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Structure and kinematics of the South Iberian paleomargin and its relationship with the Flysch Trough units: extensional tectonics within the Gibraltar Arc fold-and-thrust belt (western Betics)

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Abstract

The superposition of structures observed in a selected area of the external zones of the Betic orogen, situated in the westernmost segment of the peri-Mediterranean mountain belt, reveals an alternation of compressional and extensional events during the overall convergence of the African and Eurasian plates that bound the system. An Early Miocene compressional event produced the formation of the so-called Gibraltar Arc fold-and-thrust belt, formed by the units derived from the South Iberian paleomargin and the Flysch Trough. This wedge constitutes the footwall of the Gibraltar Thrust, whose hanging wall corresponds to the Alboran Domain. The outward migration of the compressional front (W-to-NWward in the South Iberian segment) was followed by the outward migration of the extensional locus, related to the Middle Miocene rifting that led to the Alboran Basin subsidence. The tectonic inversion of the Gibraltar Thrust was accompanied by normal faulting of the innermost part of the paleomargin and the Flysch Trough units. Finally, the Alboran region underwent N–S-to-NW–SE compression from Late Miocene onwards, which caused the uplift of the Gibraltar Arc area.

1. Introduction

The Betic and Rif chains, north and south of the Alboran Sea, respectively, are linked by the Gibraltar Arc (Fig. 1), which developed during, and partly in response to, Cenozoic convergence between Africa and Iberia (e.g. Dewey et al., 1989). This westernmost segment of the Alpine–Mediterranean orogenic belt is of great interest for the study of contractional processes contemporaneous with extensional ones. The rifting that took place in the internal part of the Gibraltar Arc and led to the formation of the Alboran Basin occurred while an arcuate fold-and-thrust belt (now emerged) developed in its external part during the Miocene (e.g. Bouillin et al., 1986; Balanyá and García-Dueñas, 1988; Platt and Vissers, 1989; Comas et al., 1992; García-Dueñas et al., 1992). This belt involved: (a) the South Iberian and Maghrebian paleomargins (Fig. 1), separated by a transform fault that was active during the Jurassic and Early Cretaceous (e.g. Dercourt et al., 1986) and is made up of Triassic to Neogene deposits (e.g. García-Hernández et al., 1980; Wildi, 1983); and (b) the Flysch Trough units (Fig. 1), containing Cretaceous to Miocene siliciclastic deposits (e.g. Chauve, 1968; Didon, 1969), sedimented in a deep basin with attenuated continental crust or possibly oceanic crust (Biju-Duval et al., 1978; Dercourt et al., 1986; Durand-Delga et al., 2000).

This fold-and-thrust belt constitutes the footwall of the so-called Gibraltar Crustal Thrust, whereas the hanging wall is the Alboran Domain (Fig. 1) (Balanyá and García-Dueñas, 1988), a post-metamorphic thrust stack formed mainly by Paleozoic to Triassic rocks, with a complex tectonometamorphic evolution (e.g. Balanyá et al., 1997; Azañón and Crespo-Blanc, 2000 and references therein). It is generally acknowledged that shortening in front of and below the Gibraltar Crustal Thrust was due to the westward migration of the Alboran Domain from the Early Miocene (e.g. Andrieux et al., 1971; Bourgois, 1978; Durand-Delga, 1980; Leblanc and Olivier, 1984; Martín-Algarra, 1987; García-Dueñas et al., 1993; Royden, 1993; Lonergan and White, 1997). Shortening obliterated the Flysch Trough, and sedimentary infill detached from its (unknown) basement was incorporated into the Gibraltar Arc fold-and-thrust belt (Balanyá and García-Dueñas, 1988). The migration of compression below the Gibraltar Thrust produced the inclusion into the wedge of the sedimentary covers of the South Iberian and Maghrebian paleomargins. The transport direction during shortening swings

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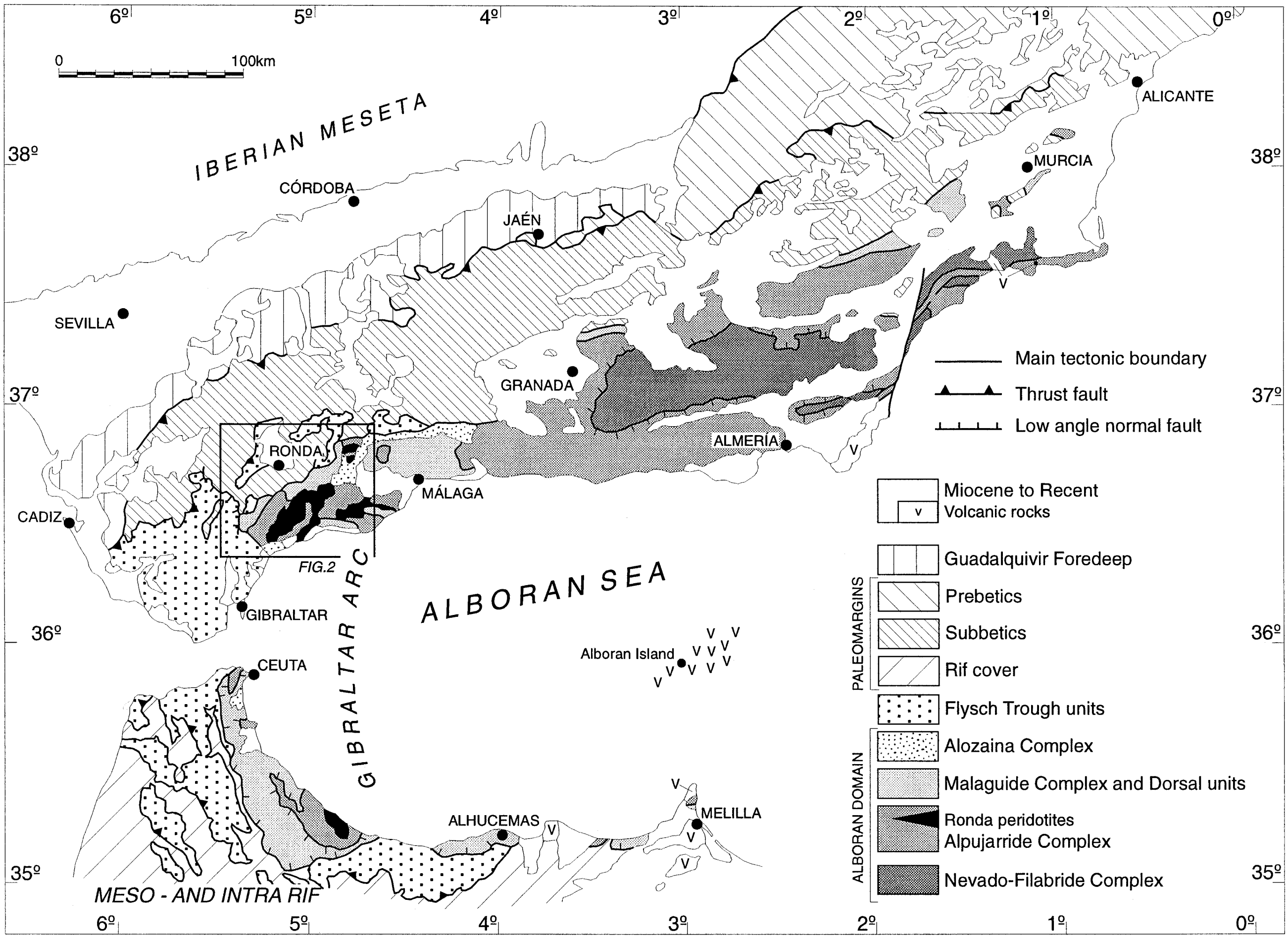

Fig. 1. Crustal domains and tectonic complexes of the Gibraltar Arc, western Mediterranean Sea.

around the Gibraltar Arc (Crespo-Blanc et al., 2001): it is broadly NW-directed in the northeastern part of the Arc (this paper), W-directed in its central sector (Luján et al., 2000) and SE-directed in the southeastern part (Morley, 1987). Finally, foreland basins with olistostrome deposits developed during the upper Miocene in the northern and southern segments of the Betic–Rif orogen (Guadalquivir and Rharb Basins, respectively).

The westward movement of the Alboran Domain continued after its thrusting over the paleomargins, due to an extensional fault system producing approximately westwards hanging wall movement. This is one of the main extensional systems associated with the Miocene rifting that produced the thinning of the Alboran Domain, leading to the formation of the Alboran Basin in the inner part of the Gibraltar Arc during the Early and Middle Miocene (Comas et al., 1992, 1999; García-Dueñas et al., 1992).

The aim of this paper is to shed light on the different tectonic events—whether compressional or extensional—involved in the Gibraltar Arc fold-and-thrust belt, and to analyze their kinematics in a selected area of the western Betics (Fig. 2). Special attention is given to the structure of the most internal unit of the Iberian paleomargin (Internal Subbetic) and its structural relationship with the Flysch Trough units. It will be shown that the propagation towards the external part of the Arc of the contractional deformation that produced the Gibraltar Arc fold-and-thrust belt is followed by an extensional event that invades the wedge. This event is most likely related with the outward migration of extension, from the Alboran Domain towards the external zones. The data presented in this paper will help establish the evolutionary model of the Gibraltar Arc during the Miocene, key to our understanding of the western Mediterranean Alpine system.

