Geologie en Mijnbouw **77:** 39–61, 1999.

Oligocene to Middle Miocene basin development in the Vélez Rubio Corridor – Espuña (Internal-External Zone Boundary; Eastern Betic Cordilleras, SE Spain)

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Received 15 July 1997; accepted in revised form 24 June 1998

Key words: stratigraphy, detritus analysis, tectono-sedimentary history

Abstract

The Vélez Rubio Corridor and the area northwest of the Sierra Espuña are located on the Internal-External Zone Boundary. The External Zone is represented by the Southern Subbetic, the most basinward part of the former passive margin of Iberia, the Internal Zone by its unmetamorphosed highest unit, the Malaguide Complex, tectonically underlain by the metamorphosed Alpujarride Complex. During the Oligocene and Aquitanian, the Southern Subbetic was the locus of slope deposition with northeastern provenance of detritus. In the Malaguides of the Espuña, the detritus of Lower Oligocene transgressive conglomerates and Middle Oligocene fan deltas indicates Sardinian proximity. The Upper Oligocene to lower Aquitanian Ciudad Granada and Pliego formations of the Malaguide Complex, carrying exclusively Malaguide detritus, were deposited in grabens within the Malaguide realm during an extensional rifting phase. The Malaguides, still far removed from the Subbetic, underwent major thrusting during the Aquitanian. Of the sedimentary units found between the Internal and External Zones, the oldest unit (the allochthonous Aquitanian Solana formation) was deposited in submarine fans outside the Subbetic or Malaguide realms proper, but in close connection with the latter. The Internal Zone collided with the External Zone in the early Burdigalian with concomitant disruption of the Southern Subbetic slope. On the suture, a deep basin was formed and filled in by the Burdigalian Espejos formation carrying detritus from the Subbetic and from the Malaguide and Alpujarride Complexes. In the late Burdigalian, the Subbetic was thrust southward over the Espejos formation, thus double-sealing the collisional contact. During the latest Burdigalian to Langhian, new basins were formed along the Internal-External Zone Boundary and within the Southern Subbetic. The onset of strike-slip faulting caused shoaling and uplift of these basins. Onset of a new pattern of strike-slip faulting induced the formation of a new suite of basins during the Tortonian, e.g. the Lorca Basin.

Introduction

The Betic Cordilleras of southern Spain can be subdivided into an External Zone, representing the former passive margin of Iberia, and an Internal Zone, consisting of a stack of allochthonous complexes (Figure 1). The External Zone comprises a northern, continentward Prebetic Zone and a southern, basinward Subbetic Zone. The structurally highest unit of the Internal Zone is formed by the Malaguide Complex, comprising unmetamorphosed rocks of Palaeozoic, Mesozoic and Tertiary age, which tectonically overlie metamorphic rocks of the Alpujarride and Nevado-Filabride Complexes. The Vélez Rubio Corridor forms at present the boundary between the External and Internal Zones. It is separated by the Tortonian-Pliocene Lorca Basin from the External-Internal Zone Boundary to the northwest of the Espuña ('zone-limite' of Paquet 1969; Figure 1). The analysis of the Tertiary sediments in the Vélez Rubio Corridor and the area northwest of the Espuña is important for the timing

† Thom Roep died suddenly September 15, 1997. His stimulating discussions were dearly missed during the revision of this paper.

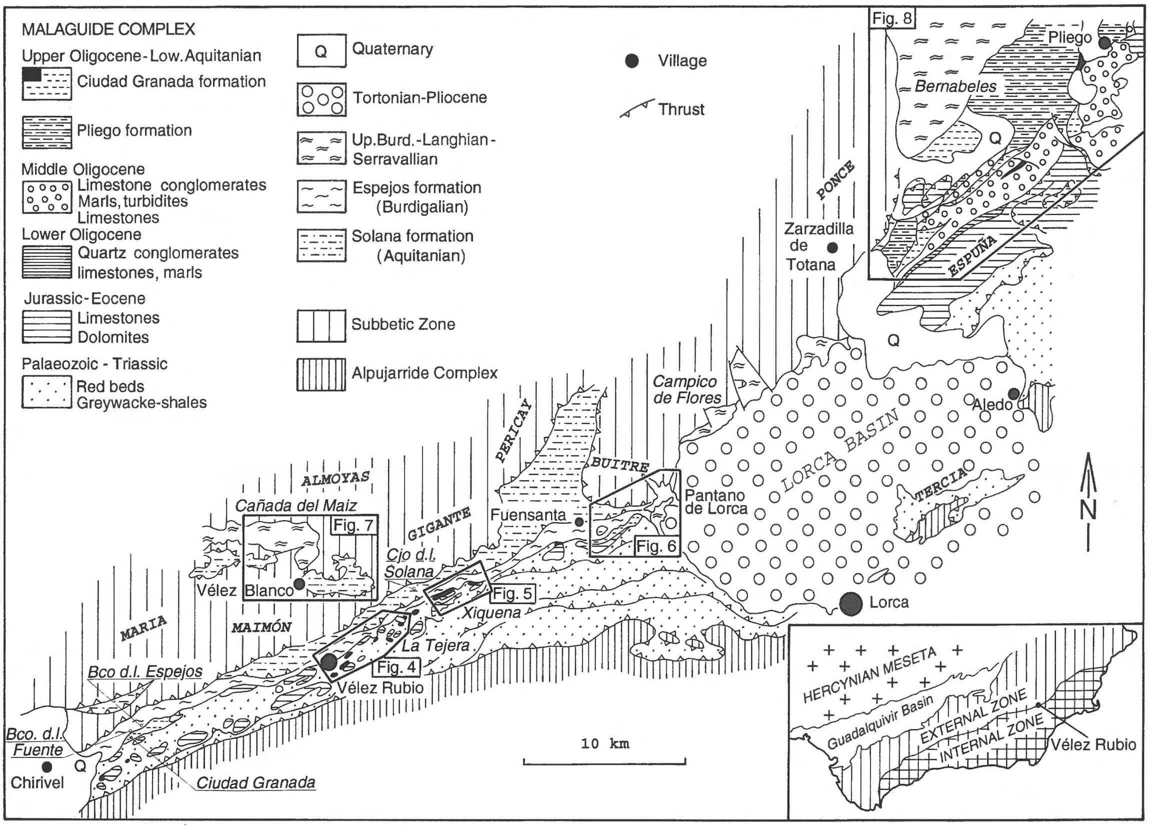

Figure 1. Simplified geological map of the Vélez Rubio Corridor and Espuña area, showing the location of the type localities of the Oligo-Miocene formations discussed in the text and the locations of Figures 4 to 8.

of the suturing of the Internal and External Zones, or in other words, of the final emplacement of the Internal Zone Complexes. Dutch workers (including the authors) started to investigate the Corridor in 1959, which resulted in a series of separate papers and two PhD theses (referenced below). At the same time, French geologists worked in the Espuña area (Paquet 1966, 1969). The one but last publication on the Corridor by the Dutch group was by Hermes (1984), a rather confusing paper that upset our previously published results. Recently, British workers (Lonergan et al. 1994) also studied the Internal-External Zone Boundary in the Corridor and the Espuña area. Their conclusions are at variance with our results, due to what we consider to be stratigraphic inaccuracies. The Tertiary deposits of the Espuña area were also analysed by Martín-Martín in his 1996 thesis. His conclusions are largely in agreement with our ideas, though some of his views merit discussion.

The purpose of the present paper is twofold: first, to place the results obtained in the 1960s and early 1970s in a more state-of-the-art plate-tectonic and sedimentological context. It should be borne in mind that in those decades the Betic Cordilleras were thought to be the result of one compressive north-south directed Tertiary movement. The Oligo-Miocene West Mediterranean extensional phase and the allochthonous Alboran or South Sardinian microplate had yet to be discovered; slope fan models were just being developed. Secondly, to discuss the work of Hermes (1984), Lonergan et al. (1994) and Martín-Martín (1996). The stratigraphic part of the paper not only summarizes data published before 1980, but also includes new or hitherto unpublished information. On the basis of the new and re-evaluated old data, the tectono-sedimentary evolution of the Vélez Rubio Corridor and Espuña area will be discussed.

Though conventionally the Oligocene is subdivided into the Lower Oligocene or Rupelian, and the Upper Oligocene or Chattian, for convenience sake we will use in the descriptive part a threefold division: Lower Oligocene (P18-P19-P20 = lower Rupelian), Middle Oligocene (P21 = upper Rupelian to lower Chattian), and Upper Oligocene (P22 = upper Chattian), in order to avoid repetitive use of such akward terms as 'lower Lower', 'upper Lower', 'upper Lower to lower Upper'. In this paper we use the time scale of Berggren et al. (1995). The planktonic foraminiferal zonation is according to the scheme of Blow (1969) as emended by Berggren et al. (1995).

Oligo-Miocene stratigraphy of the Southern Subbetic, the Malaguide Complex and intermediate deposits

Oligocene to Lower Miocene of the Southern Subbetic

The Subbetic deposits are often in a strongly tectonized, even chaotic, state. More often than not the sections are disturbed by bedding-parallel thrust planes. The following description is based on data from scattered outcrops because the relationships had to be examined at different localities (Figures 1, 2).

