV et al., 2015; Rowan et al., 2020).

In the Western Betics, secondary minibasins and carapace fragments have been found in abundance, resulting in a complex amalgamation of minibasins and rafts separated by large outcrops of Triassic salt. This has been interpreted as a previously developed allochthonous salt nappe incorporated into the evolving Betic FTB (Flinch and Soto, 2017, 2022). Overall, the structures in the Eastern Betics show much less shortening-related deformation, for example diapirs are secondarily welded and only minorly thrust. The pre-shortening location of the Elda Salt Sheet represents a more proximal domain of the passive margin system compared to the Western Betic domains and correlates to a diapiric/extensional domain close to the upper slope-mid slope transition of a typical passive margin salt system (Jackson and Hudec, 2017). In the Elda region, allochthonous salt evolved by breaching previously developed diapiric roofs before advancing laterally as individual extrusions from passive diapirs. The difference in structural style is related to the development and distribution of passive diapirs and the relative location of structures along the passive margin, where the Elda region represents the intermediate domain and the salt nappes of the Western Betics are related to the distal toe of the Mesozoic passive margin salt system (Flinch and Soto, 2017, 2022).

5.6. Implications for the structural evolution of the Eastern Prebetics

Our research on the BSM largely aligns with recent publications on the structural evolution of the Eastern External Betics. However, we suggest that the presence of allochthonous salt has been greatly underestimated throughout the region (i.e. De Ruig, 1995; Geel, 1995; Martín-Martín et al., 2018a, 2020b; Escosa et al., 2018). Although small salt sheets have been recognized in the Eastern Prebetics by Tent-Manclús et al. (2000, 2001), the extent and size of these structures has been largely overlooked in most historic and recent publications. Our data from BSM indicates that a large (at least 1.4 km thick) allochthonous salt sheet developed in the Elda region. Therefore, studying the isolated outcrops of non-evaporite lithologies within the large Triassic salt exposures, and conducting detailed studies focused on the structural and sedimentological architecture will provide a more complete history of the diapiric evolution of the area and the degree of allochthonous salt development, and provide a more complete history of the south Iberian passive margin evolution.

6. Conclusion

The Bolon Secondary Minibasin (BSM) is a pristine outcropping example of a previously encased secondary minibasin that developed in a deepwater depositional environment. Field data show that the BSM contains two distinct structural and stratigraphic domains: (1) an 800 m thick carapace/roof of terrigenous and outer platform deposits developed over a salt structure, and (2) a Miocene unit that developed in conjunction with the evolving allochthonous salt sheet. The roof strata consist of a Senonian carapace and Paleogene roof strata characterized by extensional growth faults. Subsequently, a minibasin-wide unconformity followed by fan deltas dominated by diapir-derived detritus record the uplift, dismemberment, and eventual isolation of the roof fragment within the allochthonous salt.

Throughout the middle Miocene, the continued extrusion of allochthonous salt from the feeder uplifted and transported the roof fragment. Once the fragment was no longer located over upwelling salt it began to subside. Subsidence into the salt sheet was accompanied by halokinetic folding and slumping/reworking of carapace strata around the flanks of the BSM. Continued subsidence allowed salt to fully encase the BSM before its final 60° rotation towards the NW. The horizontal axis rotation of the BSM is attributed to wedging stratigraphy within the

minibasin associated with high rates of salt extrusion on the SW flank. This was likely a factor of continued salt flow from the feeder into the sheet during regional shortening. The Elda Salt Sheet and BSM are a pristine example of secondary minibasins evolving in a sub-aqueous contractional setting providing a much-needed field-based analogue for seismic interpretation.

Key conclusions about secondary minibasins developed from rafted roof fragments:

1. Rafted roof fragments can provide an effective seed for minibasin inception providing a significant density contrast with the surrounding salt once the fragments become isolated away from the feeder.
2. The carapace bounding faults conditioned the shape of the BSM and the subsequent location of halokinetic folding, which nucleated above the roof bounding faults.
3. The BSM is characterized by syn-encasement sediments at its highest levels with high proportions of diapir derived detritus sourced from the encasing salt. Therefore, we propose that ramps, flats, fold-ramps, and debrite wedges should be expected at the top of other encased secondary minibasins.
4. Horizontal axis rotations in the BSM are attributed to both syn-depositional rotation and post-encapsulation salt flow.
5. The evolution of the BSM was driven first by carapace and roof development above a long-lived salt structure followed by dismemberment of the roof during regional shortening related salt breakout. Eventually the roof fragment subsided into the allochthonous salt sheet and was fully encased. Long after the encasement and rotation of the BSM, the Elda Salt Sheet was eroded and degraded exposing the subsalt BSM in its current position.

CRediT authorship contribution statement

David Paul Canova: Writing – review & editing, Writing – original draft, Project administration, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Eduard Roca:** Writing – review & editing, Writing – original draft, Validation, Supervision, Resources, Methodology, Investigation, Funding acquisition, Conceptualization. **Oriol Ferrer:** Writing – review & editing, Writing – original draft, Supervision, Resources, Project administration, Methodology, Investigation, Funding acquisition, Conceptualization. **Carles Ferrández-Cañadell:** Writing – review & editing, Formal analysis, Data curation. **Frederic O. Escosa:** Writing – review & editing, Validation, Investigation, Data curation. **Jawad Afzal:** Writing – review & editing, Writing – original draft, Formal analysis, Data curation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.marpetgeo.2025.107418>.

Data availability

Data will be made available on request.

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