2. The Gibraltar Arc fold-and-thrust belt in the western Betics

Our study area, to the E and N of Ronda, includes the boundaries between the Alboran Domain, the Flysch Trough and the Subbetic units (Figs. 2 and 3). The structural relationship between these different domains can be observed in the cross-section of Fig. 2 and in the panoramic view of Fig. 5a. In the study area, the boundary between the Flysch Trough and the Subbetic units is sealed by the Ronda

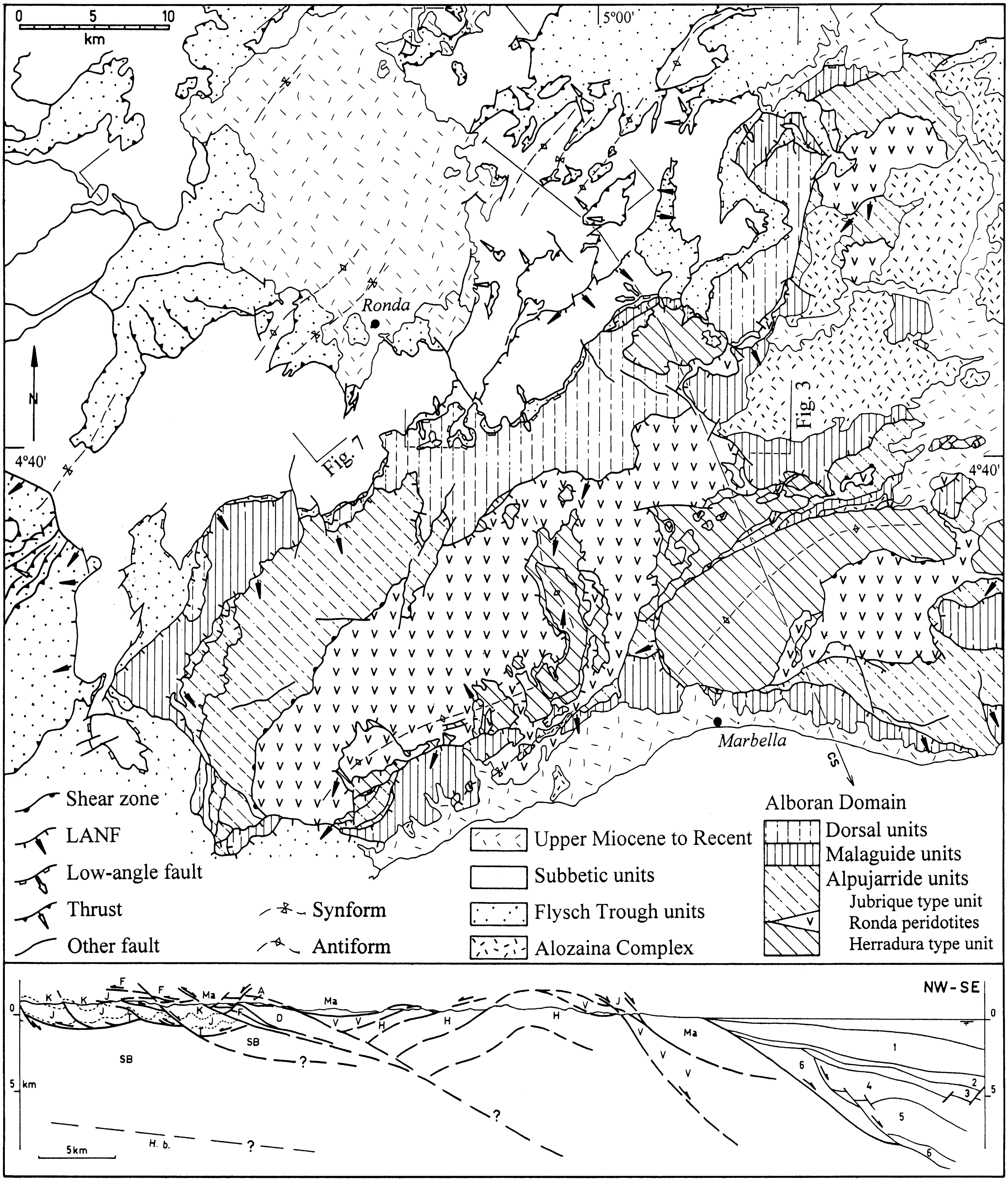

Fig. 2. Tectonic units in the western Betics [modified after García de Domingo (1994), Junta de Andalucía (1985), Cano Medina (1990), Cruz San Julián (1990), Del Olmo Sanz et al. (1990) and Moreno Serrano et al. (1990)]. Area shown in Fig. 1. Arrows: kinematic indicators along low-angle faults [in the Alboran Domain, according to Soto and Gervilla (1991), García-Dueñas et al. (1992), Balanyá et al. (1997) and Sánchez-Gómez (1997); in the Flysch Trough units, according to Luján et al. (1999, 2000)]. Synform and antiform: upper Miocene very open folds (in the Alboran Domain, according to Sánchez-Gómez, 1997). Cross-section (cs): structural relationships between the Internal Subbetic units (T, Triassic; J, Jurassic; K, Cretaceous to Paleogene) that belong to the South Iberian paleomargin, the Flysch Trough units (F), the Alboran Domain (D, Dorsal units; Ma, Malaguide; A and H, Alpujarride units; V, peridotites) and the sedimentary deposits of the Alboran Basin, according to Comas et al. (1992, fig. 5, op. cit.) (1, Pliocene to Quaternary; 2, Messinian; 3, Tortonian to upper Serravallian; 4, Serravallian; 5, lower Serravallian to Langhian; 6, Burdigalian to Aquitanian). Top of the Hercynian basement (H.b.) extrapolated from Blankenship (1992).

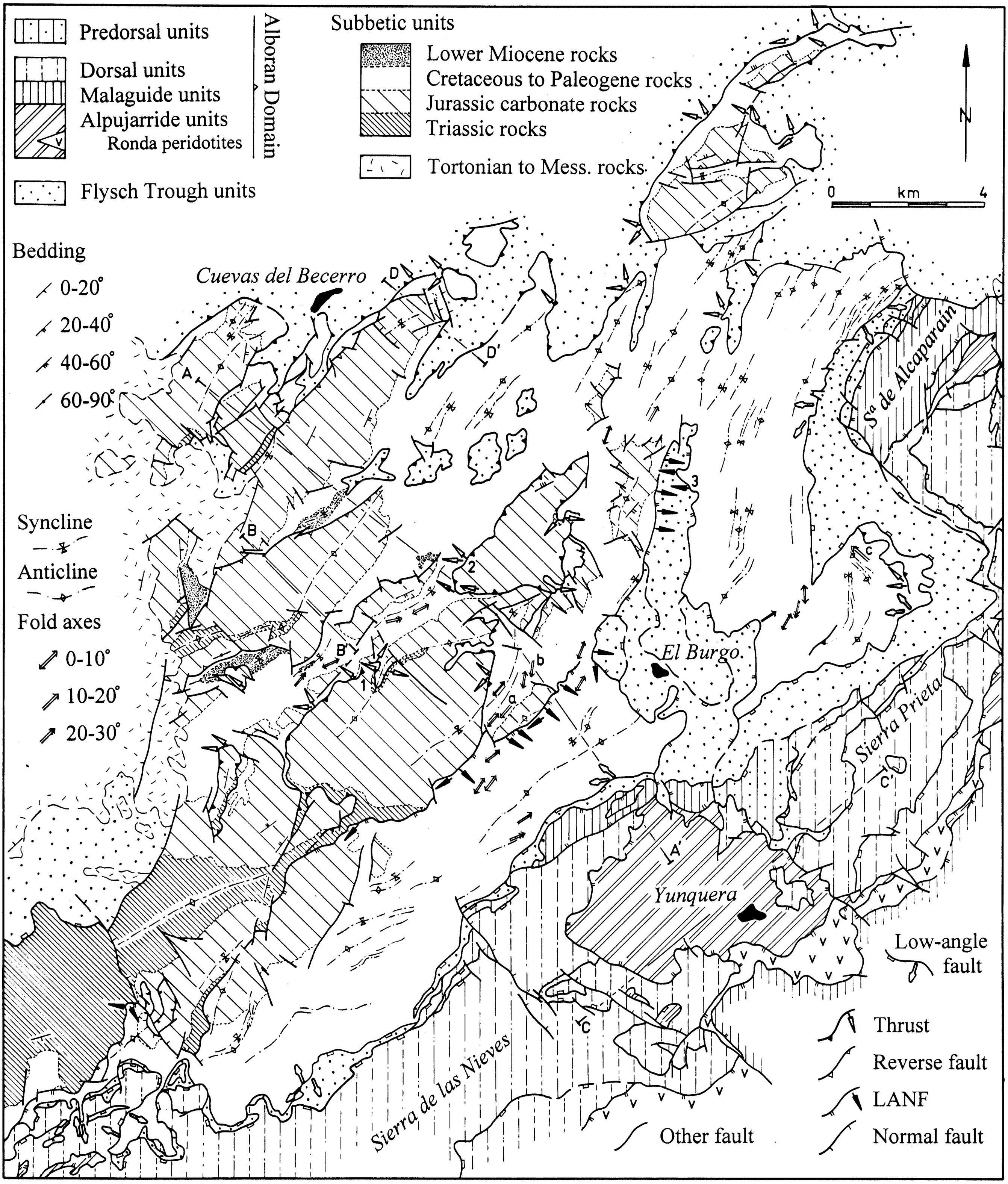

Fig. 3. Geological and structural map of the Internal Subbetic units in contact with the Flysch Trough and the Alboran Domain units, east of Ronda [modified after Cruz San Julián (1990) and Del Olmo Sanz et al. (1990)]. Location shown in Fig. 2. Syncline and anticline: early to middle Miocene closed folds. Arrows: kinematic indicators along low-angle faults. LANF: low-angle normal fault. 1, 2, 3 and a, b, c: localization of stereoplots of Figs. 6 and 8.

