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Minorca, an exotic Balearic island (western Mediterranean)

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ABSTRACT

Despite forming part of the Balearic group of islands, Minorca differs stratigraphically and structurally from Majorca and Ibiza: i) Paleozoic rocks are abundant in Minorca but are very scarce in Majorca and are absent in Ibiza. Eocene-Oligocene sediments are virtually absent in Minorca but crop out extensively in Majorca, ii) Contractional structures in Minorca differ in direction (aligned SW-NE in Majorca and Ibiza and N-S in Minorca) and in age from those in Majorca and Ibiza. In addition, Paleozoic deposits of Minorca do not correlate with those of Sardinia, where in addition the Triassic sediments are not very abundant. Contractional deformation in Sardinia is in part older (late Eocene-early Miocene) than in Minorca (early Miocene?). Given its Neogene clockwise rotation, Minorca cannot be considered a small block dragged by the early Miocene counter clockwise rotation of the Corsica-Sardinia block. Furthermore, the Paleozoic and Mesozoic stratigraphy of Minorca (siliciclastic late Paleozoic rocks, Triassic Germanic facies and Jurassic carbonates) has affinities with that of the southern part of the Catalan Coastal Ranges. Thus, of all the Balearic islands, Minorca seems to have traveled the farthest during the Valencia Trough rifting with the result that it resembles an exotic island forming part of the Balearic foreland.

KEYWORDS **Minorca. Balearic islands. Western Mediterranean. Stratigraphy. Structural Geology.**

INTRODUCTION

Minorca is the North-easternmost island of the Balearic group between the Algerian Basin to the South and the Valencia trough to the North (Fig. 1). Other islands include Majorca, Ibiza and some smaller islands. Minorca is a WNW-ESE elongated island and shows two geologically contrasting regions, one in the North and the other in the South. The former is mainly made up of late Paleozoic and Mesozoic rocks deformed during Variscan and Alpine orogenies. The southern region is constituted by upper Miocene shallow water limestone layers in a shallow dipping attitude. The contact between the northern and southern regions is rectilinear, strongly suggesting a fault beneath the uppermost Miocene limestones (Gelabert *et al.*, 2005).

The Balearic islands are the emerged areas of the continental block known as the Balearic promontory. This block occupies a central position in the western Mediterranean, which is considered to be part of a Neogene to Quaternary back-arc complex. Such a tectonic framework is thought to be the result of i) the rollback of an oceanic slab and the progressive SE retreat of a West-

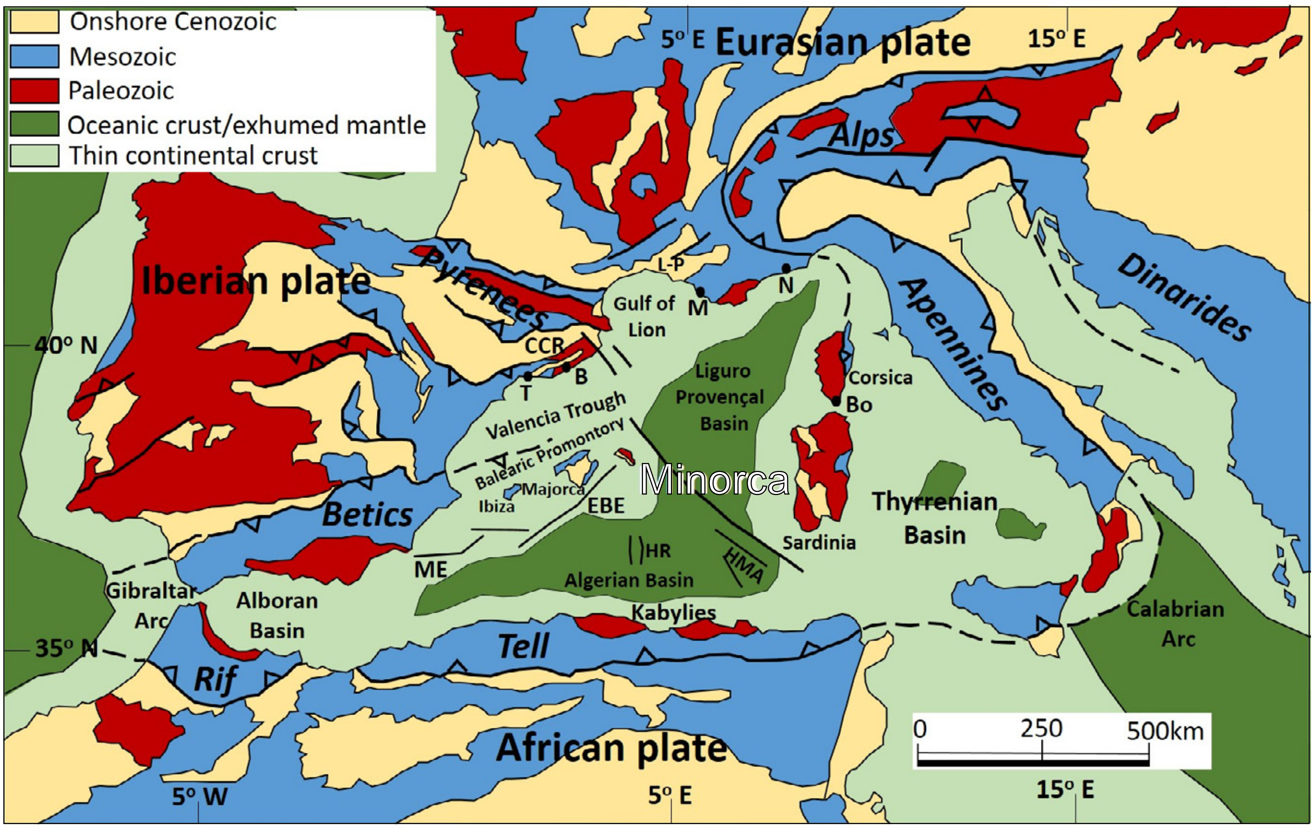

FIGURE 1. Location of the island of Minorca in the western Mediterranean and geology map of the surrounding areas. B: Barcelona; Bo: Bonifacio; CCR: Catalan Coastal Ranges; EBE: Emile Baudot Escarpment; HR: Hannibal Ridge; HMA: Hamilcar Magnetic Anomaly; L-P: Languedoc-Provence region; M: Marseille; ME: Mazarron Escarpment; N: Nice; T: Tarragona.

dipping subduction zone at present located in the Calabrian Arc (Doglioni *et al.*, 1997; Gelabert *et al.*, 2002; Spakman and Wortel, 2004) and ii) the WSW retreat of an East-dipping subduction today located in the Gibraltar Arc (Lonergan and White, 1997). Subduction retreat resulted in upper plate extension (Royden, 1993) that led to stretching and hyper-extension of the magma-poor Gulf of Lion-Valencia continental margin (Jolivet *et al.*, 2015; Granado *et al.*, 2016) and to the formation of the Valencia trough (Sàbat *et al.*, 1995; Roca, 2001), the Liguro-Provençal (Gailler *et al.*, 2009), Algerian (Booth-Rea *et al.*, 2007) and Tyrrhenian basins (Sartori *et al.*, 2004). The deepest parts of these basins are constituted by an anomalous oceanic crust or exhumed continental mantle (Hinz, 1972, 1973; Prada *et al.*, 2014; Jolivet *et al.*, 2015). Between these extended areas, continental blocks such as the Corsica-Sardinia block and the Balearic promontory act as crustal boudins (Doglioni *et al.*, 1999) which underwent less extension. These boudins were left behind during arc migration.

Rifting in the Gulf of Lion and the Valencia margin started at 30Ma (Jolivet and Faccena, 2000). Counter-clockwise rotation of the Corsica-Sardinia block of *ca.* 45º for Corsica and 55º for Sardinia and the formation of the deepest parts of the Liguro-Provençal Basin due to drifting occurred between 21 and 16Ma in the early Miocene (De Jong *et al.*, 1973; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007). Neogene clockwise rotation of *ca.* 20º is recorded in the Balearic islands (including Minorca) and could be in part attributed to stretching in the Valencia trough (Parés *et al.*, 1992).