Oligocene

Between the Sierra Ponce in the east and the Sierra de Maria in the west (north of the Sierra de Maria-Maimón-Gigante range), the dominant Oligocene facies is formed by light-coloured, fine-grained bioclastic and planktonic foraminiferal packstones, the Calcarenite-Calcipelite formation of Hermes (1978; Figure 2). The beds, either featureless or parallel laminated, are often broken up into a series of flat balls, which are gravity-induced pull-aparts. The fine-grained packstones contain intercalations of coarse-grained, larger-foraminiferal turbidites and mass flows; channelling and slumping are common features. The thickness of the formation varies laterally, from < 100 to > 300 m. In the east (Buitre, Ponce), where the formation reaches its greatest thickness, the base is formed by up to 10-m-thick grain-supported mass flows containing shallow-water fossils: coral fragments, coralline red algae, larger Foraminifera, bivalves, gastropods, etc. In several localities an overall thinning-up of beds has been observed. Almost everywhere, the calcarenites are underlain by Lower to Middle Eocene greyish-green marls, but in the Sierra Ponce they cut erosively into Upper Cretaceous pelagic limestones. All features described suggest that the calcarenites have been deposited in a submarine slope-fan environment.

The age of these slope-fan deposits is rather controversial. Because the turbidites and mass flows in the lower part of the formation contain exclusively Eocene larger Foraminifera (*Nummulites, Discocyclina*) and because they rest upon upper Middle Eocene marls, Hermes (1978) concluded to a late Middle Eocene age for the base of the formation. However, marls intercalated near the base in the Sierra de Buitre yielded a planktonic foraminiferal association indicating an Early Oligocene P19 age (presence of *Globigerina ampliapertura*, absence of *Pseudohastigerina* sp.). Turbidites and mass flows up-section contain a mixture of Eocene and Oligocene larger Foraminifera (*Nummulites, Discocyclina, Lepidocyclina*; near the top also *Miogypsinoides*) which suggests a Middle to Late Oligocene age for the remainder of the formation.

On the southern slopes of the Sierra de Maria-Maimón-Gigante range, Oligocene deposits are represented by greenish-grey marls with intercalations of planktonic foraminiferal packstones and larger-foraminiferal and algal (rhodolithic) turbidites. They form the upper part of the Taibena formation of Geel (1973; Figure 2). Several planktonic foraminiferal biozones occur within the marls: P19 (presence of *Globigerina ampliapertura*, absence of *Pseudohastigerina* sp.), P21 (co-occurrence of *Globigerina angulisuturalis* and *Paragloborotalia opima opima*) and P22 (co-occurrence of *Globorotalia opima nana* and *Neogloboquadrina mayeri* gr.; absence of *Paragt. opima opima* and *Globorotalia kugleri*). In marls of Early Oligocene P19 age, a 15-m-thick massive larger-foraminiferal and algal mass flow is intercalated, containing exclusively reworked Eocene larger Foraminifera. The coarse turbidites up-section yielded both reworked Eocene and Middle to Late Oligocene larger Foraminifera (*Eulepidina, Miogypsinoides*). The greenish-grey sediments grade downward into yellowish marls and planktonic foraminiferal packstones of Late Eocene age (P17, *Turborotalia cerroazulensis* Zone). Because of the strong imbrication here, it is impossible to give a reliable figure for the thickness of the Oligocene deposits.

We interpret the deposits south of the Sierra de Maria-Maimón-Gigante range as the downslope, basinward counterpart of the slope-fan deposits of the Calcarenite-Calcipelite formation. This interpretation is confirmed by the presence of an intermediate facies to the northeast of Vélez Blanco, indicating a marl content increasing to the southwest.

Lower Miocene

In the entire Southern Subbetic, brown, coarse *Lepidocyclina*–algal packstones are found: the Lep-

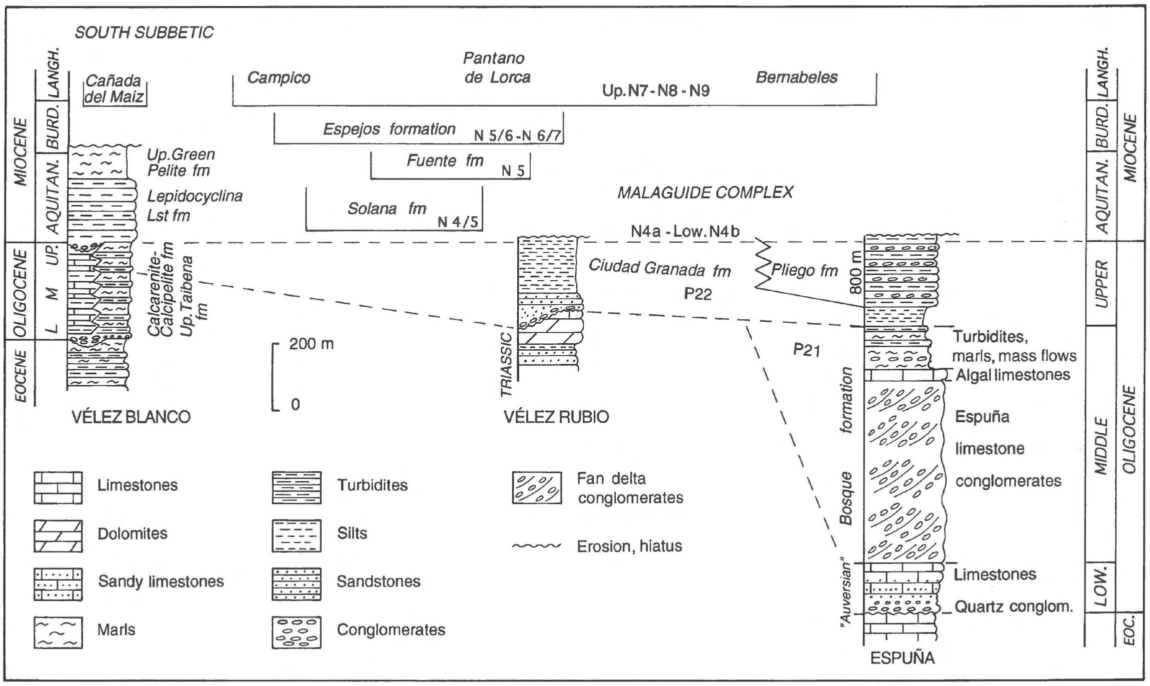

Figure 2. Simplified columnar sections showing the stratigraphic relations and correlation of the Oligo-Miocene formations discussed in the text. P21, P22 and N4 to N9 refer to the planktonic foraminiferal biozones as used by Berggren et al. (1995). The half-boxes indicate the palaeogeographic extent of the Miocene deposits and their relationships with the Subbetic Zone and the Malaguide Complex.

idocyclina Limestone formation of Hermes (1978; Figure 2). They are medium to very thick-bedded and show often the features of turbidites, for instance, normal grading, parallel and current lamination, and sole marks. Marly intercalations are subordinate. The beds are frequently discontinuous laterally. The Lepidocyclina Limestone unit may reach a thickness of 200 m (locally more). The lower boundary of the unit is generally sharp, marked by a sudden increase in grainsize and in thickness of the beds compared with the underlying Oligocene deposits. In the Almoyas, north of Vélez Blanco, the unit is composed of three superimposed channel fills, each c. 100 m thick, cutting erosively into older strata, down to Cretaceous pelagic limestones. The channels are filled in by *Lepidocyclina*-bearing turbidites, inversely graded beds, mud-supported debris flows, slumps, lime mudstones and marls. Current directions are towards the southwest. The components of the debris flows are derived from the Subbetic realm, e.g. Jurassic oolitic limestones, Tintinnid limestones, middle Cretaceous green marls, and Eocene larger-foraminiferal packstones. The features described suggest that the Lepidocyclina Limestone unit was deposited in a submarine middle to inner slope-fan environment.

Intercalated marls invariably contain planktonic foraminiferal faunas of the *Globorotalia kugleri* Zone (N4), corresponding to the early to middle Aquitanian. This age is corroborated by the presence of *Eulepidina, Nephrolepidina, Miogypsinoides* and *Miogypsina*.

Above the Lepidocyclina Limestones follow greenish-grey clay-rich marls: the Upper Green Pelite formation of Hermes (1978; Figure 2). Intercalations of silexites, thin-bedded planktonic foraminiferal turbidites and sandy turbidites are common. Locally, west and south of the Gigante and Buitre, tuffites are intercalated (De Clercq et al. 1975). On account of the presence of *Catapsydrax dissimilis, Cat. unicavus, Cat. stainforthi, Globigerinoides trilobus, Gd. altiaperturus, Gd. altispirus, Globoquadrina dehiscens, Globigerina woodi, G. angustiumbilicata* and *Globorotalia obesa*, the planktonic foraminiferal fauna can be assigned to the *Catapsydrax dissimilis* and/or *Cat. stainforthi* Zones (N5 and/or 6). Given the absence of *Globigerinatella insueta* and *Globorotalia praescitula*, a restricted assignment to the *Cat. dissimilis* Zone (N5) is more plausible. This corresponds to a latest Aquitanian to early Burdigalian age.