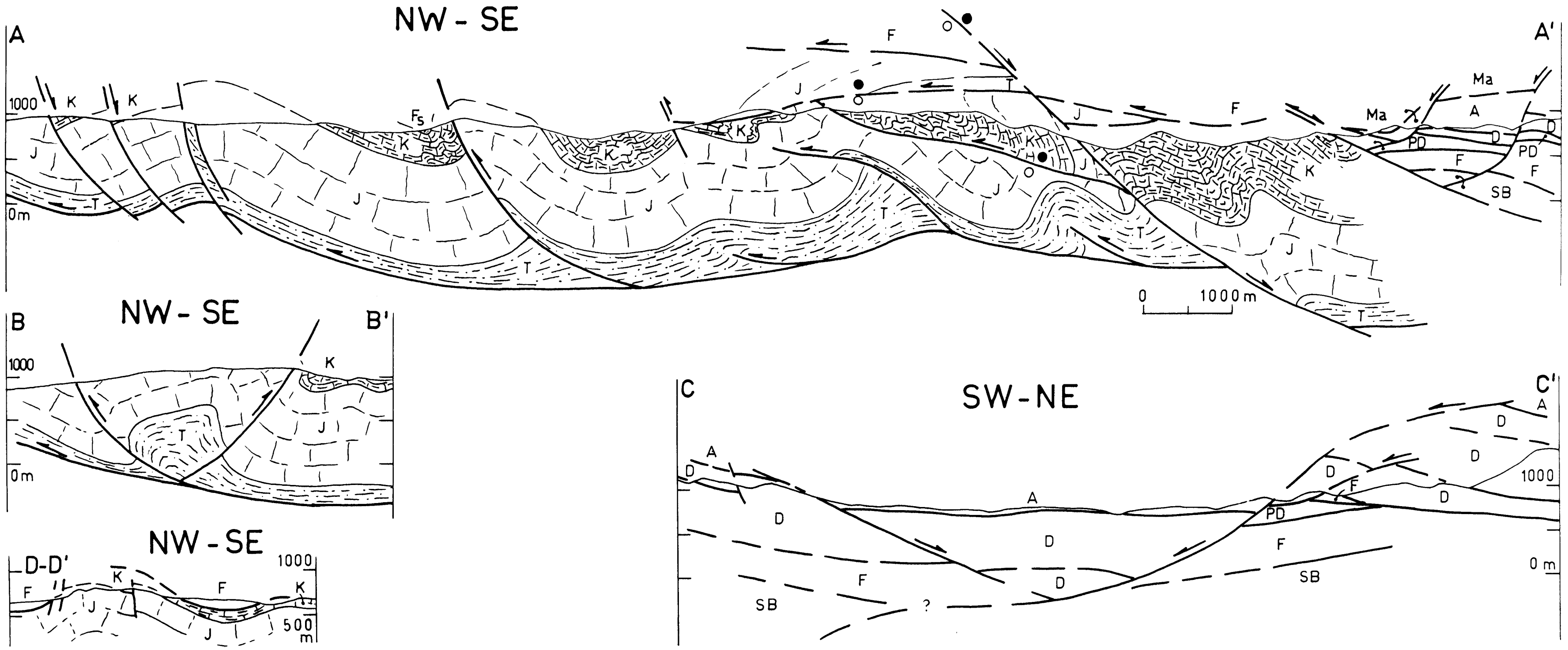

Fig. 4. Geological cross-sections: sections A–A′ and B–B′ illustrate the style of early to middle Miocene closed folds and thrusts in the Internal Subbetic units (T, Triassic; J, Jurassic; K, Cretaceous to Paleogene; SB, Internal Subbetic units) and Flysch Trough units (F); section C–C′ shows low-angle normal faults that affect the Alboran Domain (PD and D, Predorsal and Dorsal units respectively; Ma, Malaguide; A, Alpujarride units) and which can be followed in the Gibraltar Arc accretionary prism; section D–D′ shows the upper Miocene open folds defined by the boundaries between the Flysch Trough and the Internal Subbetic units. Localization in Fig. 3.

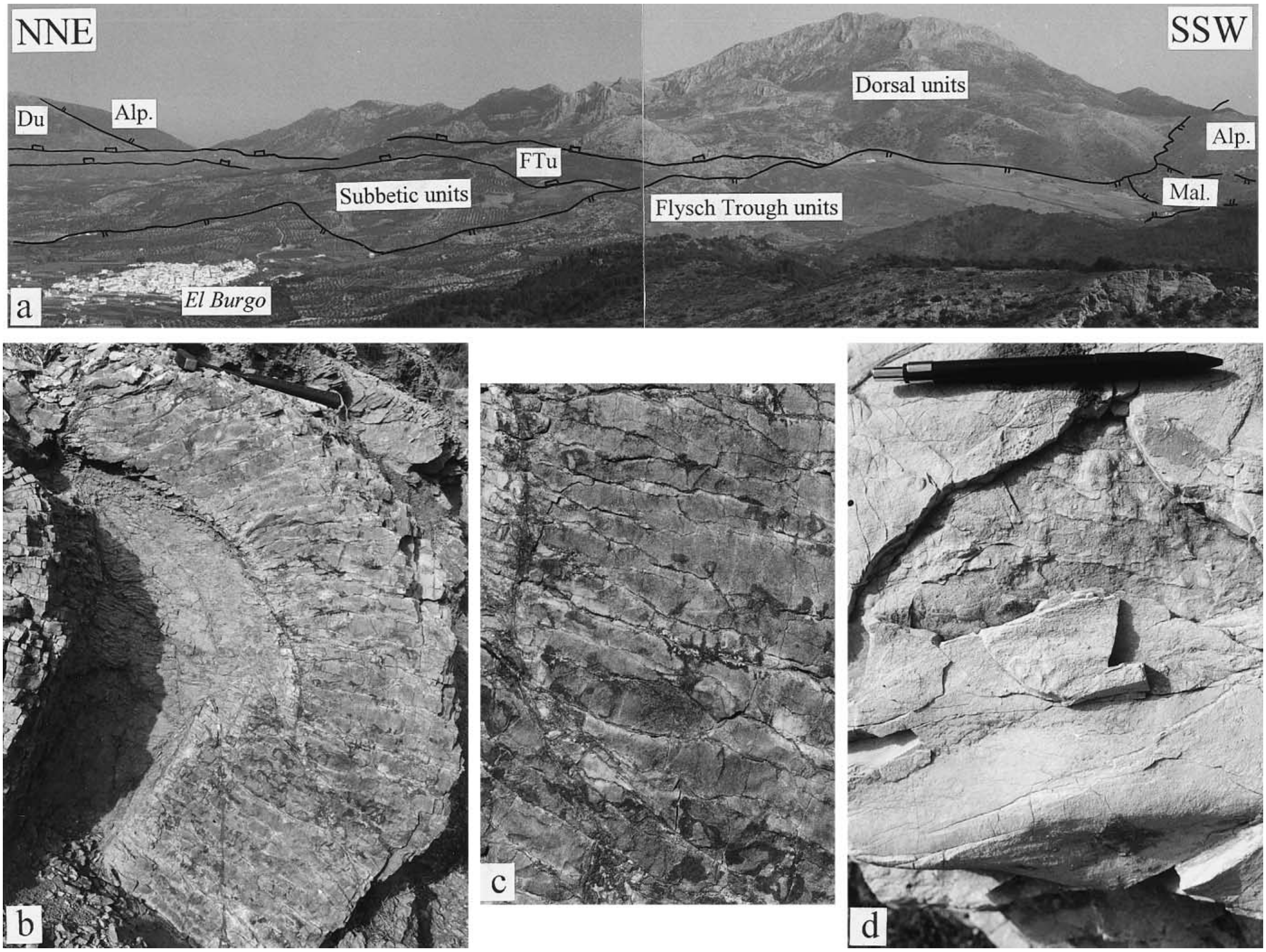

Fig. 5. (a) Panoramic view illustrating the structural relationship between the Alboran Domain (Alp., Alpujarride units; Mal., Malaguide units; Du, Dorsal units), the Flysch Trough units (FTu) and the South Iberian paleomargin (Subbetic units). Note that in the SSW part of the photograph, a middle Miocene low-angle normal fault produced the downthrow of the Alboran Domain with respect to the Dorsal units (see cross-section C–C′ of Fig. 4). (b) Lower to middle Miocene closed fold developed in the upper Cretaceous to Paleogene 'red beds', and related with main shortening in the Internal Subbetic units. Observe the associate cleavage and its fan pattern in the calcareous competent bed. (c) Detail of the central part of Fig. 5b, which illustrates the cleavage marked by insoluble material. (d) Evidence of slip parallel to the bedding, probably induced by a mechanism of flexural slip folding, on the limb of a fold whose axis appears in the lower part of the photograph.