Despite a consensus on the evolution of the western Mediterranean (Rosenbaum *et al.*, 2002; Schettino and Turco, 2006; Carminati *et al.*, 2012), certain aspects have not yet been resolved (*e.g.* the origin of the Algerian Basin). Because Minorca differs from the other Balearic islands and because it occupies a central position in the western Mediterranean, its evolution and its relationships with the surrounding areas are essential to better understand the geodynamics of the western Mediterranean.

The present study seeks to provide a comprehensive review of the geology of Minorca, highlighting its Mesozoic and Cenozoic evolution. It is based on the analysis of the geological map by Bourrouilh (1983) and is complemented by our own field work. The effects of the Variscan orogeny on Minorca fall outside the scope of this paper.

The geology of Minorca is poorly documented. Hollister (1934) considers that Minorca differs from the other Balearic islands. He regards these islands as the NE prolongation of the Betic ranges and correlates Ibiza and Majorca with its external zones and Minorca with its internal zone. Fallot (1948) agrees with this conclusion and considers that Minorca belongs to the Malaguide (*i.e.* an internal zone in the Betics), but in a earlier work he suggests that Minorca is probably an emerged part of the foreland (Fallot, 1923). The role of Minorca in the Alpine System continues to be the subject of much controversy (Bourrouilh, 2016). Bourrouilh (1983) completed a detailed map (1:50,000) of the whole island and described the stratigraphy and the Variscan and Alpine deformations. He placed especial emphasis on a set of WNW-ESE striking faults. His work constitutes a milestone in our knowledge of the geology of Minorca. Rosell *et al.* (1990) mapped the island at 1:25,000 and compiled a tectosedimentary evolution (Rosell and Elizaga, 1989). The sedimentology of the Permo-Triassic siliciclastics (Rosell *et al.*, 1988; Goméz-Gras and Alonso-Zarza, 2003; Linol *et al.*, 2009) and the late Miocene limestones (Obrador *et al.*, 1983a; Pomar *et al.*, 2002) have aroused considerable interest. Some geomorphological studies (Fornós *et al.*, 1998; Gelabert *et al.*, 2005) and paleomagnetic data (Parés *et al.*, 1992) have also been published.

GEOLOGICAL SETTING AROUND MINORCA

This section is a brief introduction to the geology of the areas surrounding Minorca: the islands of Majorca and Ibiza, the Catalan Coastal Ranges (CCR), the Languedoc-Provence region, the islands of Corsica and Sardinia, the Valencia trough, the Gulf of Lion and the Liguro-Provençal and Algerian basins (Fig.1).

Majorca is the biggest of the Balearic islands. It consists of a set of basins and ranges striking NE-SW or NNE-SSW. The ranges are mainly constituted by Triassic and Jurassic rocks. Paleozoic rocks are present only in one very small outcrop that corresponds to Carboniferous Culm facies (Rodríguez-Perea and Ramos-Guerrero, 1984). The Triassic is made up of Germanic facies composed of a lower siliciclastic series followed by carbonates and by evaporitic series whereas the Jurassic consists of shallow water and slope limestones (Alvaro *et al.*, 1989). These Mesozoic rocks are overlain by thin Eocene, Oligocene or Miocene conglomerates, marls, sandstones and limestones (Ramos *et al.*, 1989). Cretaceous and Paleocene rocks are scarce or absent. The stratigraphic gap between Mesozoic and Cenozoic increases North-westwards. The structure of Majorca is characterized by a NW-directed thrust system detached in the Triassic evaporitic facies (Fallot, 1922; Gelabert, 1998). Thrust-related deformation propagated from SE to NW, starting in the Chattian (late Oligocene) and reaching its peak in the middle Miocene (Fig. 2; Gelabert *et al.*, 1992; Sàbat *et al.*, 2011) during the formation of a turbiditic basin. Upper Miocene strata are subhorizontal and lie unconformably over folded and thrusted rocks. Basins are filled mostly by Neogene sediments (Benedicto *et al.*, 1993), they are in part contractional piggyback basins and in part extensional post-orogenic basins (Sàbat *et al.*, 2011).

The island of Ibiza is located at the southwestern end of the Balearic islands (Fig. 1). Paleozoic rocks are absent. The Mesozoic sequence is almost complete except for the Triassic Buntsandstein. It is constituted by marls, dolostones and limestones of neritic and pelagic environments. The Cenozoic sequence is incomplete following a major hiatus. This sequence starts with a basal conglomerate the age of which is disputed: Etheve *et al.* (2016) using dubious data attribute them to late Oligocene-earliest Burdigalian (early Miocene) whereas Rangheard *et al.* (2011) consider that they are probably Burdigalian. Overlying these conglomerates are Burdigalian and Langhian (middle Miocene) turbiditic marls and sandstones (Rangheard *et al.*, 2011; Etheve *et al.*, 2016). These sequences are deformed by several middle Miocene NW-directed thrusts which are unconformably overlain by upper Miocene limestones and silts (Rangheard, 1971). In a more recent paper, Etheve *et al.* (2016) focus on the stages of the deformation affecting Ibiza: i) late Oligocene to early Miocene extension, ii) middle Miocene compression and iii) late Miocene to recent extension (Fig. 2). Ibiza and Majorca have much in common and the differences in the respective evolutions shown in Figure 2 are probably due more to interpretation rather than to data (Sàbat *et al.*, 2011; Etheve *et al.*, 2016).

The CCR (Fig. 1) are constituted by a Variscan basement consisting of i) early Paleozoic mainly low-grade metasediments, ii) late Paleozoic limestones, radiolarites (lidites) and flysch Culm facies deposits (Santanach *et al.*, 2011) and iii) upper Carboniferous-Permian plutonic bodies. To the N and NE of Barcelona (Fig. 1), the cover rocks are Triassic Germanic facies that are unconformably overlain by Paleogene conglomerates, sandstones, marls and limestones. To the SW of Barcelona, the cover sequence is more complete and Jurassic dolostones and Cretaceous shallow water limestones are present (Anadón *et al.*, 1979; Institut Geològic de Catalunya, 2010). Early Paleozoic rocks are absent around Tarragona and the cover sequence is limited to the Germanic Triassic facies and Jurassic dolostones (which occurs in Minorca); nevertheless, there is a very local presence of Cretaceous rocks (Salas, 1987). The main Cenozoic tectonic events recorded in the CCR are: i) transpression during Paleogene N-S compression with the formation of NE-SW folds and thrust faults and