The boundary between the *Lepidocyclina* limestones and green marls is gradual. It probably rep-

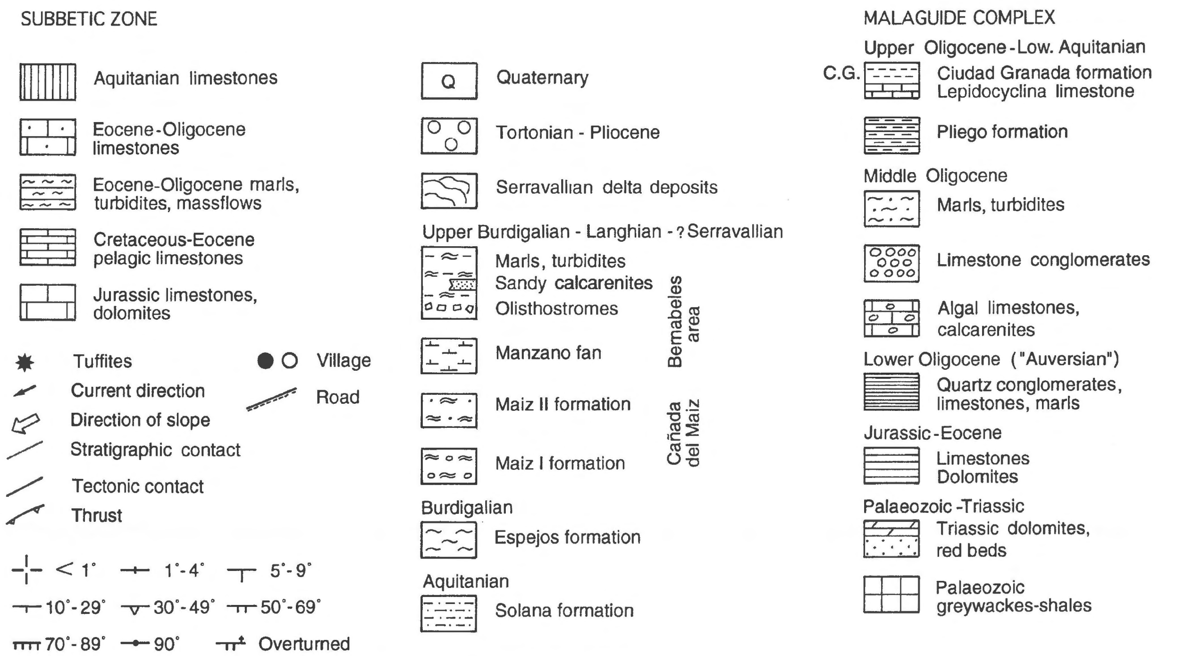

Figure 3. Legend to the geological maps of Figures 4A, 5A, 6A, 7A and 8A.

resents the depositional transition between an active slope-fan and an abandoned fan in a basin-plain environment. The common occurrence of silexites (diatomites) indicates sedimentation in a narrow basin, gulf or bay with limited connections to the open sea. Though silexites do occur in older, Eocene, pelites (Hermes & Smit 1976), they are more frequent in the Upper Green Pelite formation. This observation, together with the change in facies and the fact that the Upper Green Pelite formation is locally not deposited, suggests that the former slope of the Subbetic realm became disrupted, though in part of the Southern Subbetic deep-marine sedimentation continued. In this connection it is important to note that De Smet (1984) reported debris flows, with Subbetic components, and turbidites, unconformably overlying older Subbetic rocks to the north of the study area.

The green marls represent the youngest deposits of the Southern Subbetic proper. Though uppermost Burdigalian to Langhian deposits do occur in the area, they cannot be regarded as being Subbetic since they are connected with another suite of basins (see below).

Oligocene to Aquitanian of the Malaguide Complex

In the Malaguide realm, Upper Eocene to Middle Oligocene rocks are nearly always missing. Sedimentation started anew during the Late Oligocene (P22). The Espuña forms an exception, because its section comprises Lower to Middle Oligocene (P20–P21) sediments.

Lower to Middle Oligocene of the Sierra Espuña (Figures 1, 2, 8A, B)

Lower Oligocene sediments, the As formation of Martín-Martín (1996), erroneously attributed to the Auversian by Paquet (1966), overlie, without a noticeable angular unconformity at outcrop scale, Middle Eocene deeper open-shelf, inner-platform or continental deposits or Cretaceous pelagic rocks. They represent a transgressive shallow-shelf sequence, with mega-crossbedded supermature conglomerates (mainly quartz pebbles) at the base, a diminishing-upward amount of noncalcareous detritus, and carbonate-platform deposits at the top. They yielded a mixed, largely reworked, fauna of larger and planktonic Foraminifera of which the youngest elements indicate an Early Oligocene P20 age (Geel 1996). Both detritus and environment appear incompatible with the proximity of an active thrust front of imbricated carbonate units as envisaged by Lonergan (1993). This disagrees with her idea of early thrusting in the Malaguide realm (see also Geel 1996; cf. Martín-Martín 1996).

The Lower Oligocene is overlain by the Bosque formation (Lonergan 1991), a sequence of limestone conglomerates, several hundreds of meters in thickness, representing a backstepping fan-delta complex at the margin of a platform, situated to the northeast (Geel 1996). The presence of *Eulepidina* and *Nephrolepidina* in the associated turbidites and mass flows indicates a Middle Oligocene P21 age for the conglomerates. Martín-Martín (1996) reported the presence of NP 23 to NP 25 nannofossil associations, which supports a Middle Oligocene age. During this time, the Malaguide Complex formed part of the South Sardinian block. This block was juxtaposed to Sardinia, the North Sardinian block, and situated several hundreds of kilometres east of its present position (see 'Discussion and conclusions'). Analysis of the limestone clasts suggests that the source area did not belong to the Malaguide domain but to the North Sardinian block, given the majority of fragments of Upper Jurassic sheltered inner-platform carbonates (*Clypeina* limestones) not known from the Malaguides proper (Figure 7B in Geel 1996). This observation implies, that the majority of the clasts have not been supplied by Malaguide thrust slices in the southeast and invalidates one of the main arguments for the Oligocene thrusting and nappe emplacement suggested by Paquet (1966), Lonergan (1993) and Martín-Martín (1996).

Upper Oligocene to Aquitanian of the Vélez Rubio Corridor and the Espuña area

In the Malaguide Upper Oligocene to Aquitanian, two lithostratigraphic units can be distinguished which are partly coeval but differ in environmental conditions: the Ciudad Granada and Pliego formations.

Ciudad Granada formation. This unit was first recognized to be an important entity by Roep & MacGillavry (1962) who described it as 'Oligocene of the Salud subunit'. Soediono (in MacGillavry et al. 1963/64) introduced the name Ciudad Granada formation after a hamlet near Chirivel. Soediono (1971) restricted this name to the large outcrop north of the Murcia-Granada highway, whereas he assembled small outcrops elsewhere in his study area in the Frac and Perro Malo formations on account of small lithological differences. The three formations together form his Ciudad Granada group. In this paper, we include the Perro Malo formation and members A and B of the Frac formation of Soediono (1971) into the Ciudad Granada formation because the lithological differences are small. The top part of Soediono's Frac formation, member C, however, cannot be considered as belonging to the Ciudad Granada formation (see below, under Fuente formation).

The Ciudad Granada formation occurs in scattered small outcrops located in the Vélez Rubio Corridor from Chirivel to the southeast of Fuensanta, and on the northwest side of the Espuña (Figures 1, 2). In the Corridor, the base of the formation is formed by unsorted conglomerates with both angular and very well-rounded predominantly dolomitic pebbles, followed by light-yellow or pinkish to brown, indistinctly bedded calcareous sandstones with conglomeratic lenses and well-rounded 'frosted' quartz granules. The thickness of the sandstone is up to 65 m. Large flat *Operculina, Lepidocyclina*, echinid spines, bivalves and coral fragments are frequently present. Up-section, the sandstones grade rapidly into reddish-brown to yellowish-brown pelites. These 'red' pelites comprise diagnostic thin (c. 3 cm) layers of purple or ochreous siltstone. These layers can be used to unravel the complicated tectonics of the pelites, which, because of the lack of distinct bedding, would otherwise remain obscure. A cobble conglomerate, 'black conglomerate', is intercalated at various places. The thickness of the red pelites is difficult to estimate due to intense tectonic deformation, but is at least c. 200 m (Soediono 1971).

In the Vélez Rubio Corridor, from Chirivel to Fuensanta, the basal conglomerates and *Operculina* sandstones rest upon a paleorelief of Triassic sandstones and dolomites of the Malaguide Complex without a noticeable angular unconformity. Southeast of Fuensanta, the basal beds either overlie Triassic dolomites and sandstones, or limestones of Jurassic or Middle Eocene age, again without angular unconformity. Locally, *Lepidocyclina* and/or coral limestones are found at the base (Geel 1967, 1973). In the Espuña, where the basal conglomerates and *Operculina* sandstones are missing, the red pelites are found directly upon Middle Oligocene yellow marls and turbidites (Geel 1996).

The carbonate clasts of the basal conglomerates and *Operculina* sandstone are derived from subjacent Malaguide Triassic or Jurassic rocks. The non-calcareous detritus has been reworked from Triassic sandstones (Roep & MacGillavry 1962; Soediono 1971; Geel 1973). The black conglomerates intercalated in the red pelites contain predominantly clasts derived from the Malaguide Palaeozoic. In one locality near Chirivel, they contain in addition light-coloured gneissic rocks with radiometric ages of 323 ± 60, 332 ± 60 and 350 ± 60 Ma (Soediono 1971). Car-