Basin, an intramontane basin formed by Late Tortonian and Messinian sediments (Fig. 2).

2.1. Internal Subbetic units

The Subbetic Zone is formed by unmetamorphosed rocks of Triassic to Neogene age, in which a marked differential subsidence occurred during the Jurassic, leading to troughs (Median Subbetic) and swells (External and Internal Subbetic) (García-Hernández et al., 1980). The Subbetic units that crop out in the studied area belong to the Internal Subbetic, and their lithostratigraphy is characterized from bottom to top as (Bourgois, 1978; Martín-Algarra, 1987):

1. a Triassic sequence (minimum thickness of 450 m) made up of limestones and dolomitized carbonate rocks (Muschelkalk facies), together with gypsiferous claystones and fine-grained sandstones (Keuper facies);
2. a Jurassic sequence (minimum thickness of 600 m) formed by dolomite rocks superposed by massive, predominantly shallow marine limestones;
3. an Early Cretaceous unconformity marked by hard ground on top of the Jurassic limestones and multiple periods of karstification between the Berriasian and the Albian;
4. an alternating of pink pelagic marls and marly limestones, Late Cretaceous to Paleogene in age ('red beds', minimum thickness of 350 m);
5. an Early Miocene flysch-type sequence (minimum thickness of 25 m), made up of clays alternating with a few fine-grained sandstone levels, the uppermost dated as Early Burdigalian (Dubois and Magné, 1972).

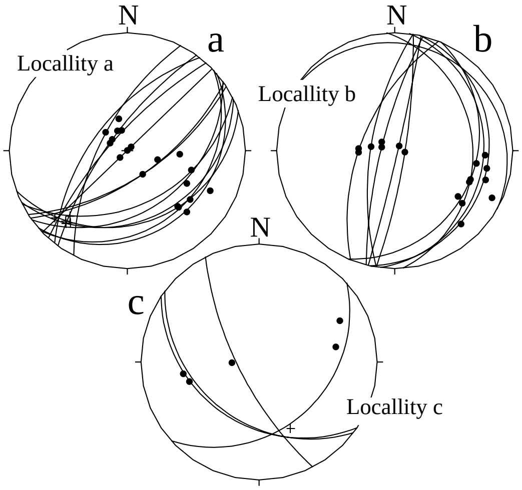

Fig. 6. Stereographic plots illustrating the transport direction during the folding of the sedimentary rocks by a mechanism of flexural slip (localities a to c situated in Fig. 3). Great circles, bedding; black circles, calcite fibers within bedding; crosses, fold axis.

The structure of the Subbetic units in the study area can be described as a NW-vergent fold-and-thrust belt, post-Early Burdigalian in age, deduced from the age of the youngest sedimentary rocks involved (see Fig. 3, 10 km W of El Burgo). The local deformation style of the folds is strongly controlled by the rheology of the Mesozoic sequence, and a disharmony along the boundary between Jurassic and Cretaceous rocks can be observed: the massive Jurassic dolomites and limestones define 3–5-km-wavelength synclines and anticlines (Figs. 3 and 4, cross-section A–A′), whereas the Cretaceous to Paleogene ‘red beds’ form metric- to decimetric-scale folds, coherent with the large-scale structure, with an interlimb angle that generally ranges between 60 and 110° (Fig. 5b). Cross-section A–A′ of Fig. 4 shows these folds to be NW-vergent, their axial plane normally dipping strongly towards the SE, although some box-folds are present (Fig. 4, cross-section B–B′). On the small scale, in the carbonate rocks and in particular in

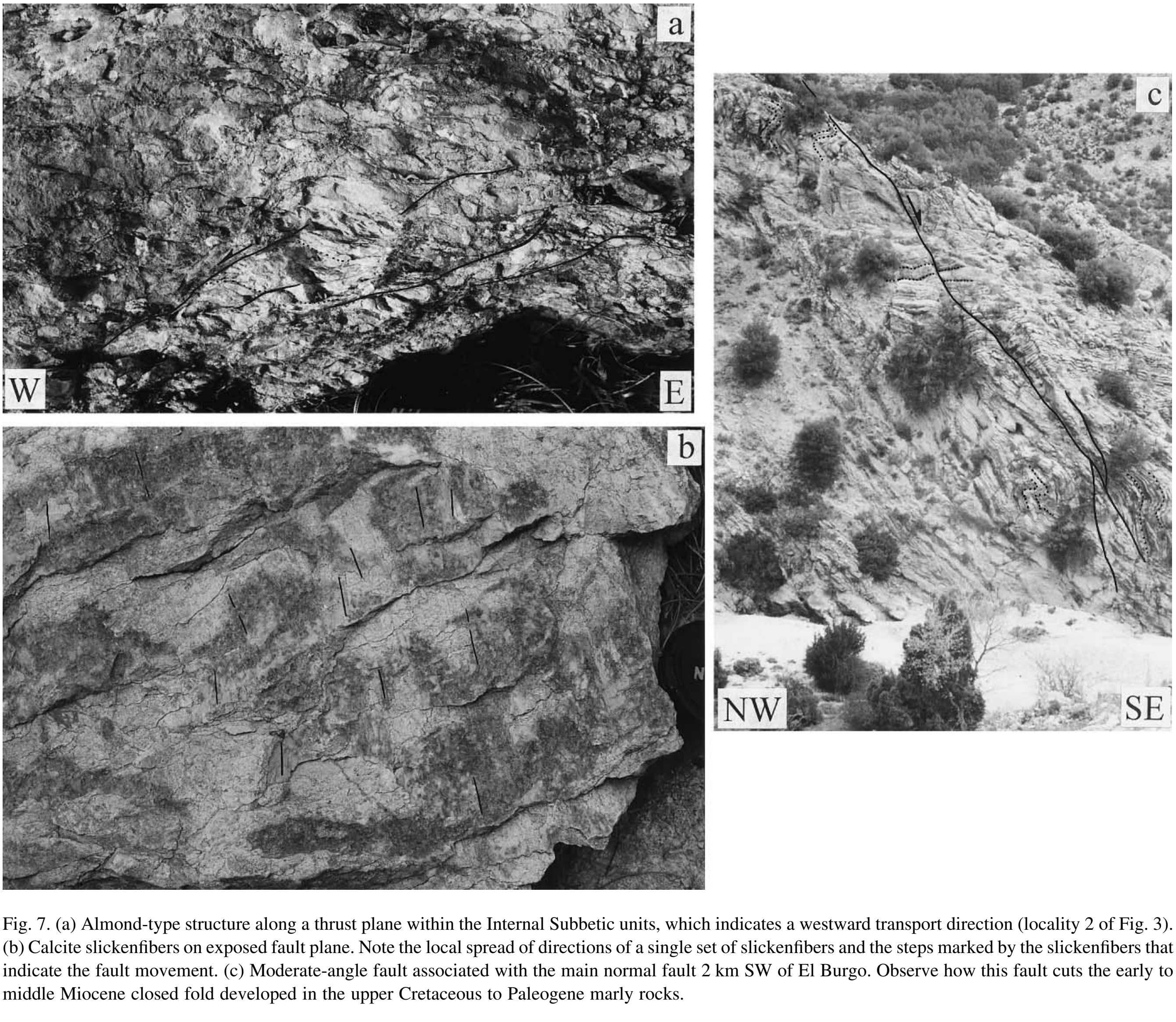

Fig. 7. (a) Almond-type structure along a thrust plane within the Internal Subbetic units, which indicates a westward transport direction (locality 2 of Fig. 3). (b) Calcite slickenfibers on exposed fault plane. Note the local spread of directions of a single set of slickenfibers and the steps marked by the slickenfibers that indicate the fault movement. (c) Moderate-angle fault associated with the main normal fault 2 km SW of El Burgo. Observe how this fault cuts the early to middle Miocene closed fold developed in the upper Cretaceous to Paleogene marly rocks.

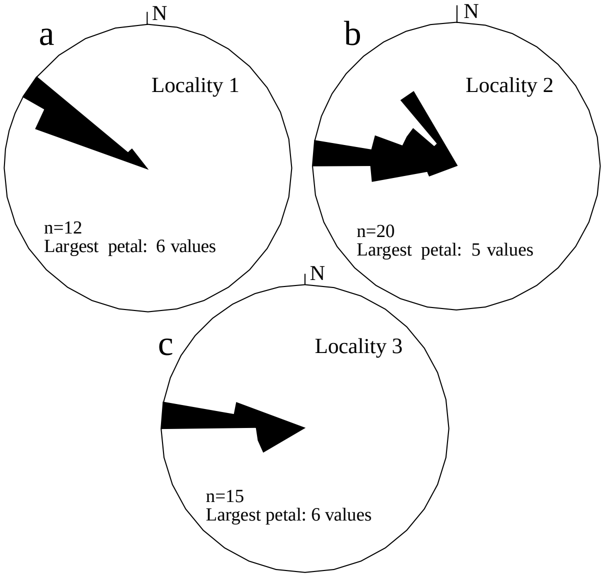

Fig. 8. Rose diagrams of the direction and sense of movement of kinematic indicators situated along low-angle faults (localities 1 to 3 situated in Fig. 3). (a) Single set of northwestward kinematic indicators along a thrust plane within the Internal Subbetic units. Observe the 30° local spread of direction. (b) Two sets of kinematic indicators, northwestward and westward directed, along the same thrust plane within the Internal Subbetic units. (c) Single set of westward kinematic indicators along a low-to-moderate angle fault which bounds the Flysch Trough and the Subbetic units.

the marly 'red beds', a cleavage of centimetric scale spacing is associated with the folds (Fig. 5b). Its insoluble (argillaceous) composition (Fig. 5c), together with a frequent stylolitic geometry, point to a pressure-solution origin for this cleavage. In the fold hinge zone, it is convergent in the inner part of the folds in carbonate beds (Fig. 5b), while divergent in more marly beds. Evidence of slip parallel to the bedding within the 'red beds' is frequent (Fig. 5d). Opposite shear senses can be seen on both fold limbs, suggesting a mechanism of flexural slip during the fold development.