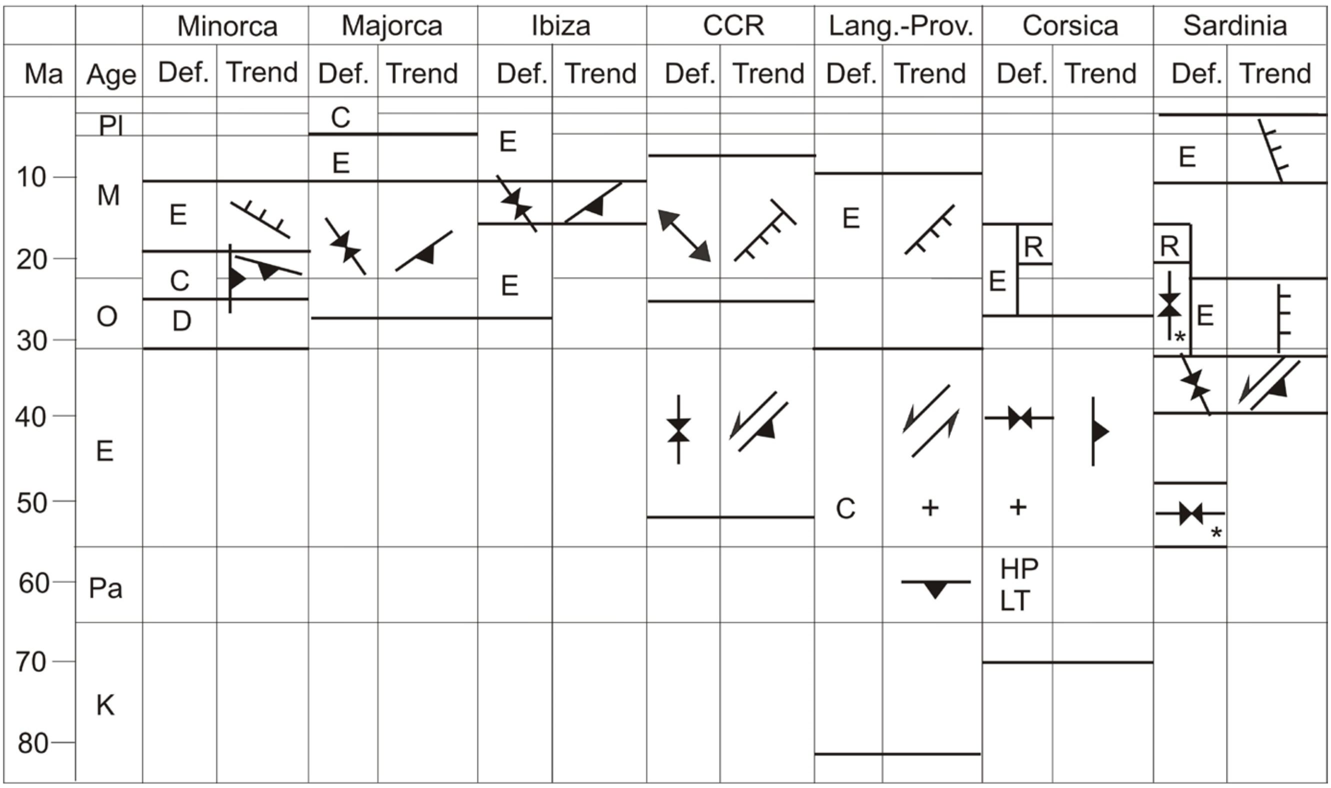

FIGURE 2. Chronological chart of deformation (def.) in the areas surrounding Minorca. Age, Pl: Pliocene; M: Miocene; O: Oligocene; E: Eocene; Pa: Paleocene; K: upper Cretaceous. Deformation, C: Compression; D: Detachment; E: Extension; HP–LT: High-Pressure–Low-Temperature; Lang.-Prov.: Languedoc-Province region; R: Rotation. Age of deformation in Majorca and Ibiza according to Sàbat *et al.* (2011) and Etheve *et al.* (2016), respectively. Deformation in Sardinia marked with * is in line with Carmignani *et al.* (2004).

ii) Neogene rifting due to NW-SE extension caused by the opening of the Valencia trough (Fig. 2; Fontboté, 1954; Anadón *et al.*, 1979; Guimerà, 1984). Rifting generated NE-SW normal faults and associated grabens infilled by Miocene sediments, and on-shore and off-shore alkaline volcanic edifices associated with NW-SE transversal faults (Martí *et al.*, 1992; Maillard and Mauffet, 1993; Roca, 2001; Savelli, 2002; Martí *et al.*, 2011).

The Languedoc–Provence region has several basement outcrops in small massifs between Marseille and Nice (Fig. 1). The basement in this region is constituted by early Paleozoic metasediments, Carboniferous granitoids and Permian volcanic rocks. These volcanic rocks are similar to those encountered in NW Corsica (BRGM, 1980). The sedimentary cover consists of Triassic in Germanic facies followed by Jurassic and early Cretaceous limestones and dolostones that are unconformably overlain by thick late Cretaceous marine limestones (Curnelle and Dubois, 1986). This unconformity is punctuated with lenses of bauxite. Upper Cretaceous limestones are overlain by uppermost Cretaceous, Paleocene and Eocene continental sediments. Oligocene continental and Miocene marine detritic sediments are uncomformable (Tempier, 1987). A large North-verging deformed area resulted from the collision between Corsica-Sardinia (Iberia) and Eurasia (Arthaud and Séguret, 1981). The northern foreland basin in the Provence region underwent widespread inversion of Permian and Mesozoic basins (Roure and Colleta, 1996). E-W-trending folds and thrusts that developed within the Mesozoic cover were detached above the Triassic evaporites. The Paleozoic basement also played a large part in the deformation. At several points along the Provençal coast, it was thrusted on Permian and Mesozoic rocks (Lacombe and Jolivet, 2005). Shortening occurred from the late Senonian to the late Eocene (Fig. 2; Séranne *et al.*, 1995) and the estimated amount is about 30km (Tempier, 1987). Late Variscan NE-trending faults such as the Cevennes, Nimes and Durance faults were reactivated as left-lateral strike slip faults during the Pyrenean orogeny and as extensional faults during the Oligocene (Benedicto *et al.*, 1996). The Languedoc-Provence region can be regarded as the eastern continuation of the Pyrenees.

The island of Corsica (Fig. 1) has been divided into two different units: Variscan and Alpine. Variscan Corsica is located in the West and South of the island and is constituted by a Variscan basement made up of Carboniferous granites and granodiorites and Permian volcanic rocks. The Variscan basement was affected by NE-SW striking faults and is overlain by a thin sedimentary cover consisting of Mesozoic shallow water limestones and Eocene detritic rocks. By contrast, Alpine Corsica is located in the NE of the island and is constituted by Mesozoic oceanic metasedimentary rocks (commonly referred to as schistes lustrées), oceanic crustal rocks such as pillow basalts, gabbros and peridotites and continental crustal rocks composed of gneisses (Rossi and Rouire, 1980). E-W compression in Alpine Corsica coeval with HP–LT metamorphism resulted in overthrusting Variscan Corsica starting in the late Cretaceous (Mattauer *et al*., 1981; Harris, 1985) and continuing from Paleocene to early Oligocene (Bézert and Caby, 1988; Waters, 1990; Malavielle *et al*., 1998). Thrust faults strike N-S with clear top-to-West displacement (Fig. 2). It should be noted that this part of the history precedes the early Miocene counter-clockwise rotation of the Corsica-Sardinia Block. This rotation and the late Oligocene-early Miocene extension in Alpine Corsica (Jolivet *et al*., 1990) are both attributed to the formation of the Liguro-Provençal Basin. Undisturbed late Miocene shallow water limestones are present at several localities such as Bonifacio.

Sardinia is a N-S elongated island (Fig. 1). Its basement is constituted by early Paleozoic metasediments intruded by Variscan granitoids. These intrusions are located in the northern part of the island as a prolongation of those cropping out in Corsica. The main Variscan folds trend WNW. The Mesozoic cover begins as a discontinuous Triassic in Germanic or transitional to alpine facies (Costamagna and Barca, 2002). Jurassic dolostones and limestones overlie the Triassic beds or lie unconformably over the basement; Jurassic is overlain by Cretaceous limestones (Chabrier and Mascle, 1984). Paleogene is scarce and consists of thin conglomeratic beds, coals and limestones (Carmignani *et al*., 2004). According to Cherchi and Tremolieres (1984), late Eocene and early Oligocene NW-SE compression generated open folds and reactivated Variscan NE-SW faults as oblique thrust faults in the northern part of the island; moreover, folds and strike-slip faults resulting from early Eocene E-W shortening and Oligocene-Aquitanian (early Miocene) N-S shortening are recognised in southwestern and North-Central Sardinia (Fig. 2; Carmignani *et al*., 2004). Rifting during the Oligocene (33 to 23Ma) resulted in the N-S Sardinia rift which is infilled by continental sediments. This rifting was associated with the rotation of Corsica-Sardinia Block (*i.e*. 21 to 16Ma) but preceded the rotation and was coeval with calc-alkaline volcanism (*i.e*. andesites). The Sardinia Rift is overlapped by middle and upper Miocene neritic limestones. Finally, upper Miocene and Pliocene alkaline basalts and the NNW-SSE rift of Campidano are attributed to the Tyrrhenian extension (Carmignani *et al*., 1994).

The Valencia trough (Fig. 1) attains a depth of 2,200m; its crust is continental and is only 10–15km-thick along the axis of the trough. The structure along NW-SE cross-sections is asymmetrical: the Iberian slope shows steeply dipping extensional faults whereas the Balearic slope is characterized by highs bounded by both normal and thrust faults. The crustal structure resulted from Mesozoic rifting, the latest Cretaceous-Eocene partial inversion, early to middle Miocene rifting and local development of contractional faults in the middle-late Miocene (Sàbat *et al*., 1995; Roca, 2001; Carminati *et al*., 2012; Ayala *et al*., 2015; Etheve *et al*., 2018).