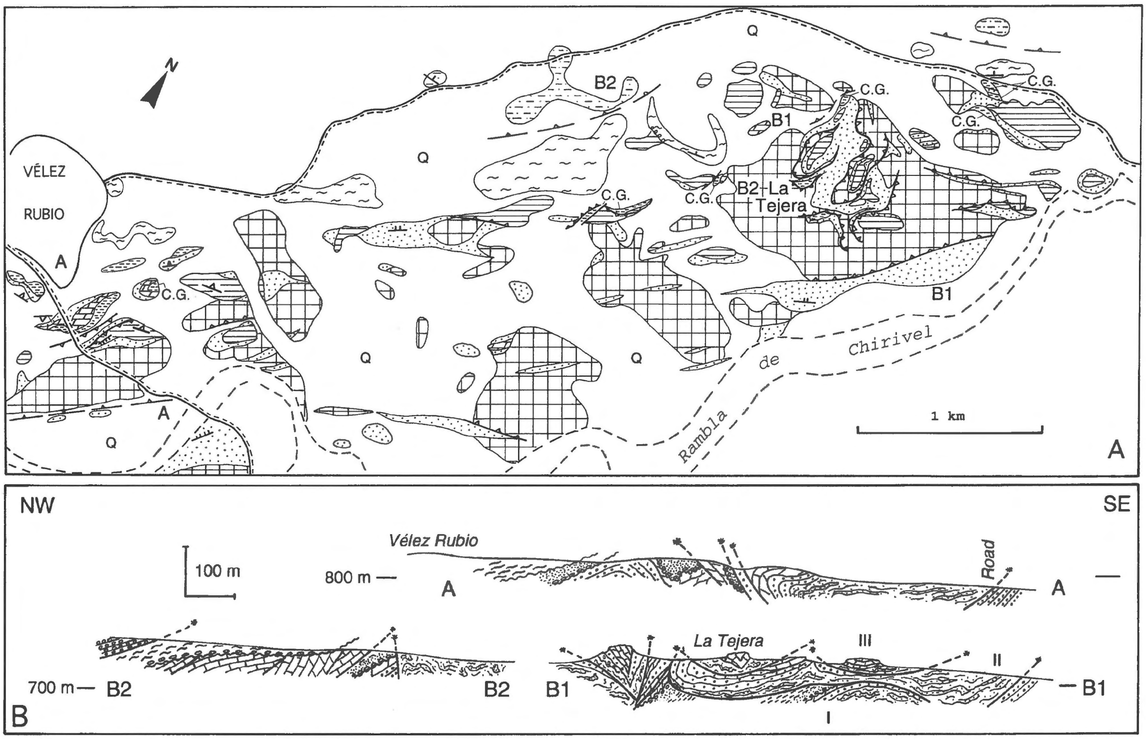

Figure 4. Vélez Rubio-Tejera area. A) Geological map showing locations of cross sections AA, B1B1 and B2B2, and scattered outcrops of the Ciudad Granada formation (C.G.) (for location, see Figure 1, for legend, Figure 3). B) Cross sections (for legend, see Figure 5B). Cross section AA: the Ciudad Granada fm, overlying with an erosional unconformity Triassic dolomites and sandstones, is folded and thrust together with its substrate. Cross sections B1B1 and B2B2 show three superimposed imbricated units (I, II and III), the lowermost one (I) comprising Ciudad Granada deposits overlying Triassic dolomites. The pile of imbricated units is dissected by younger faults. The Espejos fm overlies transgressively the imbricated and faulted Malaguide units, including the Ciudad Granada fm. The Burdigalian Espejos fm is tectonically overlain by the Aquitanian Solana fm.

boniferous gneisses are unknown in the Malaguides of Spain but do occur in the basement of their Kabylian equivalents in northern Africa (e.g. Bourrouilh et al. 1979; Boissière & Peucat 1985; Monié et al. 1988).

The basal part of the Ciudad Granada formation was deposited in very shallow, normal-marine water as can be inferred from the occurrence of larger Foraminifera, bivalves, coral colonies and *Balanus*. Up-section, the larger benthonic fauna diminishes and finally an association rich in planktonic Foraminifera and well-preserved, highly diversified smaller benthonic Foraminifera (e.g. *Almaena*) is found. The change from sandstones to pelites records a rapid deepening to more than 200 m in waterdepth (Soediono 1971; Gelati & Steininger 1984). The relationships, distribution and sedimentological characteristics of the Ciudad Granada formation suggest, that at the time of its deposition the Malaguide realm was subdivided into elongated areas showing either uplift and erosion or uplift and erosion followed by subsidence and deposition, which calls forth a picture of extensional horsts and grabens.

The basal beds contain e.g. *Nephrolepidina morgani* and *Miogypsinoides*; higher stratigraphic levels may contain *Miogypsina gunteri*. The planktonic foraminiferal and nannofossil assemblages reported from the red pelites (faunal lists in Soediono 1971; Gelati & Steininger 1984) indicate that the lower part of the formation is late Chattian (P22), and the upper part earliest Aquitanian (N4a + early N4b).

The Oligocene to lowermost Aquitanian Ciudad Granada formation carries exclusively Malaguide detritus and rests transgressively upon predominantly Triassic rocks of structurally relatively high Malaguide units. Nowhere could an angular unconformity be ascertained at its base. The formation does not seal thrust contacts but is imbricated with other Malaguide rock units. An example of such a relationship can be studied northeast of Vélez Rubio in the hills of la Tejera (Figures 4A, B, cross sections

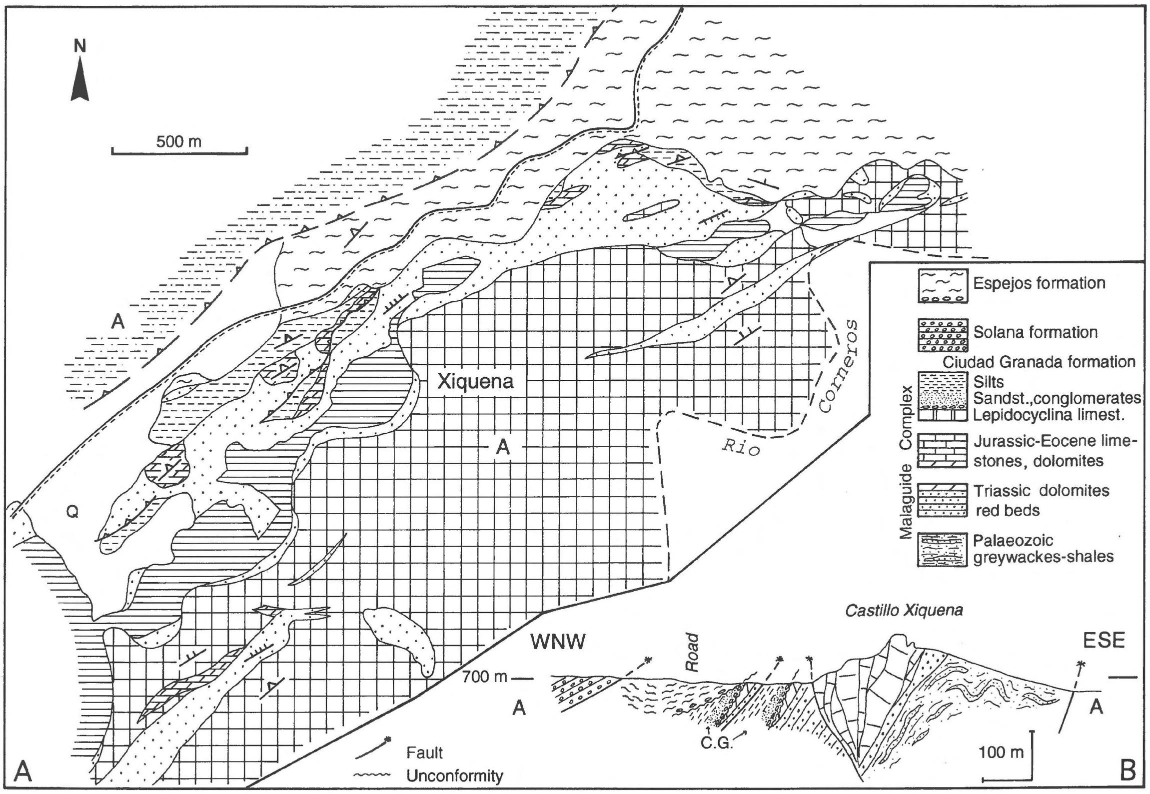

Figure 5. Xiquena area. A) Geological map showing location of cross section AA, modified after Roep (1972); for location, see Figure 1, for legend, Figure 3. B) Cross section AA. The Burdigalian Espejos fm, resting transgressively on tectonized Malaguide units, the Ciudad Granada fm (C.G.) included, is tectonically overlain by the Aquitanian Solana fm.

B1B1 and B2B2). Here, three superimposed imbricated units are separated by sub-horizontal, slightly undulating thrust contacts. The lowermost unit (I) consists of Palaeozoic greywackes and Triassic sandstones and dolomites transgressively overlain by the Ciudad Granada formation. The middle and upper units (II and III) show both a succession of Palaeozoic greywackes, Triassic sandstones and dolomites, and Jurassic and Eocene limestones. The pile of imbricated units is dissected by younger steeply north-dipping faults. The complicated structure of la Tejera is unconformably overlain by Burdigalian deep-water deposits (Espejos formation, see below), which are clearly post-thrusting and post-faulting. From the situation in la Tejera and also from other localities (Figure 4B, cross section AA; Figures 5A, B and 6A, B) it can be concluded that an important phase of thrusting succeeded by faulting took place after deposition of the Oligocene to lowermost Aquitanian Ciudad Granada formation but before sedimentation of the Burdigalian Espejos rocks.

Stratigraphically and structurally comparable to the Ciudad Granada formation are the basal parts of the Estepona flysch (Didon 1960) and the Pantano de Andrade formation (Bourgois 1978) in the Western Betic Cordilleras, and the Fnideq formation in the Rif (Wildi 1983; Feinberg et al. 1990) and the lower part of the transgressive Oligo-Miocene of the Kabylies (Gélard et al. 1973) in northern Africa. Generally, the 'Alozaine formation' (Bourgois et al. 1972) is also considered to belong to this suite of rock units. In the type locality of this formation, however, only imbricated units of Malaguide Triassic and Burdigalian breccias can be seen (T. Geel personal observation, corroborated by Martín-Algarra 1987).