Fig. 3 shows the trend of the axial traces, drawn from 1:20.000 scale aerial photographs and from fold axes measured in the field, in particular in the 'red beds'. These fold axes appear mainly NE–SW-directed. Nevertheless, locally—for instance around Sierra de Alcaparaín—these axes may veer to a NW–SE direction, almost parallel with the irregular design of the Alboran Domain (Fig. 3). Moreover, over a stretch of 2 km, a single fold may be seen to swing from a N–S to a NE–SW axial direction (3 km NE of El Burgo). The fold axes are subhorizontal or plunge slightly, and there are no morphological variations between the NE–SW- and the NW–SE-oriented folds. It is therefore suggested that these folds do not represent two distinct fold generations, but rather formed during a single deformation episode. The fold axes were modified by subsequent vertical axis rotations, at least locally, most likely produced by late thrusts (see below). Significant is the fact that where the fold axes are locally NW–SE-directed, 6 km NE of El Burgo (Fig. 3), the main movement along the surrounding late thrusts is towards the NW, as in most of the thrusts of the study area. This is consistent with the measurement of slip running parallel to the bedding, unambiguously associated with the fold development involving a mechanism of flexural slip. In localities with fold axis orientation varying from NE–SW to NNW–SSE, the mean direction of the slickenfibers measured is subperpendicular to the fold axis, regardless of its orientation (Fig. 6). If the N–S to NNW–SSE trending folds had developed as lateral structures during a main NW–SE shortening, they would show oblique transport direction with respect to the fold axes.

The described folds are cut by thrusts, in-sequence and out-of-sequence, located within the Internal Subbetic units (Figs. 3 and 4, cross-section A–A′). The in-sequence thrusts are rooted in the gypsiferous claystones and fine-grained sandstones of Triassic age, suggesting that a sole thrust is present within the evaporites. Out-of-sequence thrusts are illustrated in the southeastern part of cross-section A–A′ (NW of the main normal fault, Fig. 4). The latter show an oblique transport direction with respect to the cross-section (see below and Fig. 3), cutting both limbs of previous synclines and bounding unrooted slices of Subbetic rocks.

Within the thrust fault zone (up to 50 m in width, but generally around 10 m), several small scale structures—including almond-type structures (Fig. 7a), slickenside lineations, slickenfibers on exposed fault planes, secondary fractures and drag folds—reveal the kinematics along the thrusts. Useful criteria are observed only when one of the fault blocks corresponds with the marly limestones of Late Cretaceous to Paleogene age. The hanging wall movement along the thrusts within the Internal Subbetic units is mainly towards the NW, that is, subperpendicular to the fold axes in this zone (Fig. 3). For a single thrusting episode, the movement direction given by the kinematic indicators varies slightly in a given locality. For instance, the direction of the kinematic indicators in locality 1 (Figs. 3 and 8a) varies from N40°W to N70°W. A mean of N60°W for its thrusting direction is plotted in Fig. 3.

Striae or slickenfibers of a given direction often cut through or have formed over previous ones, indicating that the same fault surface was reactivated. For example, the rose diagram of locality 2 (Fig. 8b) illustrates two sets of kinematic indicators, NW- and W-directed, along the same thrust plane. In this locality, mean movement directions of N85°W and N35°W are plotted (Fig. 3). Along the studied thrusts, however, movements towards the W, and very occasionally towards the SW, NE or E, may also be observed. At the moment, the relative timing of movements along these thrusts is not clear.

Balanced cross-sections and reliable calculations of total shortening cannot be made due to: the internal deformation of the rocks, the oblique and possibly variable transport direction along some of the thrusts with respect to the cross-sectional plane, late tectonic events affecting the folds and thrusts (see below), and a lack of information at

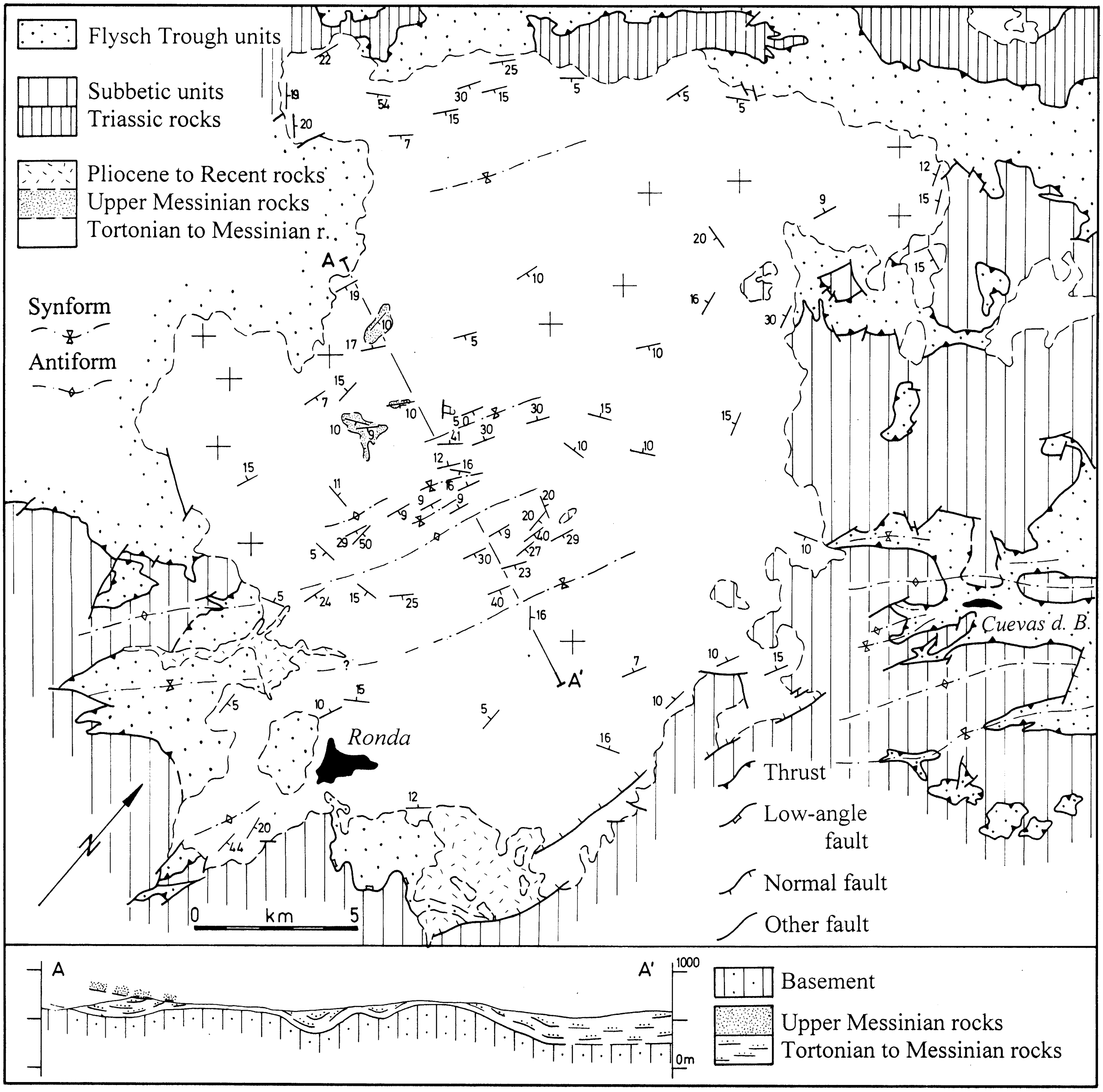

Fig. 9. Structural map and cross-section of Ronda Basin (localization in Fig. 2), modified after Cano Medina (1990), Cruz San Julián (1990), Del Olmo Sanz et al. (1990) and Moreno Serrano et al. (1990).

depth (due to lack of relief and seismic lines). Nevertheless, a rough estimation of 20% shortening due to the folds and thrusts in sequence can be calculated in cross-section A–A′, along the Triassic–Jurassic and Jurassic–Cretaceous rock boundaries (below the out-of-sequence thrusts). To be added to this minimum value would be an estimation of shortening due to: (a) internal deformation of the rocks, (b) transport along the out-of-sequence thrusts, and (c) possible duplication of the Subbetic units in order to account for the empty area between the sole thrust shown in cross-section A–A′ and the Hercynian basement situated 5 or 6 km below sea level (Fig. 2).