The Gulf of Lion and the Liguro-Provençal Basin (Fig. 1) comprises i) a broad NW extensional margin (Gulf of Lion) where the crust is 20 to 14km-thick, ii) a central flat distal to oceanic part with a maximum water depth of 2,800m and a minimum crustal thickness of 5km and iii) an abrupt Corsica-Sardinia margin (Pascal *et al*., 1993; Gailler *et al*., 2009; Carminati *et al*., 2012). The Gulf of Lion and the Liguro-Provençal Basin resulted from the continental margin hyper-extension (Jolivet *et al*., 2015). Rifting started in the early Oligocene (34Ma) and evolved to drifting from the late Aquitanian to the late Burdigalian (21–16Ma); drifting is attributed to the counter clockwise rotation of the Corsica-Sardinia Block (Séranne *et al*., 1995; Carminati *et al*., 2012). The margin displays latest Messinian to recent gravity-driven salt tectonics arranged in structural domains: up dip extension, down dip translation and distal compression. These salt tectonic domains are closely related to the underlying crustal arrangement that resulted from the rifting hyper-extension (Granado *et al*., 2016). The thin continental crust, the exhumed mantle and the oceanic domain (if present) are all covered by a thick Neogene to recent sedimentary cover in a sag-type basin.

The Algerian Basin (Fig. 1) has a crust that is 4–6km thick (Hinz, 1972, 1973) and has an oceanic character (Booth-Rea *et al*., 2007). It has three prominent features: i) the Emile Baudot-Mazarron escarpment, ii) the Hannibal Ridge and iii) the Hamilcar magnetic anomaly. The Emile Baudot-Mazarron escarpment is a zig-zag fault system striking NE-SW to E-W that partially corresponds to the boundary between the continental crust of the Balearic promontory and the oceanic crust of the Algerian Basin. The Hannibal ridge is located North of the Great Kabylie striking NNE-SSW and is probably a late Miocene volcanic feature (Mauffret *et al*., 2004). The Hamilcar magnetic anomaly consisting of a set of fan shaped anomalies striking NW-SE is located at the boundary between the Liguro-Provençal and Algerian oceanic basins (Galdeano and Rossignol, 1977). The age of the Algerian Basin is unknown but must be older than the Messinian salt present in the basin and could be as old as the Liguro-Provençal Basin (summary in Carminati *et al*., 2012). The kinematics of the Algerian Basin oceanic opening is a contentious subject. Some works such as those by Vergés and Sàbat

(1999), Schettino and Turco (2006) and Vergés and Fernàndez (2012) suggest that the main mechanism is the early Miocene NNW-SSE extension between 23 and 16Ma (*i.e.* prior to the collision between Kabylie and Africa). Alternatively, Mauffret *et al.* (2004) proposed a 560km of E-W extension between 16 and 8Ma (*i.e.* Langhian to Tortonian). Moreover, Martínez-Martínez and Azañón (1997, 2002) describe both the late Burdigalian–Langhian NNW-SSE and Serravallian WSW-ENE continental extensions in the Betic Range. As a result, models considering two orthogonal extensions for the opening of Algerian Basin have been proposed (Rosenbaum *et al.*, 2002; van Hinsbergen *et al.*, 2014), but the amount, place, age and nature (either continental or oceanic) of these two extensions have not been satisfactorily addressed to date.

STRATIGRAPHY

The Paleozoic of Minorca consists almost entirely of non metamorphic terrigenous clastic deposits and embraces the Devonian to the Carboniferous and the upper Permian.

The Devonian is almost complete and shows flysch facies. According to Bourrouilh (1983), the Devonian rests upon black siltstones and clays that could be Silurian; but later authors (Rosell and Elizaga, 1989) consider that these black deposits may be resedimented facies in the lower Devonian. The Devonian sequence is made up of fine terrigenous clastic sediments, mostly thin-bedded sandstones and shales (Bourrouilh, 1983). Rosell and Elizaga (1989) interpreted this sequence as distal lobes of turbidite fans and basin plain. The total thickness of the Minorca Devonian has been estimated at one thousand meters (Fig. 3).

Carboniferous deposits consist of mostly terrigenous clastic sediments. Two cycles have been described (Bourrouilh, 1983; Rosell and Elizaga, 1989): the lower cycle is composed of terrigenous and calcareous turbidites. The transition to the upper cycle is defined by an olistostrome up to 50m thick, followed by slumped radiolarites and dark to red shales with some olistoliths of volcanic rocks. The upper cycle is made up of facies Culm consisting of thin-bedded turbidites with many intercalated

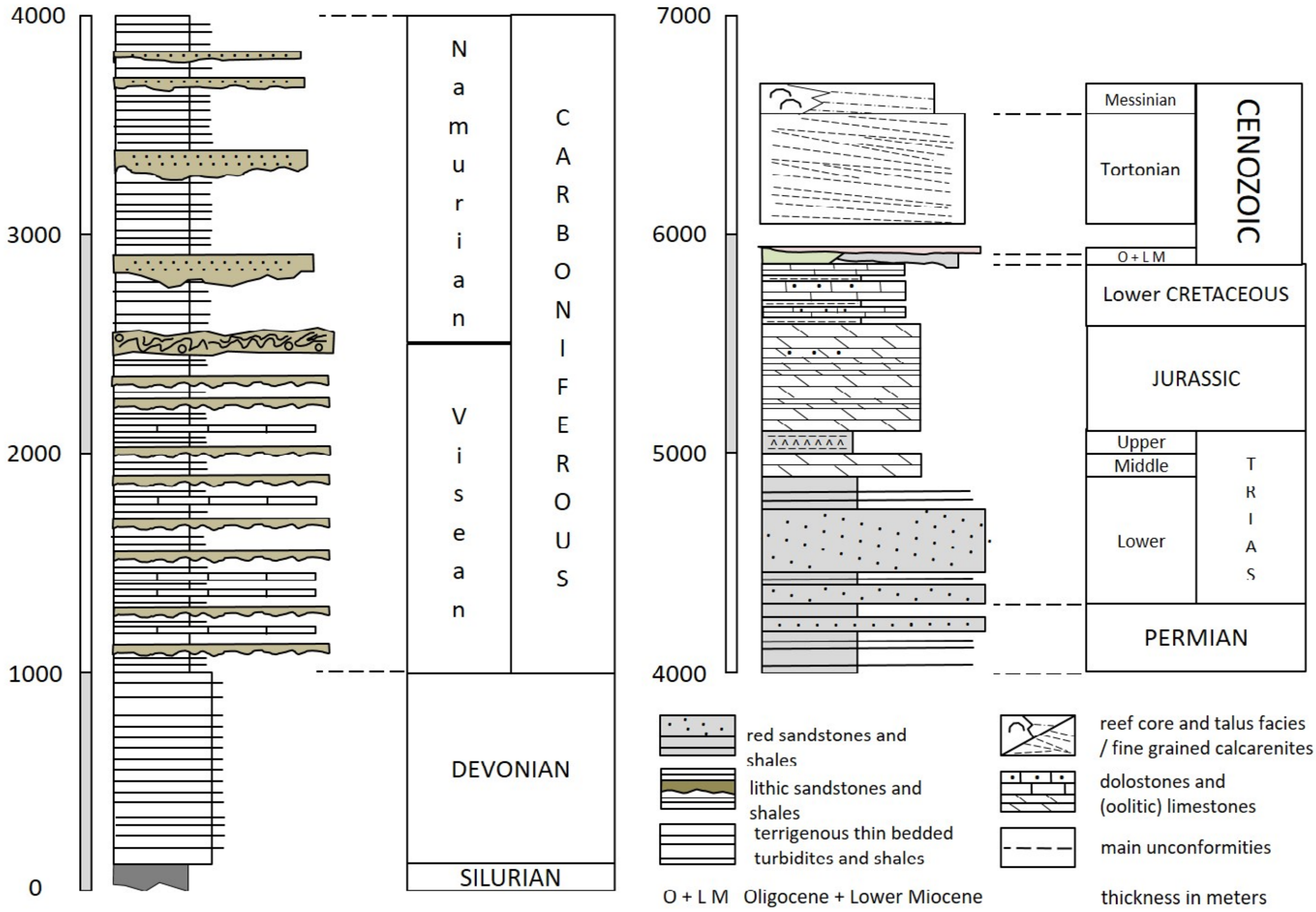

FIGURE 3. Stratigraphic column of Minorca.

sandstone beds generated by proximal turbidites. The lower cycle is mostly Visean in age and the upper cycle has been dated as Namurian.The Minorca Carboniferous attains a total thickness of 3,000m (Fig. 3).