Pliego formation. This formation is named after the village of Pliego, northeast of the Sierra Espuña. Its name was introduced by Jerez Mir (1981) for the assemblage of red *Almaena*-bearing silts and overlying varicoloured conglomerates, sandstones and silts in this area. Red pelites with *Almaena* are found everywhere in the Malaguides, but the varicoloured coarse

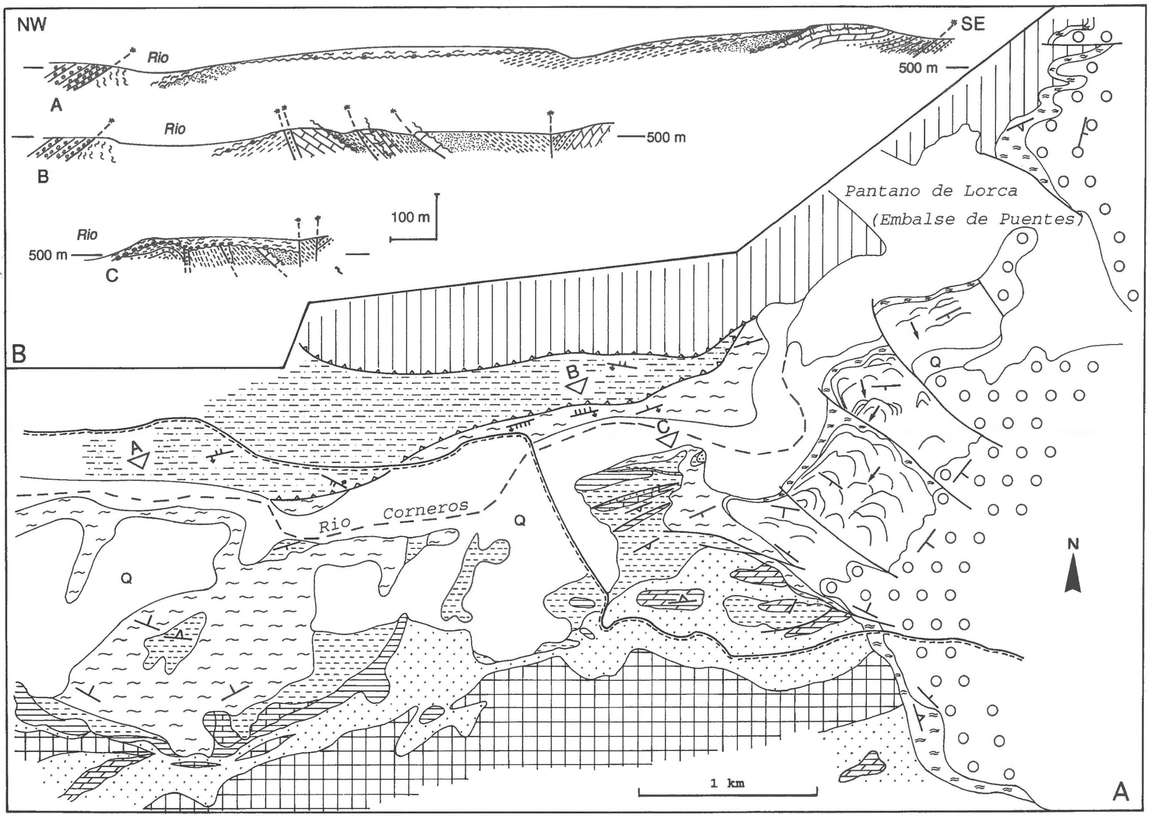

Figure 6. Pantano of Lorca area. A) Geological map showing relationships between Miocene formations and Malaguide and Subbetic deposits, and positions of cross sections A to C (for location, see Figure 1, for legend, Figure 3). The Burdigalian deep-marine Espejos fm unconformably overlies tectonized Malaguide sediments, the Ciudad Granada fm included. North of the Rio Corneros, the Espejos fm is tectonically overlain by the Aquitanian Solana fm which in turn is overthrust by the Subbetic. The tectonic contacts between the Subbetic, the Solana fm and the Espejos fm are sealed by Middle Miocene, Langhian, slope sediments. Serravallian deltaic deposits prograde, first southeastwards, then southwestward onto the Espejos fm. Lower Tortonian shallow-marine deposits of the Lorca Basin rest unconformably on the Middle Miocene formations. B) Cross sections A to C (for legend, see Figure 5B). The Espejos fm rests unconformably on folded and/or faulted Malaguide sediments, the Ciudad Granada fm included. The Solana fm is thrust over the Espejos fm.

clastic unit is restricted to the Espuña area. The two lithologies are partly coeval but were deposited in different environments (Figure 2). Therefore, Geel (1996) proposed a distinction in the Espuña area between red *Almaena* silts (Ciudad Granada formation) and varicoloured coarser clastics (Pliego formation).

The Pliego formation, up to 800 m in thickness, consists of green, brown, yellow, white and red micaceous silts, calcareous sandstones, clast-supported conglomerates and pebbly mudstones and of red, fine-grained ferruginous siltstone beds. Large-scale slumping, normal and inverse grading, and channelling indicate deposition by turbidites and mass flows in a submarine fan complex, including midfan and outer fan, and a basin plain. The formation carries detritus from all levels of the Malaguide Complex. It also contains a fair amount of phyllite and quartzite fragments derived from deeper levels of the Malaguide Palaeozoic (Geel 1996).

The Pliego formation can be assigned a Late Oligocene (P22) to earliest Aquitanian age on account of the presence of *Catapsydrax dissimilis, Cat. unicavus, Globigerina angustiumbilicata, G. tripartita, Globorotaloides suteri, Globigerinoides primordius, Neogloboquadrina mayeri, Globigerina woodi*, and at the very top, *Globoquadrina dehiscens*, and the presence of *Lepidocyclina* and *Miogypsinoides*. Martín-Perez et al. (1994) and Lonergan et al. (1994) reported NP25 and NN1 nannofossil assemblages, which corroborates our age determination.

Thus, whereas in the Vélez Rubio Corridor the red silt facies of the Ciudad Granada formation persisted into the earliest Aquitanian, this facies is replaced in the Espuña area during the Late Oligocene by submarine fans.

Martín-Martín (1996; cf. Martín-Martín et al. 1997; Guerrera et al. 1997) considered his Rio Pliego formation (= Ciudad Granada and Pliego formations of the present paper) to be the lateral, coeval equivalent of the Bosque formation, consisting of the Espuña limestone conglomerates and associated turbidites, marls, mass flows and carbonate-platform deposits. These formations would have been deposited into the same basin, but supplied by different sources, respectively situated to the north and southeast. This conclusion is mainly based on the allegedly close vertical and lateral relationships of the formations. As can be concluded from his maps, Martín-Martín included large parts of the outcrops of the siliciclastic Ciudad Granada and Pliego formations into the calcareous Bosque formation, probably considering the tectonic imbrication of the formations to be stratigraphic interfingering (for instance southwest of Pliego and east of the Almoloya, Figure 8A). Especially near Pliego, interfingering of the Bosque formation (here in carbonate-platform facies) and heavily slumped submarine (mid)fan deposits of the Pliego formation is hardly feasible. Moreover, the Bosque (or Espuña) conglomerates are of northeastern, not southeastern provenance.

Lower Miocene between the Subbetic Zone and Malaguide Complex

Between the Malaguide Complex and the Subbetic Zone, several Lower Miocene rock units can be distinguished. These are the allochthonous Solana formation and the autochthonous Espejos and Fuente formations (Figures 1, 2). In the Espuña area only the Espejos formation is represented (= El Niño formation of Martín-Martín 1996).

Espejos and Fuente formations

Soediono (in MacGillavry et al. 1963/64) introduced the names Fuente formation and Espejos formation to distinguish severely tectonized grey turbidite successions from younger, less tectonized olisthostrome-turbidite sequences occurring in the Chirivel area. Accordingly, all grey turbidite assemblages in the Vélez Rubio Corridor were called 'Fuente' on account of their supposedly strongly tectonized and chaotic state, whereas the Espejos formation was considered to be restricted to the Chirivel area (MacGillavry et al. 1963/64). After the realization that slump folds had been mistaken for tectonization and that difference in detritus was a more important criterion, the Amsterdam group incorporated into the Espejos formation all grey turbidite sequences of the Vélez Rubio Corridor carrying both Alpujarride and Malaguide detritus. The name Fuente became restricted to sediments without Alpujarride detritus. Hence, the 'Espejos' formation of Soediono (1970, 1971) and the 'Espejos' formation of Geel (1973) refer to the same rock unit which is continuous in the field. Hermes (1984) unfortunately recurred to the terminology of MacGillavry et al (1963/64), considering his 'Espejos formation' which is restricted to the Chirivel area, to be Serravallian. This caused considerable confusion in the literature afterwards (cf. Martín Algarra 1987; Sanz de Galdeano & Vera 1992). The extensive descriptions by Soediono (1971) and Geel (1973), under the heading 'Espejos formation', of hardly tectonized Burdigalian sediments, overlying with an angular unconformity tectonized Malaguide rocks and carrying both Alpujarride and Malaguide detritus, were ignored. Hermes called these sediments 'Fuente formation' of which he stated that detritus nor relationships were described. Consequently, an important observation for the dating of the final emplacement of the Internal Zone Complexes became lost.

Fuente formation (sensu Soediono 1971, non sensu Hermes 1984). The formation has been named after the Barranco (= dry gully) de la Fuente (Figure 1). It occurs in a narrow tectonized zone with a length of c. 1.5 km in the Chirivel area. Outside this strip it has not been recognized elsewhere in the Vélez Rubio Corridor. It consists of greenish-grey fine-sandy planktonic foraminiferal packstones, marls and silexites, and is devoid of Alpujarride detritus. Its lower boundary is always of a tectonic nature. Though the contact between the Fuente and Espejos formations is not exposed, Soediono (1971) inferred the existence of an angular unconformity despite the fact that a vertical fault zone delineates the boundaries of the outcrop. Thus, the nature of the contact remains obscure. The planktonic foraminiferal fauna of the formation indicates a latest Aquitanian to early Burdigalian age (Soediono 1971).