2.2. The Flysch Trough units

The Flysch Trough units (Fig. 3) consist mostly of deep-water turbidite deposits from Early Cretaceous to Early Miocene, whose basement is unknown, unmetamorphosed and thrusted over the Internal Subbetic (e.g. Didon, 1960; Chauve, 1968). The internal structure of the Flysch Trough

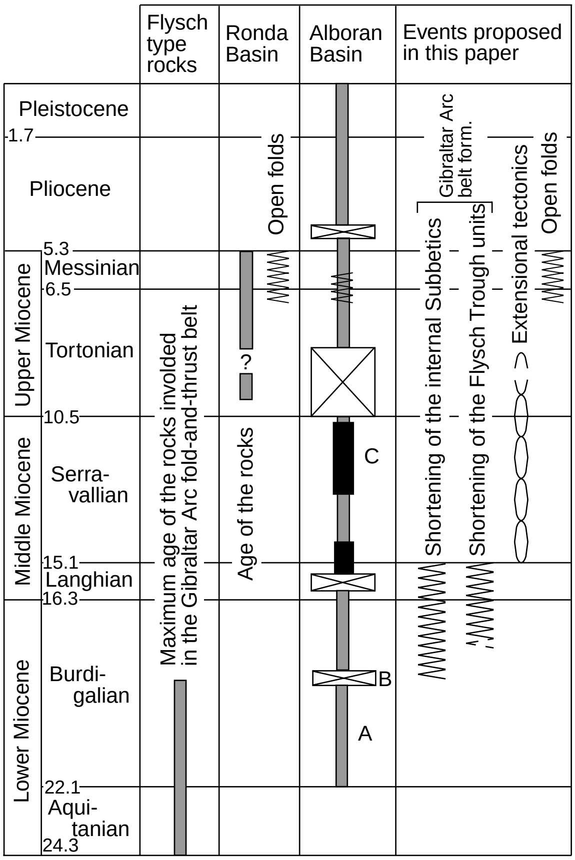

Fig. 10. Time relationships between the stratigraphic age of the affected rocks (Flysch type rocks of the Subbetic and Flysch Trough units, Ronda Basin rocks), the different structures of folding, thrusting and normal faulting observed in the study area, the rifting with deposits of marine sediment (A), major unconformities (B), main subsidence events (C) and inversion tectonics observed in the Miocene Alboran Basin (Comas et al., 1992; Rodríguez-Fernández et al., 1999) and the tectonic events proposed in this paper. Time scale according to Berggren et al. (1985).

units is not known for the study area due to poor outcrops. Yet SW of the study area, the structure of the Flysch Trough units consists of folds and thrusts in sequence, rooted in a clayey formation. They are W- to WNW-vergent in the eastern part of the outcrops, and characterized by a backthrust system in the western part (Luján et al., 1999).

Most of the boundaries between the Flysch and the Subbetic units appear as thrusts in Fig. 3. They bound units of different paleogeographical domains placed one upon the other, and there is no clear evidence of extensional reactivation along these boundaries. The thrusts are sealed by sediments of the Ronda Basin, Late Tortonian to Messinian in age (Serrano, 1979; Rodríguez-Fernández et al., 1999) (Figs. 3 and 9).

The thrusts cut the polymetric-scale folds developed in the Cretaceous to Paleogene 'red beds'; nevertheless, the Flysch units generally lie over the 'red beds', and the thrusts do not cut the 3–5 km wavelength folds marked by the massive sequence of Jurassic dolomites and limestones (Fig. 3). Kinematic indicators in a 10–50-m-wide fault zone indicate the hanging wall movement along the thrust planes, with criteria similar to those observed along the thrusts cutting the Subbetic units. Northwestward thrusting is observed consistently along these boundaries, though it is common to observe additional sets of indicators, either W- or SW-directed (Fig. 3). In a few outcrops, the cross-cutting relationships between slickenfibers on exposed shear planes indicate that the northwestward movement is the oldest. These thrusts are folded by late, kilometric-scale very open folds (see below).

The compressional or extensional character of the boundary between the Flysch Trough and the Subbetic units, labeled low-angle fault in Fig. 3, has not yet been established. The fact is that the geometric relationships do not permit the distinction of an out-of-sequence thrust from a low-angle normal fault, characteristic of the Gibraltar Arc fold-and-thrust belt and of its hinterland, respectively. Moreover, the few kinematic indicators available to date show the hanging wall movement along this boundary to be highly variable. The northwestward movement observed E and S of El Burgo (Fig. 3), for instance, fits with the thrusting episode described in Section 3.1; but the E, SE, SW and W movements observed on the same tectonic boundary are compatible with a low-angle normal faulting episode described below (Section 4). At map scale, this low-angle fault produces thinning of the Cretaceous to Paleogene rocks of the Subbetic units towards the SW, which could be consistent with a low-angle footwall ramp of a normal fault with southwestward movement. Nonetheless, in this part of the low-angle fault, not only southwestward indicators have been observed: NW and SE indicators, together with WNW and ESE ones (Fig. 3) suggest that this low-angle fault has a complex history.

3. Extensional tectonics within the Gibraltar Arc fold-and-thrust belt

A few low-to-moderate-angle faults affected the aforementioned fold-and-thrust wedge. For example, two conjugate low-to-moderate-angle normal faults bound the 'finger' of Flysch units N and S of El Burgo (Fig. 3). They are vaguely N–S directed and dip between 30 and 45°. Kinematic indicators along these normal faults, such as the slickenfibers shown in Fig. 7b, show an approximately E–W extensional direction, as illustrated in the rose diagram of Fig. 8c (locality 3). Both this diagram and Fig. 7b show the directional variation of a single set of kinematic indicators, and in this particular case, movement towards N85°W is plotted in Fig. 3. At the outcrop scale, these normal faults clearly cut the close-to-open folds that affect the Subbetic units. At map scale, the normal faults invade the Alboran Domain and produce the downthrow of the boundary between the Alpujarride and Dorsal units, from 1500 m in the Sierra Prieta and Sierra de las Nieves to 700 m near Yunquera (Figs. 3 and 4, cross-section C–C′).

This structure is well represented in the panoramic view of Fig. 5a.

A main NE–SW-trending normal fault, low-to-moderate angled, penetrates the Subbetic units SW of El Burgo (Fig. 3). Fig. 7c illustrates how minor faults associated with this normal fault cut the folds affecting the Subbetic units. The hanging wall of this normal fault moved mainly towards the SE, which produced the downthrow of the Cretaceous to Paleogene rocks with respect to the Jurassic rocks in the northeastern part of the fault (Fig. 3 and cross-section A–A′ of Fig. 4). Kinematic indicators towards the SW have also been observed along this fault. This is coherent with the footwall ramp structure viewed at map scale: the Cretaceous to Paleogene materials that lie directly upon Triassic rocks toward the SW (Fig. 3), indicating that a progressive excision of the Jurassic rocks took place along the fault. It is not clear if the SW movement is later than the SE one.

These low- to moderate-angle normal faults and their cross-cutting relationships with the folds and thrusts associated with the Gibraltar Arc fold-and-thrust belt formation evidence post-Early Burdigalian extensional tectonics subsequent to the compressional tectonics. Certain westward movement indicators observed along some thrusts of the study area (Fig. 3) would appear to be related with the reactivation of the same surfaces during this extensional event.

4. Late folding

In the northern part of Fig. 3, the thrust planes that bound the Flysch Trough and the Subbetic units exhibit NE–SW directed folds, thus producing the alignment of small sierras formed by Jurassic limestones. These undulations are very open kilometric-scale folds, as illustrated in Fig. 4 (cross-section D–D′), and they do not produce internal deformation. The folds can be followed within the Ronda Basin. They probably formed in the Late Tortonian to Late Messinian age interval (see Section 5).

The fold trend of these late folds is parallel with the axes of the previous close-to-open folds drawn by the Subbetic rocks. As the wavelength of the late folds is much larger than that of the previous ones, no outcrop-scale type 3 interferences (Ramsay, 1967) are observed. These late folds did not develop elsewhere in the studied area, and therefore their relationship with the low-to-moderate-angle normal faults described in the previous section cannot be elucidated by direct observation (see Section 6.1).

5. Ronda Basin

The Ronda Basin consists of Late Miocene sediments that seal the boundaries between the Flysch Trough units and the Subbetic units (Fig. 9). These sediments include, from bottom to top (Serrano, 1979): (a) undated sandstones and conglomerates; (b) calcarenites, marls and clays whose planctonic foraminifera associations pertain to the *Turborotalia acostaensis* subzone (Early Tortonian?) up to the *Globigerinoides elongatus*subzone (Late Messinian); and (c) algae limestones discordant upon the underlying formations (Late Messinian?).

In a recent review of the sedimentary record of the Miocene Alboran Basin, Rodríguez-Fernández et al. (1999) show that Early Tortonian sediments are missing from most of the marine sedimentary sequences recognized on land, which suggests a significant hiatus at this time. The undated continental sandstones and conglomerates of the Ronda Basin could be correlated with this hiatus in marine sediments. The assignment by Serrano (1979) to the Early Tortonian of the foraminifera association of the rocks overlying the conglomerates therefore needs some revision (Rodríguez-Fernández et al., 1999). In this paper, we assume that Early Tortonian marine sediments are lacking in the Ronda Basin.