Unconformably overlying Carboniferous and Devonian and the late Variscan paleotopography is a red clastic section that is 650m thick and corresponds to Permo-Triassic (Bourrouilh, 1983; Rodríguez-Perea *et al.*, 1987; Rosell *et al.*, 1990; Linol *et al.*, 2009). The lower part consists of silty-clay layers with intercalated sandstones bars and conglomerate lenses; these sediments correspond to floodplain and pond deposits and belong to the upper Permian (Rosell *et al.*, 1988; Linol *et al.*, 2009). The uppermost part of this succession is truncated by an erosional surface. The sediments above this unconformity are constituted by basal conglomerates followed by red shales and sandstones (Buntsandstein facies) that belong to the Lower Triassic (Bourrouilh, 1983; Rosell *et al.*, 1990). Permo-Triassic rocks display lateral facies transitions and varying thicknesses; the depositional environment was conditioned by extensional faults, resulting in weak tectonic instability that controlled fluvial filling of a system composed of small basins and thresholds (Rodríguez-Perea *et al.*, 1987; Linol *et al.*, 2009).

Through a transitional contact, these materials are overlain by a Middle Carbonate Unit -Muschelkalk facies- of Anisian and Ladinian age which shows only small variations in thickness (Rodríguez-Perea *et al.*, 1987; Llompart et at., 1987). The Upper Triassic corresponds to Keuper facies with iridescent marls and gypsum. The estimated thickness of the two units ranges between 50 and 100m (Fig. 3).

The Jurassic of Minorca is mainly dolomitic; it usually has a thickness of 300m but attains 600m at Cape Cavalleria. The Lower Jurassic (Lias) is composed of massive dolomite beds with stromatholitic structures that are overlain by Toarcian marls. The Middle Jurassic is made up of dolomitic limestones with oolithes, encrinites and quartz grains. The Upper Jurassic also consists of dolomitic limestones (Bourrouilh, 1983).

The Lower Cretaceous outcrops of Minorca are constituted by oolithic and intraclastic limestones and marls. Hardgrounds are also frequent as are some hummocky structures. The lower Cretaceous series has been estimated to be between 120 and 400m thick (Bourrouilh, 1983). Upper Cretaceous outcrops have not been found.

The Cenozoic deposits of Minorca start with few decimeters thick lacustrine and palustrine carbonates of middle Eocene age. They are followed by Oligocene alluvial fans and fan delta conglomerates (Ramos-Guerrero, 1988). Both units are present in small outcrops in the North of the island (*e.g.* Macar de sa Llosa).

Pre-Serravalian Miocene deposits are locally present. They correspond to a transgressive unit, ten-meters thick, made up of reefal limestones (cropping out at Punta Nati, West of Cala Morell) and interfingered with conglomerates. These conglomerates crop out at Cala Morell. They show no sorting and have blocks and rounded cobbles. These conglomerates are massive locally, vary in thickness and show clear intraformational unconformities (Fig. 3). They are interpreted by Bourrouilh (1983) as debris flows. Serravalian deposits composed of boulder-size conglomerates that interfinger with marine calcarenites have also been recorded in Minorca (Fornós *et al.*, 2002).

The Upper Miocene, which is similar to that in Majorca, covers the S and W parts of the island. It displays two depositional sequences (Fig. 3): the lower sequence, which is of Tortonian age (Obrador *et al.*, 1983a, b), is up to 500m thick. This sequence is constituted by cross-bedded pebbly sandstones at the base followed by highly bioturbated marly calcarenites, cross-bedded grainstones, large-scale clinobeds of red-algae calcarenites and marls, and corresponds to prograding ramp-slope deposits (Pomar *et al.*, 2002). The upper sequence, which is of Messsinean age, is a reef complex displaying a reef core and talus facies (Obrador *et al.*, 1983a, b).

Quaternary deposits of Minorca comprise marine terraces and aeolian sheets and dunes.

STRUCTURE

The geological map of Minorca (Fig. 4) shows that the Northeastern part of the island is made up of Mesozoic rocks (mostly Triassic Buntsandstein and Early Jurassic dolostones) but Paleozoic rocks crop out in three areas: i) the westernmost area (located NW of Ferreries) is constituted by a dome of Carboniferous rocks; ii) the central area is a roughly N-S band mainly composed of Carboniferous rocks except for a narrow strip of Devonian rocks close to Mercadal in the East; iii) the eastern area (along the northeast coast) is also made up of Carboniferous rocks.

High-angle normal faults

In the southern part of the island, where upper Miocene limestones crop out, no faults are visible but several sets of joints are present. The main sets strike W-E to WNW-ESE and NNE-SSW (Gelabert *et al.*, 2005).

The boundary between the South and the North of the island is a blind normal fault striking WNW-ESE and

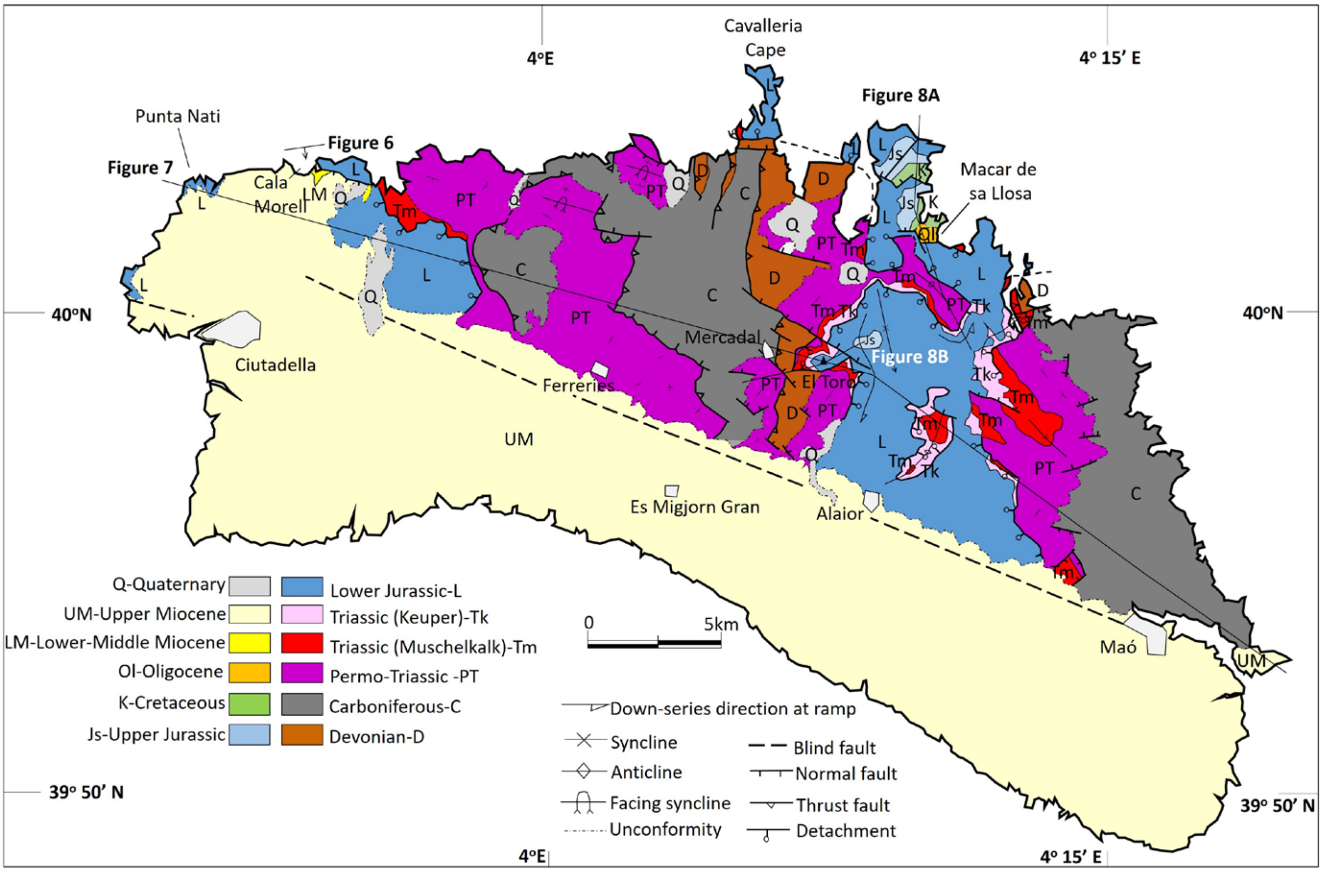

FIGURE 4. Synthetic geological map (modified from Bourrouilh, 1983).