A qua age and lithology comparable unit of greenish-grey marls with intercalations of graded conglomeratic layers, found in an isolated, tectonically bounded outcrop in the midst of Malaguide rocks, was incorporated by Soediono (1971) in his Frac formation as member C. Members A and B clearly belong to the Ciudad Granada formation (see above). Like the Fuente formation, member C of the Frac formation carries only Malaguide detritus; the relation with member B (red pelites) seems to be conformable.

The Fuente formation may be compared with the 'Supra-Numidian' in the Rif and western Betic Cordilleras described by Didon et al. (1984).

Espejos formation (sensu Soediono 1970, 1971; Geel 1973; non sensu Hermes 1984). This formation has been named after the Barranco de los Espejos in the Chirivel area. It crops out in a zone that can be traced from the Chrivel area to the area east of Fuensanta (Soediono 1970, 1971; Figure 1). The formation is far less tectonized than the Malaguide and the Subbetic rocks in the Vélez Rubio Corridor. Except near tectonic contacts, dips are usually on the order of less than 20 to 30° (Figures 4A, B, 5A, B, 6A, B). Folds due to slumping and internal angular unconformities due to packet sliding may give the impression of more severe tectonization (Soediono 1971; Geel 1973).

Closely interbedded are: fining upward turbidite sequences (coarse conglomeratic to marl) of up to several dozens of meters in thickness, slumped deposits (up to 10 m thick), mud-supported debris flows, grain-supported mass flows, algal and larger-foraminiferal beds, graded oolite-bearing beds, silexites, tuffites and undisturbed greenish-grey marls. Fragments of the Malaguide Palaeozoic are by far predominant among the detritus of the formation. Alpujarride phyllites and quartzites are always found in lesser quantities and are sometimes absent in lower levels. Locally, in higher levels, some graded beds and mud-supported debris flows occur with predominant Subbetic detritus, e.g. Jurassic oolitic limestones, Cretaceous pelagic limestones and Aquitanian larger Foraminifera (Geel 1973). In the southern part of the outcrop in the Chirivel area, an exceptional lenticular body of coarse grain-supported mass flows (olisthostromes) is present consisting of large blocks, from c. 1 m to some tens of meters in size. These blocks are mainly derived from the Malaguide Complex with up-section an increasing admixture of metamorphic Alpujarride rocks. Occasionally, a lense of thin-bedded graded sandstone is intercalated. Also remarkable are large, up to several tens of meters, gently north-dipping tabular monomict breccias of Malaguide Palaeozoic greywackes. According to Soediono (1971) these tabular bodies, the very polymict character of the deposit and the occurrence of sandstone layers between the large blocks, rule out the possibility that the olisthostromes represent a tectonic breccia. Northward, up-section, there is also a decrease in size of the components: unbedded conglomerates are replaced by bedded turbiditic conglomerates and sandstones, the large boulders then being restricted to certain levels. According to Hermes (1984), the olisthostrome of Chirivel described above is a tectonic breccia, his 'Chirivel breccia', along a vertical fault zone. This explanation is at variance with the above observations.

As stated above, there is a general decrease in grain-size in the Espejos formation towards the north. Along-strike a similar change in lithology can be observed with the result that to the east of Vélez Rubio, the dominant lithology is formed by greenish marls with polymict turbidite intercalations in which large blocks still do occur locally, for instance north of la Tejera. On account of the lack of continuous section and the fact that the formation is tectonically overlain by Subbetic thrust masses, no reliable figure can be given for the thickness of the formation. The exposed thickness is of the order of several hundreds of meters.

The basal olisthostromal beds in the Chirivel area contain a poorly preserved and broken nannofossil assemblage that may belong to NN2. Further, a poorly preserved planktonic foraminiferal fauna has been reported from the basal beds indicating the *Catapsydrax dissimilis* Zone (N5; Gelati & Steininger 1984). Marly intervals between turbidite sequences and marl drapes on top of mass flows yielded in the entire Vélez Rubio Corridor a more diversified planktonic foraminiferal fauna of the *Globigerinatella insueta* Zone (N6; Soediono 1970). Considering that only the basal few meters yielded a N5 fauna, that the duration of biozone N5 is rather long (2.7 Ma), and that the sedimentation rate is high in this type of deposits, it is highly probable that only sediments corresponding to the uppermost part of biozone N5 are present near the base. Several samples of higher levels, again over the entire Vélez Rubio Corridor, yielded assemblages with e.g. *Globigerinoides diminutus, Gd. bisphericus, Gd. altiaperturus*. Given the absence of *Praeorbulina*, the assemblages can be assigned to biozone N7. These data result in a Burdigalian (uppermost N5, N6 and N7) age for the Espejos formation.

Hermes (1984) claimed his 'Espejos' formation, that is, only the outcrop in the Barranco de los Espejos, to be late Serravallian. Resampling by González Donoso of this outcrop in 1986, resulted in a late Burdigalian age on account of the presence of *Globigerinoides bisphericus* (Martín-Algarra 1987). This result is in keeping with our own data and conclusions mentioned above.

The southern (basal) boundary of the Espejos formation is usually hidden under soil or scree. Where it is well exposed, the relation is often disturbed tectonically. There are, however, several important localities, where the original stratigraphic contact can be studied, e.g. La Tejera (Figure 4), an especially good outcrop near Tirieza baja 2 km E of Xiquena (Figure 5), and near Fuensanta (Figure 6). These outcrops show that the Espejos formation rests transgressively, with an angular unconformity upon imbricated Malaguide rocks, the Ciudad Granada formation included. The basal beds are formed by conglomerates or turbidites or by deposits which contain but little terrigenous detritus. In all instances, they contain immediately a planktonic foraminiferal fauna and a high share of radiolarians indicating a depositional depth of 800 to 1000 m. All facts considered point to an extremely rapid subsidence at or before the onset of deposition, from c. 200 m during the Aquitanian part of the Ciudad Granada formation to 800 to 1000 m during the early Burdigalian (MacGillavry et al. 1963/64; Gelati & Steininger 1984).

Only to the south of the Pantano of Lorca, the Espejos formation is overlain (unconformably) by Middle Miocene sediments (Figure 6A). Everywhere else, the northern, i.e. top, boundary of the formation is formed by a gently north-dipping thrust plane: from the Pantano of Lorca to 5 km W of Vélez Rubio, the formation is tectonically overlain by the Solana formation (Figures 1, 4–6), and from 5 km W of Vélez Rubio to Chirivel by Subbetic rocks (Figure 1). The boundary is nowhere a vertical strike-slip fault zone as suggested by Hermes (1985). The text and cross section (enclosure) of Soediono (1971) are rather inconsistent on this subject. In the text, the contact between Subbetic and Espejos formation is said to be a zone of tectonic disturbances. On the cross section, the Espejos formation is depicted as onlapping upon Subbetic rocks. This controversy is due to the use of a wrong version of the section (Soediono, personal communication 1971) and added to the confusion.

The features described above indicate, that the Espejos rocks were deposited in a submarine fan complex at a depth of 800 to 1000 m. The outward thinning and fining, away from the Chirivel area, and upward thinning and fining of the deposits suggest that the apex of the fan and the feeder channel were located in that area, though isolated channel fills do occur outside it. The presence in the Espejos formation of both Malaguide and Alpujarride detritus and, in higher levels also of Subbetic components, e.g. Jurassic oolitic limestones devoid of quartz grains in contrast to those of the Malaguide Complex, and the unconformable position upon Malaguide imbricated units, including the Ciudad Granada formation, witness the denudation of superposed and partly metamorphic Betic units and the juxtaposition of Internal and External Zone during the Burdigalian (Geel 1973).

Comparable rock units have been described from elsewhere in the Betic Cordilleras and their equivalents in northern Africa, e.g. the Millanas formation (Bourgois et al. 1972), Viñuela formation (Boulin et al. 1973), Alamo Series (Völk & Rondeel 1964), Cartama section (Sanz de Galdeano et al. 1993), and the upper part of the transgressive Oligo-Miocene in the Kabylies (Gélard et al. 1973).

Solana formation

The name Solana formation is derived from the Cortijo (= farmhouse) Solana near Xiquena. The main outcrops occur in the Vélez Rubio Corridor proper. Besides, the formation has also been found in tectonic windows in the Subbetic Zone (Figures 1, 7).

Two different types of siliciclastics can be distinguished, both interbedded with varicoloured (brownish-green, brown, ochreous-yellow or grey) clay-rich marls. The first and probably the older type (Quartz-sandstone member of Geel 1973; Solana III of Hermes 1984) consists of very thin-bedded to medium-bedded quartz sandstones, made up of densely packed, well-rounded 'frosted' quartz grains, and thin-bedded, sandy bioclastic packstones. They are associated with graded larger-foraminiferal packstones (Solana II of Hermes 1984) and silexites. In this type of arenites, Bouma's turbidite divisions are common. The second type (Polymict sandstone member of Geel 1973; Solana I of Hermes 1984) comprises mostly thick-bedded or even massively bedded polymict sandstones containing conglomeratic streaks and lenses, and clay pebbles. They are likewise associated with graded larger-foraminiferal bioclastic packstones. These sandstones frequently show amalgamated grading. Lenticular bodies of boulder and pebble-bearing mudstones and sandstones also occur. Hence, the quartz sandstones are of a 'distal' outer fan type, the polymict sandstones of a 'proximal' fan type. The components in the polymict sandstones and conglomerates are all derived from the Malaguide Complex.