The Tortonian to Late Messinian sediments are folded by kilometric-scale very open NNE–SSW-directed folds (Fig. 9), and the few outcrops of algae limestones dip slightly towards the E or SE. According to Serrano (1979) and Rodríguez-Fernández (1982), a progressive unconformity resulted during the deposition of the Late (?) Tortonian to Late Messinian formations, which is interpreted by these authors as due to folding or tilting during their deposition. This accords with a centrifugal pattern of paleocurrents around the main anticline situated 5 km NNW of Ronda (Fig. 9; Rodríguez-Fernández, 1982). In the southern part of the Ronda Basin these folds affect the basement, as evidenced by the undulation of the thrusts between the Subbetic and the Flysch Trough units (Fig. 9). Therefore, it is very likely that these large-scale folds are also responsible for the undulating of thrusts observed in the northern part of Fig. 3. In both areas, these late folds show similar amplitude and wavelength (compare cross-section D–D′ of Fig. 4 together with cross-section of Fig. 9).

A small outcrop of basement that consists of Triassic rocks appears 10 km N of Ronda. The overlying calcarenites include clasts of these rocks (Rodríguez-Fernández, 1982). Thus, the Triassic rocks were already at erosion level when the first sediments of the Ronda Basin were deposited. The extensional faults described earlier may have contributed to the denudation of the Triassic rocks.

6. Discussion

6.1. Time relationships between structures

The relative age of the structures described in this paper is well-known. The oldest tectonic structures are the close-to-open NW-vergent folds that affected the internal Subbetic unit. They are post-Early Burdigalian in age (see Section 2.1). They are cut by thrusts, in-sequence and out-of-sequence, in the Subbetic units and by the sole thrust that produced the superposition of two different paleogeographic domains (see Section 2.2). As a result, the Flysch Trough units, in which the youngest thrusted rocks are Early Burdigalian in age (García de Domingo et al., 1994; NN1-2 biozone, Esteras et al., 1995) lie over the Subbetic units. In the study area, the folds and thrusts produce a mainly NW–SE-directed shortening. The normal low- to moderate-angle faults described in Section 3, in turn, cut the aforementioned structures. Finally, very open kilometric-scale folds formed during the Late Tortonian and Messinian (see Sections 4 and 5). This means that both the shortening that produced the Gibraltar Arc fold-and-thrust belt and the subsequent extensional tectonics demonstrated in this paper took place between the Early Burdigalian and the Late Tortonian.

In order to refine the age of these tectonic events, some correlations can be made with structures observed in a larger area, which includes the southwestern Betics and the Alboran Basin, in the inner part of the Gibraltar Arc. The time relationships between the different structures of folding, thrusting and normal faulting observed in the study area together with: (a) the stratigraphic age of the affected rocks (Subbetic and Flysch Trough units together with Ronda Basin rocks), and (b) the rifting, main subsidence events, major unconformities and inversion tectonics observed in the Miocene Alboran Basin (Comas et al., 1992; Rodríguez-Fernández et al., 1999) are presented in Fig. 10 (according to the time scale of Berggren et al. (1985)).

In the Miocene Alboran Basin, backstripping analysis by Rodríguez-Fernández et al. (1999) reveals two periods of main subsidence, during Langhian and Serravallian times (Fig. 10). Listric, synsedimentary faults developed, such as those illustrated in the cross-section of Fig. 2 [see also Comas et al. (1992)]. These faults penetrate the Alboran Domain and the present-day Alboran Domain units are extensional tectonic units, bounded by low-angle normal fault systems with different extensional directions that developed at different times (García-Dueñas et al., 1992; Crespo-Blanc et al., 1994; Crespo-Blanc, 1995; Balanyá et al., 1997).

Kinematic indicators along low-angle extensional faults affecting the study area and the Alboran Domain are shown in Fig. 2. Transport directions vary essentially from SW to SE, though E–W extension is also observed. Nevertheless, no well defined pattern of extension can be made out. It is likely that the normal faults observed in the study area and in the Alboran Domain south of this area developed during the same age interval, although those faults situated in a more external position could have developed slightly later due to the outward migration of the extensional locus (see Section 6.2 and García-Dueñas et al., 1992). Thus, if the extensional faults developed during Langhian and Serravallian times, the folds and thrusts observed in the innermost part of the paleomargin and in the Flysch Trough units would have occurred between the Late Burdigalian and the Early Langhian (Fig. 10). During this time interval, the most internal part of the Gibraltar Arc fold-and-thrust belt developed. It must be stressed that this belt developed in the uppermost crustal levels. The maximum temperature reached by the rocks constituting the paleomargin and Flysch units never surpassed 80°C, as indicated by the immaturity of Late Cretaceous organic facies of the internal Betic units (Reicherter et al., 1994) and the high proportion of smectite (over 50% of the clayey fraction) of the clayey formations belonging to the Flysch type sequence (López-Galindo and Martín-Algarra, 1992). Moreover, basement slices have not been observed in either the Subbetic units or the Flysch Trough units. The Gibraltar Arc fold-and-thrust belt is characterized, then, by thin-skinned tectonics.

The last main event recognized in the study area is related with the kilometric-scale very open folds, NNE–SSW-directed, Late (?) Tortonian to Late Messinian in age (see Section 5). These folds are the result of the contractive reorganization observed by Comas et al. (1992) and Rodríguez-Fernández et al. (1999) in the Alboran Basin from the end of Tortonian (Fig. 10). In fact, from Late Tortonian to Pliocene, the Alboran region underwent N–S to NW–SE compression: a sudden uplift took place and the extensional fault systems of the Alboran Domain were folded and faulted (e.g. García-Dueñas et al., 1992; Sánchez-Gómez, 1997; Rodríguez-Fernández et al., 1999; Luján et al., 2000). The study area provides no direct observation of the folding of the normal faults that affect the Flysch and Subbetic units. However, the current boundary between the internal Subbetic units and the Flysch Trough units SE of the study area is a normal low- to moderate-angle fault, later affected by NE–SW to NNE–SSW-trending very open folds (western border of Fig. 2; Luján et al., 2000). The very open folds that affected the low-angle faults belonging to the Alboran Domain trend roughly NE–SW (Sánchez-Gómez, 1997), that is, broadly parallel with the late folds observed in the Internal Subbetic and Flysch Trough units (Fig. 2).

6.2. Tectonic evolution of the Gibraltar Arc fold-and-thrust belt within the western Mediterranean framework

The superposition of Miocene structures observed in the Internal Subbetic units, as demonstrated in this paper, reveals an alternating of compressional and extensional events similar to the tectonic evolution observed in its hanging wall, the Flysch Trough units (Luján et al., 2000). The proposed evolution differs substantially from interpretations of the structure of the Subbetic units as a result of strike-slip tectonics (e.g. Leblanc and Olivier, 1984; Sanz de Galdeano, 1990), or of a single compressional event (e.g. Bourgois, 1978; Blankenship, 1992; Platt et al., 1995; Kirker and Platt, 1998). In particular, Kirker and Platt (1998) assert that transport directions throughout the Gibraltar Arc are consistently WNW-directed, regardless of the structural trend. Without discriminating the different events recognized in the present paper, these authors