running across the villages of Maó, Alaior and Ferreries (Fig. 4). The downfaulted block is the South. There is a right step relay of the fault that continues through Cala Morell. Here the fault is clearly visible and the pre-Serravalian Miocene conglomerates in the hanging-wall are in clear fault contact with Early Jurassic dolostones in the foot-wall (Fig. 5). The fault plane dips 55º to the SSW and shows grooves and striations with an average pitch of 78º to the East.

In the northern part of the island there is a set of normal faults striking WNW-ESE that are parallel to the boundary-fault between the South and the North (Fig. 4). These faults dip steeply and have a trace that is several kilometers long; in most cases, the down-faulted block is the one in the North. Some shorter faults with slightly different trends (W-E and WSW-ENE) are present too. These high-angle normal faults cut the thrusts and the detachment (Fig. 4), and cut all stratigraphic levels including the pre-Serravalian Miocene conglomerates cropping out at Cala Morell. At this locality, the pre-Serravalian Miocene conglomerates display intraformational unconformities with the result that they are considered syncinematic with the high-angle normal fault. Moreover, normal faults are overlain by upper Miocene rocks (Fig. 5). These high-angle extensional faults are therefore considered to be of early-middle Miocene age.

Bourrouilh (1983) interpreted these faults as left-lateral strike-slip faults because of the apparent left-lateral displacement they produce in older contacts (*i.e.* unconformities, thrust faults) in the western part of the island. We challenge this interpretation because these faults cannot be restored as strike slip faults: i) apparent displacement depends on the dip of the displaced contact and ii) several apparent right lateral displacements are observed in the E of the island (Fig. 4).

Normal faults are also clearly visible at outcrop scale affecting Mesozoic and Paleozoic rocks at different localities in the island. They do not show a systematic orientation (Fig. 6A). Slickensides are present on some of these fault planes and most of the slickensides virtually correspond to dip slip faults.

Thrust faults

In the West and middle of the island there is a set of N-S thrust faults with the hanging-wall to the West tectonic transport direction. Three main thrust faults are identified (from West to East, Figs. 4; 7): i) In the dome NW of Ferreries, Carboniferous rocks are thrust over Triassic Buntsandstein rocks, the trace of the fault is curved and the displacement seems to be small. ii) Carboniferous rocks thrusted over Triassic Buntsandstein rocks, with the N-S striking thrust fault displaced by W-E and WNW-ESE high-angle normal faults; displacement could be as much as one kilometer. iii) Close to Mercadal, Devonian rocks are thrust over Carboniferous and Permo-Triassic rocks, the N-S striking thrust fault is also cut by WNW-ESE normal faults. Other smaller N-S striking thrust faults are present (*i.e*. those along the coast W of Cape Cavalleria).

From the above it follows that these N-S striking thrust faults are younger than Triassic and older than high-angle normal faults.

According to Bourrouilh (1983) these N-S thrust faults are Variscan structures that reactivated during the Cenozoic. The geological cross-section of the island (Fig. 7) shows the Triassic Buntsandstein unconformably overlying the Paleozoic with a marked angularity in the hanging-wall of one of these thrust faults, thus corroborating the hypothesis of Bourrouilh (1983).

Detachment

In line with Bourrouilh (1983), lower Jurassic dolostones are detached on top of Triassic rocks. The main outcrop of the lower Jurassic dolostones is located to the North of Alaior and extends in patches along the coast as far as Cape Cavalleria (Fig. 4). Stratigraphic cut-offs with subtraction of part of the Triassic levels were mapped in all this area and along the contact between the Triassic and Jurassic rocks. These geometries correspond both to foot-wall and hanging-wall ramps. The detachment must be interpreted as extensional because of the missing stratigraphic section. Nevertheless, given the down stratigraphy directions observed in the mapped ramps it is not possible to deduce a consistent direction for the hanging-wall displacement (*i.e*. to the North in the northern parts; to the West at El Toro; to the South, S of El Toro; and between Alaior and Maó to the North and to the Southeast; Fig. 4). This detachment seems to be folded by the WNW-ESE folds.

At Macar de sa Llosa (on the nothern coast), Bourrouilh (1983, his fig. 171) describes in detail the geometry in the hanging-wall of this detachment where Triassic Keuper facies, Jurassic and Cretaceous carbonates and Oligocene conglomerates are present. The Cretaceous carbonates dip around 40º to the South and the Oligocene conglomerates onlap the Cretaceous and dip around 15º. However, Mesozoic and Oligocene beds are involved in a roll-over anticline in the hanging-wall of an E-W extensional fault rooted in the detachment (Fig. 8B). This structure is consistent with top-to-North tectonic transport direction of the detachment.

Briefly, the lower Jurassic dolostones are detached above the Triassic rocks. This detachment is extensional but a consistent hanging-wall tectonic transport direction

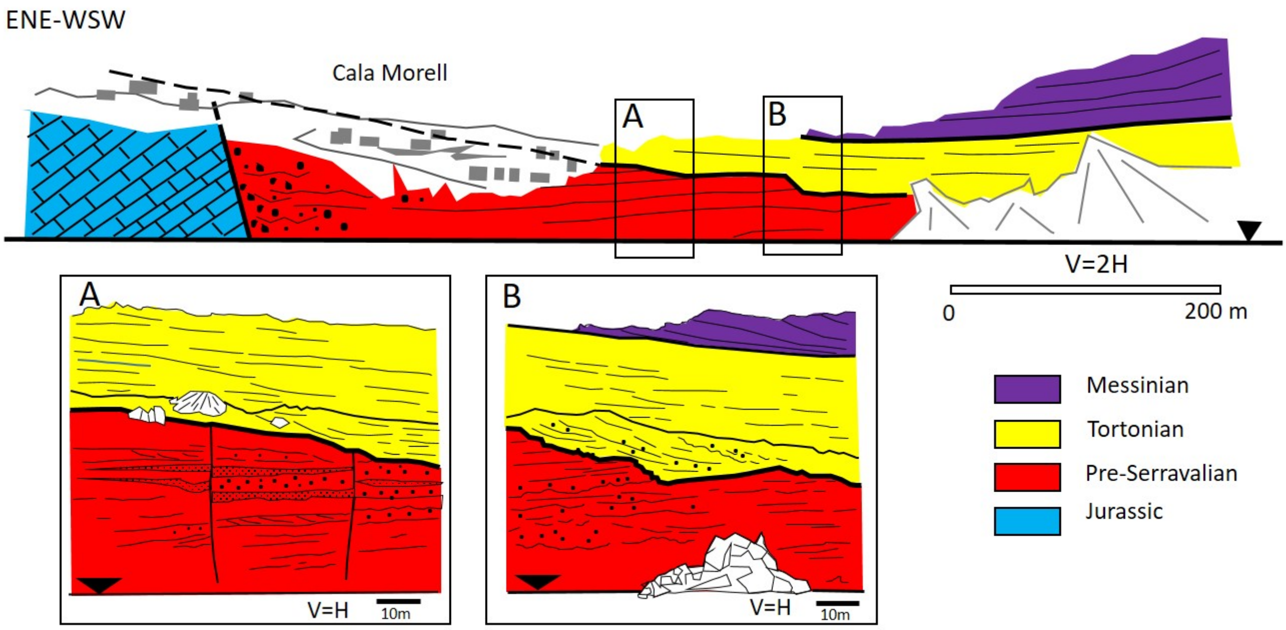

FIGURE 5. Geological section across Cala Morell sketched from a photographic assemblage shot from the sea. Vertical exaggeration 2:1. Location in Figure 4. Modified from Obrador *et al*. (1983c).