The fossil content of the larger-foraminiferal packstones indicates an Aquitanian age. The marls contain a poor, largely reworked planktonic foraminiferal

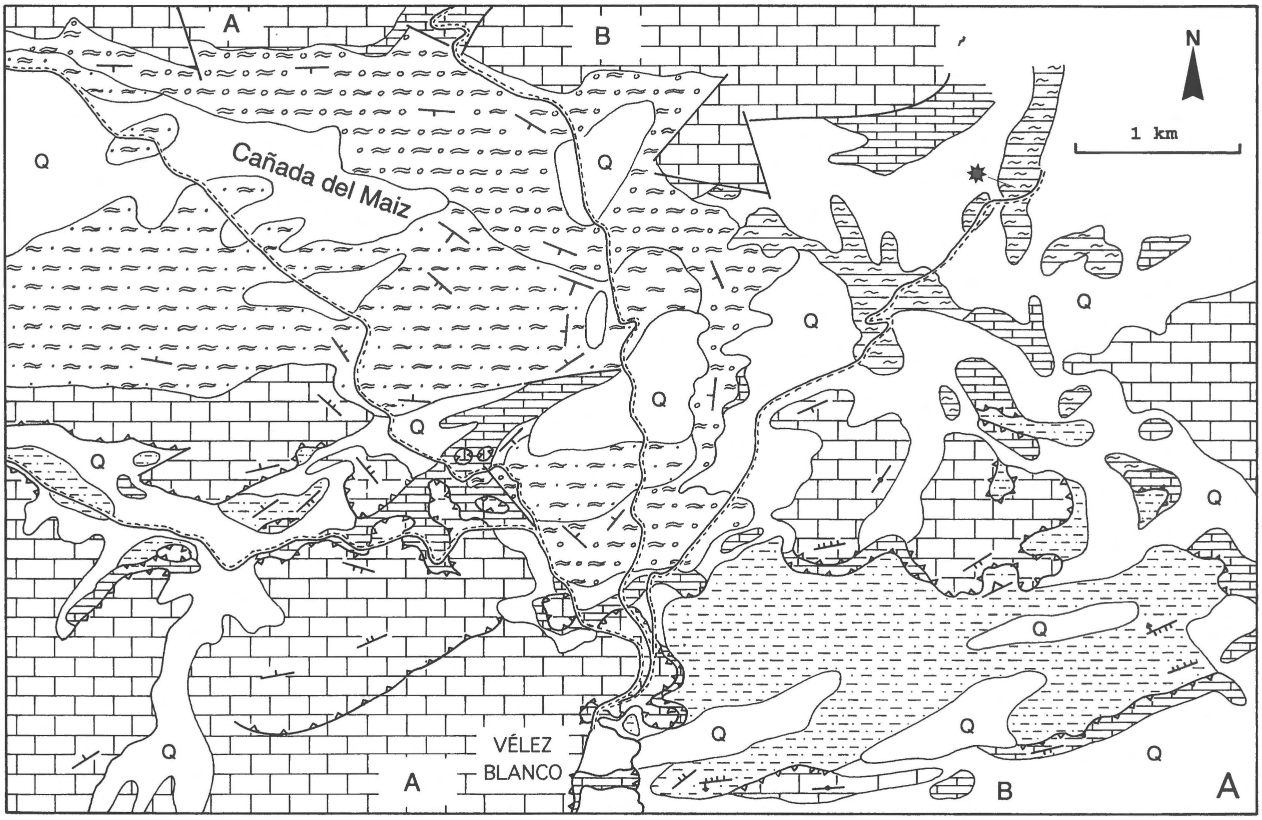

Figure 7A. Geological map of the Vélez Blanco-Cañada del Maiz area (for location, see Figure 1, for legend, Figure 3). The Middle Miocene Maiz I and Maiz II fms cover Subbetic thrust masses unconformably. In the tectonic double window of Vélez Blanco, Subbetic Jurassic dolomites rest tectonically upon Subbetic Cretaceous pelagic limestones, whereas the Subbetic Mesozoic carbonates are thrust over the Aquitanian Solana fm.

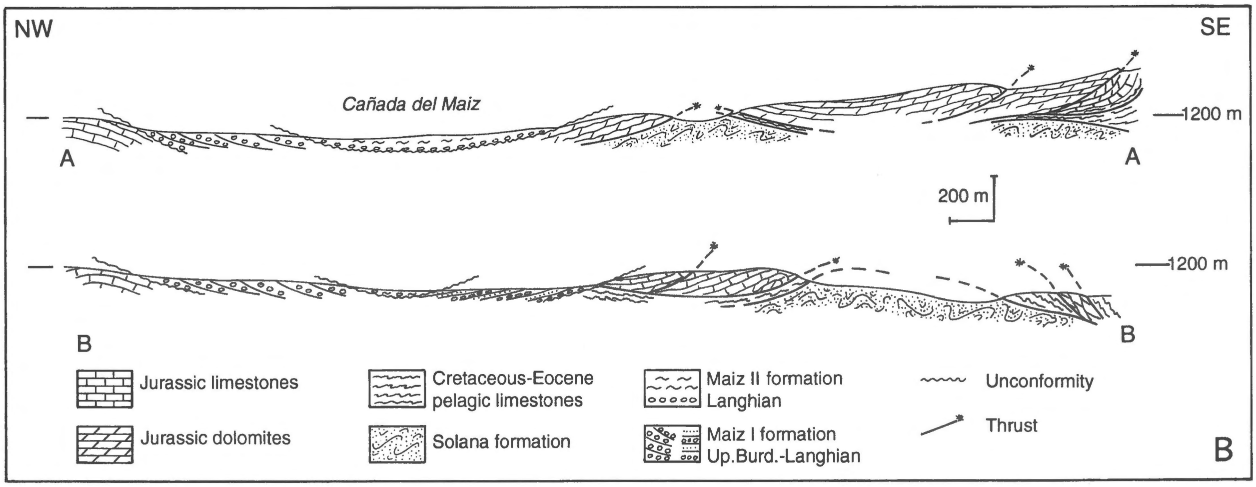

Figure 7B. Cross sections of the Vélez Blanco-Cañada del Maiz area (for location, see Figure 7A). Cross section AA: unconformable position of Langhian deposits upon Subbetic thrust masses covering tectonically the Solana fm. Cross section BB: Langhian deposits resting transgressively upon the tectonic double window of Vélez Blanco.

fauna, which points to a middle to late Aquitanian or younger age (faunal lists in Geel 1971). Geel (1973) argued that the Solana formation must be older than the Espejos formation, based on the facts that the Solana formation only contains Malaguide detritus and no Alpujarride detritus, whereas the Espejos formation carries both, and the occurrence of Solana fragments in the Espejos formation. Therefore, the Solana formation has to be of Aquitanian (N4B and early N5) age.

The Solana formation is always bounded by tectonic contacts. It overlies the Espejos formation tectonically, with gently north-dipping or near-horizontal thrust planes. It is, in turn, overlain by Subbetic rocks, again with gently north-dipping thrust planes (Figure 7B).

Hermes (1984) interpreted the Solana outcrops of the Vélez Rubio Corridor to consist of tectonic blocks with fundamentally different facies: shallow-water Solana I, and deeper-water Solana II and III. Given their allegedly steep contacts, these blocks would mark a vertical transcurrent fault zone along which fragments from Gibraltar were dragged eastward to their present position. However, firstly, 'Solana II'-type and 'Solana III'-type sandstones are closely associated: often the base of a specific bed is formed by densely packed larger Foraminifera, the top of the same bed by quartz sandstone. Secondly, outside the Vélez Rubio Corridor, for instance near Colmenar in the Western Betics, quartz sandstones ('Solana III'-type) and polymict sandstones ('Solana I'-type), can be seen to alternate (Geel, personal observation 1967; Martín Algarra 1987). Thirdly, the contacts are not vertical but dip gently northward.

The Quartz-sandstone member is microscopically indistinguishable from the Aljibe Sandstone of the Campo de Gibraltar, the Polymict sandstone member is identical to the Algeciras Flysch (Geel 1973). There is, however, megascopically a difference because the Quartz sandstones show a distal fan facies, the Aljibe Sandstone a very proximal midfan facies, whereas the Polymict sandstones of Vélez Rubio are of a proximal midfan type against the outer fan facies shown by the Algeciras Flysch.

Middle Miocene deposits in the Southern Subbetic and along the Internal-External Zone Boundary

Middle Miocene, uppermost Burdigalian to Serravallian, deposits overlie unconformably rocks of both the Internal and External Zone. They occur to the north of Vélez Blanco (Cañada del Maiz), along the northwestern rim of the Lorca Basin (Campico de Flores, Pantano of Lorca) and northwest of the Sierra Espuña (Bernabeles area; Figures 1, 2).

Cañada del Maiz

Slightly folded calcareous and marly deposits cover the strongly tectonized Subbetic rocks (Jurassic limestones, Cretaceous pelagic limestones and Palaeogene to Early Miocene marls and turbidites) with a marked angular unconformity. Two formations can be distinguished, separated by a slight angular unconformity: Maiz I and Maiz II (Figures 7A, B).

Maiz I formation. The first formation (up to 150 m thick) is found along the northern and eastern border of the basin. It consists of boulder and pebble-bearing oolitic calcarenites and boulder-pebble conglomerates, all sediment-gravity flows in which transport occurred by grain-to-grain interaction of the oolites. Some, with a more marly matrix, may be debris flows. Further, there occur olistholiths of Jurassic limestones (up to 30 m long), bedded bioclastic calcarenites and marly interbeds with planktonic faunas. Slumping is common. Laterally, the coarse beds show a distinct thinning and fining, the marly interbeds increase in thickness. Calcareous components, derived from the Subbetic are predominant. Noncalcareous detritus is subordinate. The fan shape, predominance of mass flows and mega-foresetting to the southwest suggest that the Maiz I formation represents a southwestward prograding fan delta.