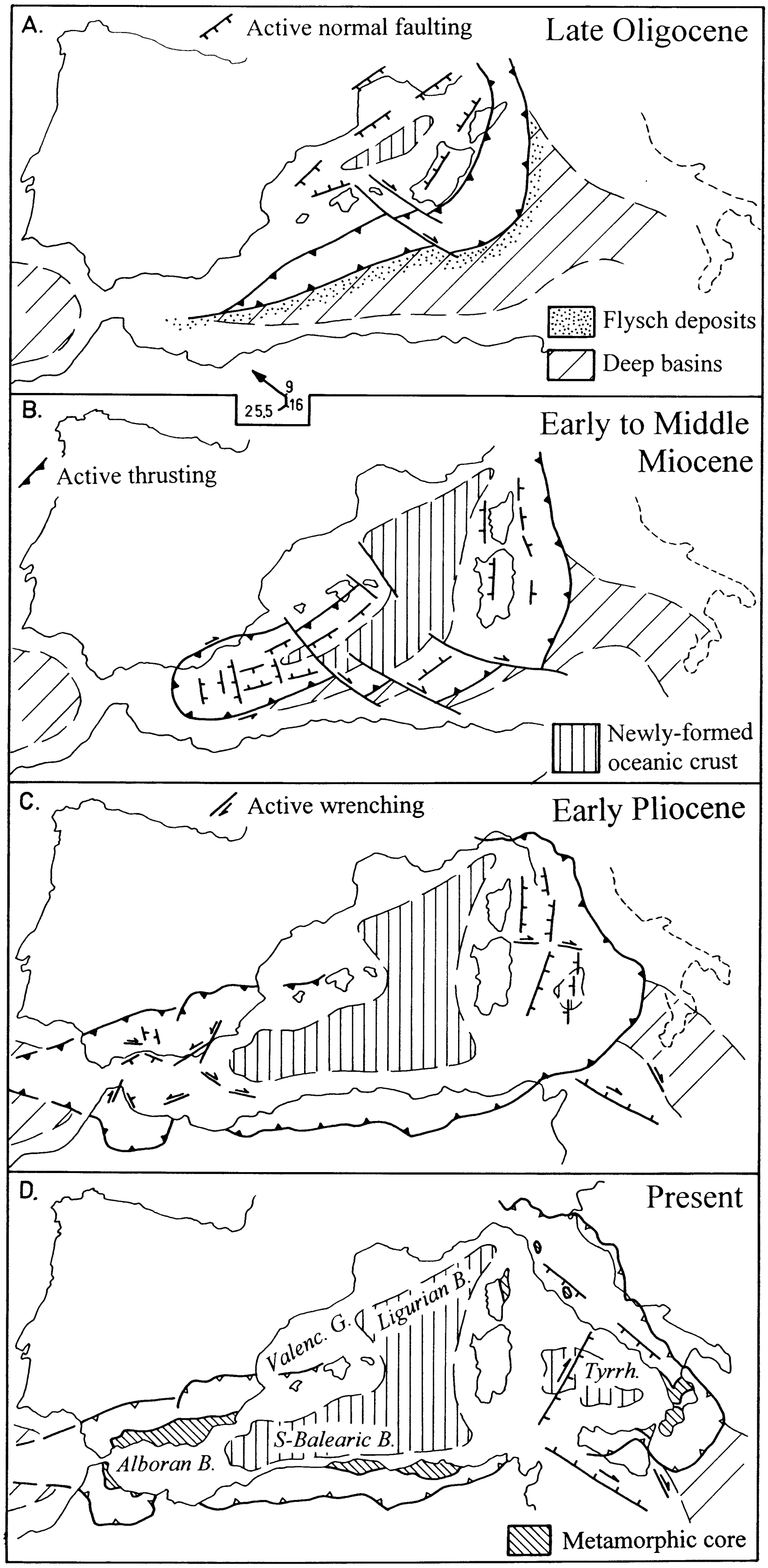

Fig. 11. Schematic paleotectonic evolution of the western Mediterranean area, based on Dewey et al. (1989) and Jolivet et al. (1998); Late Oligocene-to-Present Africa–Europe relative motion vector (arrow on sketch A) according to Mazzoli and Helman (1994). Valenc.G.: Valencia Gulf; Tyrrh.: Tyrrhenian Sea.

average out indicators measured on single surfaces, even when local spread departs by up to 50° from the mean direction (see Fig. 10 in op. cit.). The local variations of up to 15° in azimuth from the mean direction of the transport movement along thrusts, together with the likely reactivation of the same surfaces during the late extensional event observed in the study area (see Section 3), call for caution in interpreting the indicators along the thrusts. At the same time, the varied pattern of contractional transport direction in the Gibraltar Arc fold-and-thrust belt as it swings around the Arc must be taken into account: transport is NW-directed in the study area, WNW-directed in the northernmost outcrops of Flysch Trough units (Fig. 3; Luján et al., 1999), W-directed near the Strait of Gibraltar (Balanyá et al., 1995; Crespo-Blanc et al., 2001), and SW-directed in the Rif (e.g. Morley, 1987). Rotations about vertical axis determined in paleomagnetic studies (e.g. Platzmann et al., 1993; Villalaín et al., 1994; Platt et al., 1995; Kirker and McClelland, 1996) may have modified the original pattern of transport direction, as they took place after the main shortening event (Crespo-Blanc et al., 2001); however, the present-day displacement vector field suggests that radial shortening occurred during the formation of the Gibraltar Arc fold-and-thrust belt (see Fig. 11d).

The tectonic evolution of the Flysch Trough and Subbetic units is framed within the evolution of the western Mediterranean in the sketches of Fig. 11. This figure is based on the paleotectonic reconstructions of Dewey et al. (1989) and Jolivet et al. (1998), and the convergent African–Eurasian plate motion bounding the system described by Mazzoli and Helman (1994). The position of the active thrust fronts and the areas of active normal or strike-slip faulting is shown for Late Oligocene, Early to Middle Miocene and Early Pliocene times. The present-day position of the internal and external zones, along with the pattern of transport direction around the western Mediterranean, is drawn in Fig. 11d.

During Late Oligocene (Fig. 11a), active rifting in the Liguro–Provençal and Valencia Gulf basins took place (from Late Eocene to Middle Aquitanian; Rehault et al., 1985; Dewey et al., 1989). The opening of these basins is classically considered the consequence of the W-dipping retreating subduction of the Adrian–Ionian lithosphere (present-day coordinates; e.g. Malinverno and Ryan, 1986). Associated with this subduction, which began during the Eocene, a large accretionary complex developed. In Late Oligocene times, the last effects of compression are felt in Alpine Corsica, where the W-vergent thrusts are considered as backthrusts (Jolivet et al., 1998). This bivergent accretionary complex, in which high-pressure-type metamorphism affects what will constitute most of the outcropping Alboran Domain (e.g. Azañón and Crespo-Blanc, 2000 and references therein) can be followed towards the Gibraltar area (Dewey et al., 1989), where tomographic data show that the subduction plane dipped towards the NW (present-day coordinates, Blanco and Spakman, 1990). Sedimentation of flysch-type deposits occurred within the Apennine and Sicilian sectors of the subduction trench (Dercourt et al., 1986), while in the North African sector, quartzo–arenitic flysches (Numidian flysch of Aquitanian age; e.g. Didon, 1969) could be related to the formation of a peripheral bulge (e.g. Dewey et al., 1989).

In the western Mediterranean, compressional tectonics went hand in hand with extension during Early to Middle Miocene times (Fig. 11b). According to Dewey et al. (1989) and Jolivet et al. (1998), among others, oceanic crust was created in the Liguro–Provençal basin while thrust faults were reactivated as extensional shear zones in Alpine Corsica; the Corso–Sardinian block rotated counter-clockwise, probably as the Burdigalian and early rifting of the Tyrrhenian Sea took place simultaneously with the eastward migration of the thrust front in the Apennines. In a similar way, the Alboran Domain was structured as a post-metamorphic nappe-stack during the Aquitanian (Azañón and Crespo-Blanc, 2000) and was thrusted around the Gibraltar Arc, over the Flysch Trough and the South Iberian and Maghrebian paleomargins. Meanwhile, in the inner part of the Arc, the same domain is affected by rifting: low-angle normal fault systems developed from Burdigalian to Serravallian, and the Alboran Basin formed (Comas et al., 1992, 1999; García-Dueñas et al., 1992). Westward migration of the extensional locus can be inferred, as the South-Balearic Basin, formed from Late Oligocene to Late Langhian (Vergés and Sàbat, 1999), developed slightly earlier than the Alboran Basin. Rifting took place within the Arc, with the formation of new oceanic crust in the South-Balearic Basin (Rehault et al., 1985), while radial thrusting and shortening occurred in the peripheral belt, the compressive front migrating from the Alboran Domain to its footwall. The Gibraltar Thrust, at the front of the Alboran Domain, was blocked during the Middle Miocene (Balanyá and García-Dueñas, 1988), and shortening was accomplished by progressively deeper and more external thrusts located in the Flysch Trough and paleomargin units. In the South Iberian segment, this shortening is dated as Late Burdigalian and/or Langhian (Fig. 10), which is consistent with the Middle and Late Miocene shortening event observed in the Maghrebian segment (Morley, 1992). The outward migration of the compressional front is also documented by the migration of the olistostrome deposits in the foreland Guadalquivir Basin from E to W during Middle and Late Miocene (e.g. Sanz de Galdeano and Vera, 1992). The migration of the compressional front, coupled with that of the extensional locus situated in the inner part of the Gibraltar Arc, produced the extensional reactivation of the Gibraltar Thrust (García-Dueñas et al., 1992) and the development of the extensional structures observed in the fold-and-thrust belt situated in its footwall (this paper). The switch between compression and extension probably took place within a 2 or 3 Ma interval (Fig. 10). Analogous switches and time intervals have been described in other Mediterranean backarc environments: for example, the formation of the

Tyrrhenian Sea during the eastward migration of the Apennine thrust front (e.g. Jolivet et al., 1998).

In the Late Miocene and Early Pliocene (Fig. 11c), rifting is still active in the Tyrrhenian Sea, and oceanic crust starts to form in its southern part. Extension progressively shifts eastward, following the Apennine thrust front, toward the Adrian–Ionian foreland (Jolivet et al., 1998). In the Gibraltar Arc area, a contractive reorganization took place from the Tortonian onwards: tectonic inversions of previous normal faults were followed by wrench tectonics in the Alboran Sea (García-Dueñas et al., 1992; Comas et al., 1999), while onshore, kilometric-scale open folds produced the final emersion of the prism and the part of its hinterland that constitutes the Alboran Domain.

It must be stressed that the intention of the paleotectonic reconstructions of the western Mediterranean outlined in Fig. 11 is merely qualitative. The Miocene tectonic evolution of the Internal Subbetic units—interpreted in this paper in terms of compressional tectonics giving rise to the formation of a fold-and-thrust belt, followed by extensional tectonics—provides an important constraint for future models of the Betic–Rif orogen. The proposed evolution should now be tested in other areas of the external Betics, not only qualitatively but also quantitatively, with balanced cross-sections and large-scale displacement estimations.

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