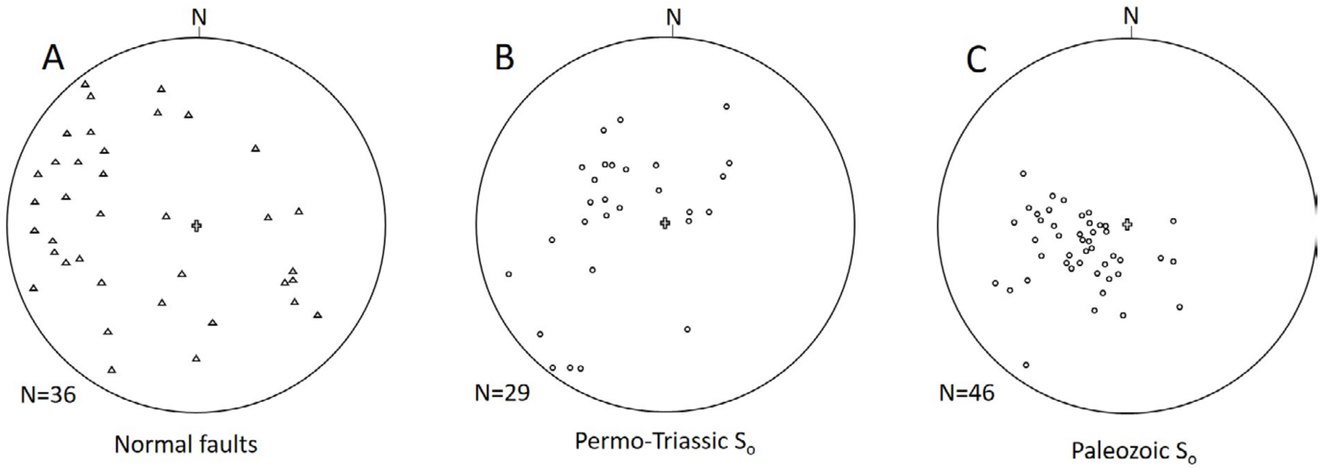

FIGURE 6. Stereogram plots (lower hemisphere, equal-angle projection): A) Poles of normal faults, B) Poles of bedding in Permo-Triassic rocks, C) Poles of bedding in Paleozoic rocks. N: indicate the number of measurements.

cannot be deduced. This detachment is of Oligocene age or younger, and seems to be older than the NW-trending folds (see the anticline located just South of Macar de sa Llosa; Fig. 4).

Folds

At least two sets of folds are visible at map scale (Fig. 4):

i) Two WNW-trending synclines are exposed to the West of Cape Cavalleria and near the coast. They face SW and have one vertical limb. These folds involve the Permian and Triassic Buntsandstein rocks. One of these folds (the one located in a southwestern position) seems to be cut by a N-S striking thrust fault (Fig. 4). Similar oriented folds (a syncline and an anticline) are present to the N and NE of Alaior; they are open folds and involve Triassic rocks and seem to affect the detachment (Fig. 4). Thus, these folds are clearly post-Triassic, and could be post-detachment and pre-thrusts.

ii) To the N of Alaior a set of NNE-SSW folds involves the Triassic and Early Jurassic rocks.

At outcrop scale, bedding is clearly visible in Permian and Triassic rocks and is folded in open and tight folds. The most common bedding dip attitude is toward SE, SW and NE (Fig. 6B). Scatter of bedding data on a stereogram plot is considerable as a result of interference between the two fold sets described above. Nevertheless, the influence of the WNW-trending set of folds is appreciable in the plot.

Orientation patterns of the Permo-Triassic and Paleozoic bedding are markedly different. The predominant dip of the bedding in Paleozoic rocks is NE (Fig. 6C). Comparison of plots in Figure 6B and C shows that the dispersion of the Permo-Triassic bedding poles is greater than that of the Paleozoic bedding poles. This is probably owing to the small number of outcrops of Paleozoic rocks surveyed and because these outcrops are distributed in an small area around Mercadal and along the coast West of Cape Cavalleria.

El Toro

This isolated hill is the highest point in the island (358m). Its structure consists of Jurassic and Triassic beds in the hanging-wall of the extensional detachment described above, which is located in the hanging-wall of a thrust fault. The extensional detachment has a hanging-wall ramp geometry consistent with top-to-SW tectonic transport direction (Fig. 8A, note that the extensional detachment could be mistaken for a thrust fault). The thrust fault is part of the N-S Alpine thrust fault system described above with top-to-West tectonic transport direction and is responsible for the altitude of El Toro (Fig. 8A). A NW-SE normal fault on the northern side of El Toro is partly the reason for the isolation of this hill (Fig. 4).

DISCUSSION

Structural evolution

After Variscan orogeny and Permian extensional instability, the Mesozoic corresponds to an era of relative quiescence. The Triassic and Jurassic rocks do not display variations of facies/thickness or significant structures that could be attributable to major deformation events. The total surface of Cretaceous outcrops is small with the result that the structural evolution of Minorca during this period is not easy to decipher.

Alpine evolution is hampered by the incomplete stratigraphic record of Cenozoic rocks. Nevertheless, cutting relationships between structures described above allow us to establish a relative chronology and an upper and lower bound can be determined by stratigraphy. The oldest Alpine structure is the detachment which is Oligocene or even younger. Detachment is followed by WNW-trending folds (and maybe by NNE-trending folds), by West directed thrust faults and by WNW-ESE high-angle normal faults. These normal faults are early-middle Miocene in age. Thus, contractional structures, folds and thrust faults are probably of early Miocene age (Fig. 2).

Comparison with the areas surrounding Minorca

The Paleozoic stratigraphy of Minorca is similar to that of the southern part of the CCR (in the vicinity of Tarragona). Upper Paleozoic rocks with similar facies are present in both areas. The Paleozoic stratigraphy of Minorca clearly differs from that in the northern part of the CCR (*i.e.* Barcelona) and that in Sardinia where lower Paleozoic and granitic rocks are widespread. Paleozoic rocks crop out extensively in Minorca whereas they are very scarce in Majorca and absent in Ibiza. In this way Minorca differs markedly from the other Balearic islands.

In the Balearic islands (including Minorca) and in Sardinia, the Alpine stratigraphic gap comprises at least the Paleocene-early Eocene and usually includes part or most of the Upper Cretaceous and most of the Paleogene.This gap in the eastern continental margin of Iberia (including Balearic and Corsica-Sardinia blocks) is probably due to uplift associated with the onset of the Africa-Eurasia convergence in the Late Cretaceous. Moreover, middle Miocene turbidites are widespread in Majorca and Ibiza but are absent in Minorca. This is another important difference between Minorca and the other Balearic islands.

Near-bedding parallel subtractive tectonic contacts have been described in the eastern ranges of Majorca (Sàbat, 1986) and in the CCR (*i.e.* in Corbera de Llobregat near Barcelona, P. Santanach, personal communication; Instituto Geológico y Minero de España, 1975). These contacts have affinities with the extensional detachment in Minorca, which could be attributed to the spread of the continental margin as described in Rowan *et al.* (2012). This interpretation is based on two observations: i) the fact that the extensional detachment is coeval with rifting and the formation of the continental margins in the NW of the Mediterranean and ii) the fact that a consistent hanging-wall tectonic transport direction cannot be deduced for the extensional detachment in Minorca.

The Eocene-Oligocene NW-SE or/and the early Eocene W-E and Oligocene-Aquitanian N-S compression in Sardinia and the Paleocene to early Oligocene E-W compression in Corsica were, respectively, N-S, NW-SE and NE-SW, and NW-SE directed when the Corsica-Sardinia block is restored to its pre-Miocene position. Compression in Corsica and Sardinia is associated with the Pyrenean and Alpine deformations as shown by its direction and age, and has no equivalent in Minorca. Compression in Corsica is clearly older and in Sardinia is partly older than and partly coeval with that in Minorca. Early-middle Miocene NW-SE compression in Ibiza and Majorca has no counterpart in Minorca where the most marked compressional structures are N-S striking, W-directed thrust faults the age of which could be early Miocene. Minorca therefore differs from the other Balearic islands in the orientation of the compression, and from Corsica and Sardinia in the age and orientation of the compression (Fig. 2).

Cenozoic folds observed in Minorca are generally open except for the two WNW-trending synclines described above. These folds are parallel to but older than the high-angle normal faults discussed above, and are of the same age and are parallel to some folds in Majorca. Sàbat *et al.* (1988) interpreted the NW-SE folds of the eastern ranges of Majorca as oblique ramps of the thrust system. At present, it is not possible to offer a hypothesis about the origin of the WNW-trending folds of Minorca.

The early-middle Miocene WNW-ESE high-angle normal faults in Minorca are parallel to the NW-trending fault system between the Liguro-Provençal and Algerian basins (Fig. 1). This boundary is a right lateral transfer zone that bounded the opening of the Liguro-Provençal Basin by the counter clockwise rotation of the Corsica-

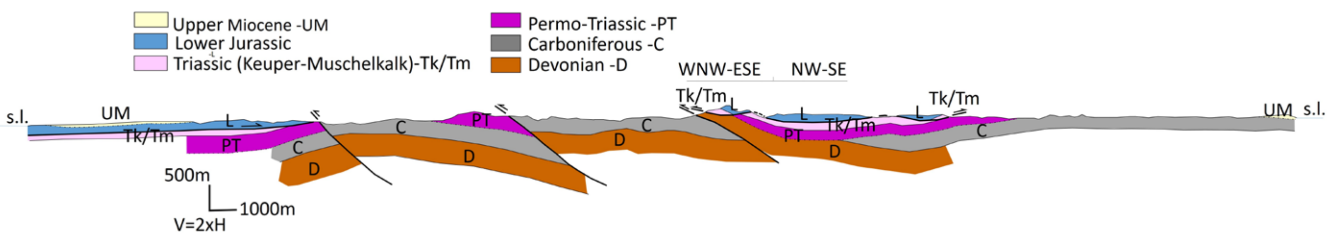

FIGURE 7. Geological cross-section along Minorca. Vertical exaggeration 2:1. Location on Figure 4.

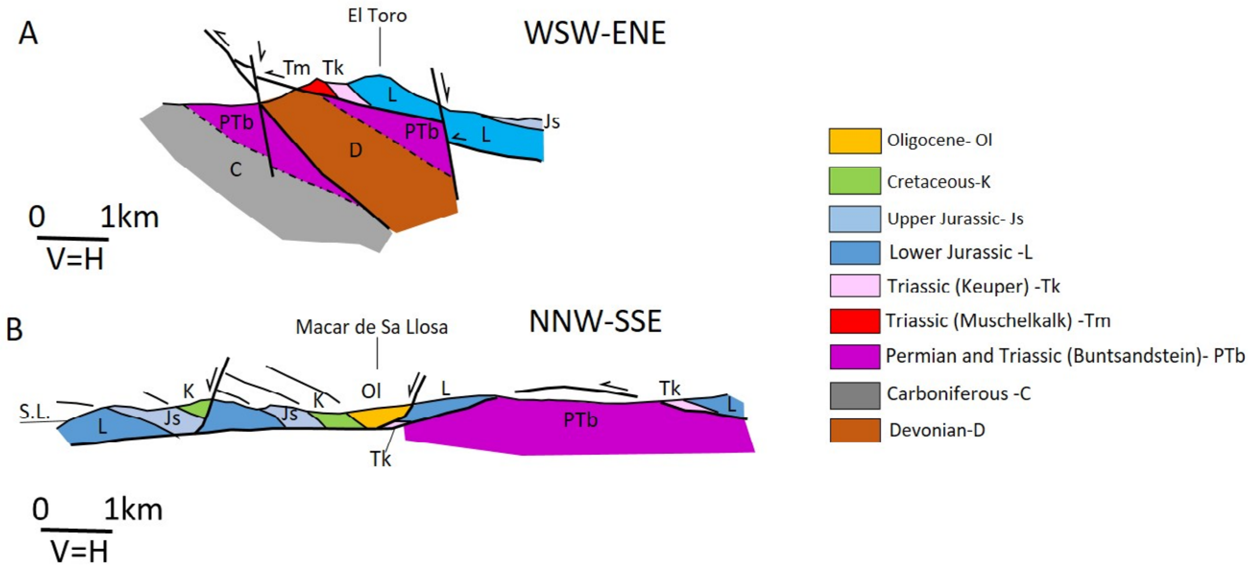

FIGURE 8. Local geological cross-sections: A) El Toro, B) Macar de sa Llosa. Location in Figure 4.

Sardinia block during the early Miocene. This fault system can be followed northwestward along the northeastern edge of the Valencia trough and the CCR by a set of NW-trending extensional faults. These faults control Neogene to Quaternary alkaline volcanism and could be interpreted as transversal faults of the Miocene rift. NW- or WNW-trending faults are also present in the eastern ranges of Majorca. Sàbat *et al*. (1988) interpreted these faults as transfer faults of the Oligocene-Miocene thrust system. However, it is our belief that the WNW-ESE high-angle normal faults of Minorca are part of the afore mentioned fault system and that they are due to stretching of the eastern continental margin of Iberia. This stretching is linked to the opening of the western Mediterrranean as a result of back-arc processes, and especially to the opening of the Liguro-Provençal Basin owing to the rotation of the Corsica-Sardinia block. Accordingly, the island of Minorca can be interpreted as an extensional ridge parallel to the fault system.

The shape of Majorca in map view is outlined by the structural grain determined by both contractional and extensional NE-trending structures, whereas Minorca is WNW-ESE elongated and controlled by WNW-trending high-angle normal faults (Fig. 1).

Minorca differs stratigraphically and structurally from the other Balearic islands. Moreover, the island cannot be considered to be a small block dragged by the early Miocene counter clockwise rotation of the Corsica-Sardinia block since it displays Neogene clockwise rotation. Given its stratigraphic affinity with the southern part of the CCR, Minorca, of all the Balearic islands, traveled the farthest during the rifting in the Valencia trough. For this reason, it must be interpreted as part of the Balearic foreland as suggested by Fallot (1923).

CONCLUSIONS

i) The Paleozoic and Mesozoic stratigraphy of Minorca island is akin to that encountered in the southern part of the CCR (Tarragona). Moreover, there are significant differences between the stratigraphy of Minorca and that of the other Balearic islands. In addition, the Paleozoic deposits of Sardinia do not correlate with those in Minorca; however, the Mesozoic series of the two islands are fairly similar except for the Triassic sediments which are scarce in Sardinia.

ii) Eocene-Oligocene sediments are very scarce in Minorca and Sardinia whereas they crop out in Majorca.

iii) Oligocene-Miocene structures of Minorca are characterized by an extensional detachment, two sets of mostly open folds, W-directed thrust faults and WNW-ESE high-angle normal faults.

iv) Contractional structures of Minorca differ in geometry from those in the other Balearic islands and from the those in Sardinia.

v) Early-middle Miocene WNW-ESE high-angle normal faults of Minorca form part of a complex system, including normal faults and right-lateral strike slip faults. This fault system acted as a right lateral transfer zone that bounded southwestwards the spread of the Liguro-

Provençal Basin and the rotation of the Corsica-Sardinia Block.

vi) Minorca is an extensional ridge parallel to the aforementioned fault system.

vii) Upper Miocene stratigraphy of Minorca is similar to that of Majorca since both islands are adjacent to each other after the rifting in the Valencia trough.

viii) Minorca resembles an exotic Balearic island because it probably forms part of the Balearic foreland. Of all the Balearic islands, Minorca travelled the farthest

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