The marly interbeds contain a fairly well preserved planktonic foraminiferal fauna that can be assigned to N8 indicating a Langhian age for the formation (Hermes 1977). Hermes' (1977) *trilobus* Zone fauna which he stated to have collected from the base of the formation in one of his sections, was probably sampled in underlying Subbetic Burdigalian (N5) sediments.

Maiz II formation. In the north and east, the second formation overlies the Maiz I delta with a slight angular unconformity. In the south, it rests with a marked angular unconformity on thrust masses consisting of Subbetic rocks and Solana formation (Figures 7A, B). It consists mainly of light-coloured marls and calcarenites with intercalations of thin-bedded, graded turbidites and some debris flows. The detritus is more polymict than in Maiz I and comprises besides predominant Subbetic oolitic limestones and Cretaceous pelagic limestones also a fair amount of Alpujarride metamorphic components: phyllites, quartzites, and dark dolomite.

The age of Maiz II is difficult to ascertain, because the planktonic foraminiferal fauna consists largely of reworked Cretaceous to Early Miocene species; the only younger species are *Globigerinoides bisphericus* and *Globorotalia scitula. Orbulina* has not been reported. The formation shoals rapidly up-section towards continental conditions (Hermes 1985). Consequently, shoaling occurred before the Langhian-Serravallian boundary.

Campico de Flores (El Campico) and Pantano of Lorca

Middle Miocene sediments are found all along the northwestern border of the Lorca Basin. A complete section can be studied in Campico de Flores (or El Campico), west of the Lorca-Caravaca road (Figure 1). The relationships between Lower, Middle and Upper Miocene deposits are well exposed around the Pantano of Lorca (Figure 6A).

Campico de Flores (El Campico). The Middle Miocene deposists here reach a thickness of up to 500 m. The lower 300 m are slope deposits made up by yellow, rhythmically (medium) bedded sandy planktonic foraminiferal packstones with gravity-induced pull-aparts and marly interbeds. Fining-up and thinning-up, greenish-grey, badly sorted polymict conglomerates and laminated sandy packstones represent intercalated mass flows and turbidites. The clasts of the conglomerates are hardly rounded and reach lengths of up to 3 m. They consist of Malaguide Palaeozoic greywackes, Malaguide Triassic sandstones and dolomites, Alpujarride phyllites and quartzites, and bivalves (oysters and pectinids), and are embedded in a matrix of sandy to conglomeratic bioclastic packstone. One of the marly interbeds is topped by a slump-debris flow of several meters thickness, with Alpujarride and Malaguide components. Near the top of the lower 300 m, Subbetic material and reworked *Nummulites* also occur. The upper 200 m of the formation in El Campico consist of badly exposed, slumped, grey-green marls. These are succeeded by ochreous-yellow, medium-bedded to massive, polymict sandy bioclastic packstones (up to 50 m thick), in which two distinct thickening-up sequences can be observed.

The lower boundary of the Middle Miocene is mostly hidden under soil or scree. In one locality, 400 m due west of the cortijada (= hamlet) of Campico de Flores, the boundary is well-exposed: sandy-pebbly planktonic foraminiferal packstones onlap northward a hardground on top of truncated Subbetic larger-foraminiferal packstones. The upper boundary of the Middle Miocene is formed by a flat, eroded and bored surface, settled by oysters, and transgressively overlain by lower Tortonian, shallow-marine bioclastic limestones.

The Campico deposits contain a complete *Praeorbulina-Orbulina* bioseries (F. Bianchi personal communication) from *Globigerinoides bisphericus* at their base via Praeorbulinids to *Orbulina suturalis* in the marls below the sandy bioclastic packstones at the top. This indicates the upper N7, the N8 and the N9 biozones and a latest Burdigalian, Langhian and early Serravallian age. The change from predominantly coarse siliciclastics and planktonic foraminiferal packstones to dominantly marly deposits is within the lower part of biozone N8. The change from dominant marls to thick-bedded packstones occurs in biozone N9.

Pantano of Lorca (Figure 6A, B). The coarse clastic lower part of the Campico deposits reappears from below the Tortonian fill of the Lorca Basin in outcrops along the northern border of the reservoir and near the Lorca-Fuensanta road to the south of the lake (Figures 1, 6A). Marly interbeds between the coarse polymict graded conglomerates contain *Praeorbulina* sp. indicating a Langhian age. The deposits straddle the External-Internal Zone Boundary because in the north they rest on tectonized Subbetic limestones, in the south on Malaguide rocks. The equivalent of the marly upper part of the Campico deposits is hidden below recent lake deposits or Tortonian shallow-water sediments. However, the equivalent of the Serravallian thick-bedded bioclastic top of the Campico section is found again along the southeastern border of the Pantano of Lorca (Figure 6A). The southeastern steep border of the lake is formed by c. 20 m of white barren marls with intercalations of thickening-up packages of fine calcarenites. These marls are overlain by medium to thick-bedded, coarse sandy and pebbly, algal and bryozoan packstones. The conglomeratic parts of the beds show inverse grading. The clasts are hardly rounded and consist predominantly of Subbetic Upper Jurassic limestones. Minor amounts of Malaguide Palaeozoic and Triassic occur at the bases of the conglomeratic intervals. Locally, reworked *Nummulites* are abundant. The coarse packstones are organized into prograding deltaic lobes. The lower sets pro-

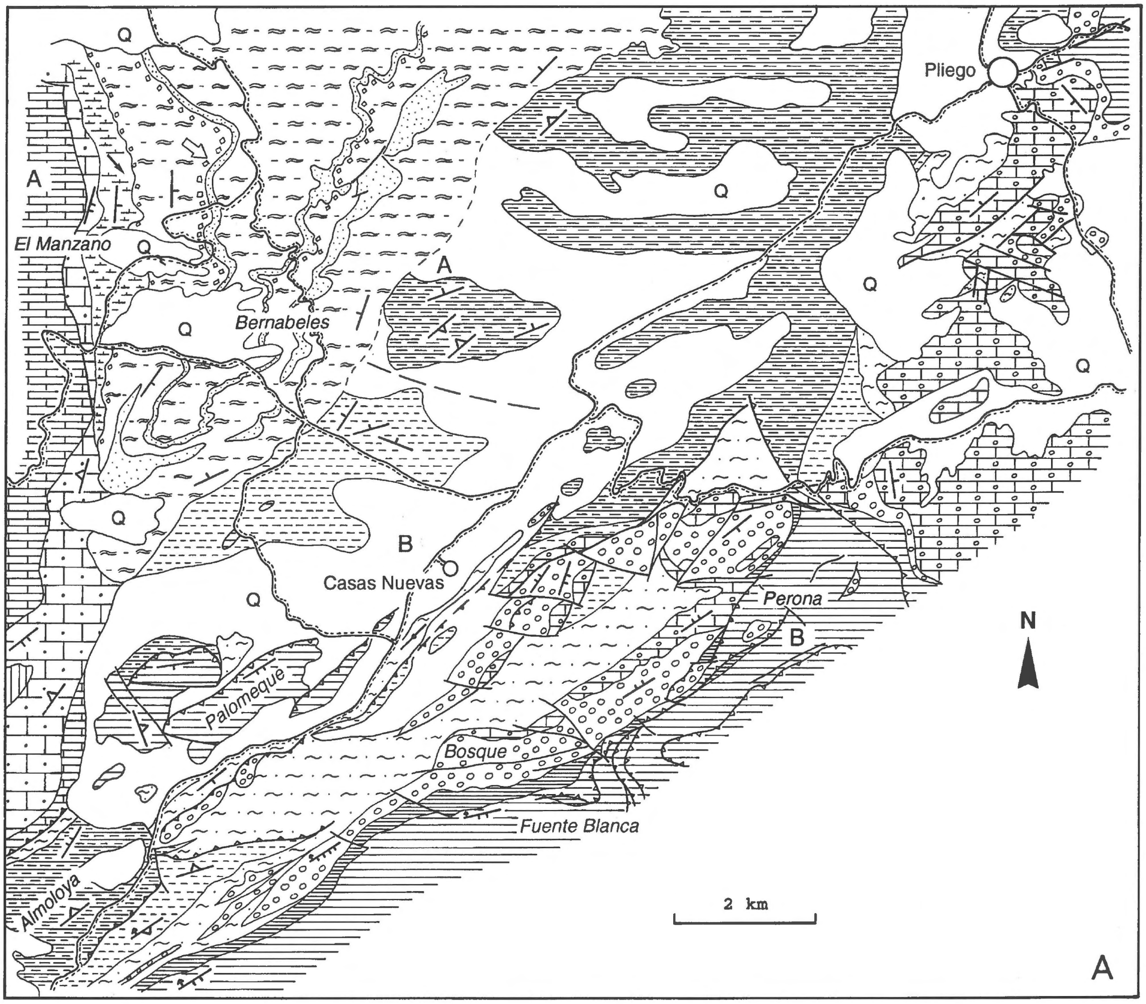

Figure 8A. Geological map of the Internal-External Zone Boundary northwest of the Sierra Espuña (for location, see Figure 1, for legend, Figure 3). The Middle Miocene slope deposits of the Bernabeles area rest onlapping upon Subbetic sediments on the west side, and upon Malaguide Upper Oligocene to Aquitanian deep-shelf and slope-fan deposits (Ciudad Granada and Pliego fms) on the east side. In the Almoloya area, Subbetic sediments are thrust over the Espejos fm (here only represented by lower Burdigalian deposits). The Ciudad Granada and Pliego fms being the youngest sediments present in the Malaguide imbricated units, the main thrusting of the Malaguides occurred after their deposition, after the earliest Aquitanian.

grade towards the southeast, the upper sets towards the southwest (Figure 6A). The deltaic deposits are transgressively overlain by lower Tortonian shallow-marine deposits of the Lorca Basin.

The deltaic fan lobes prograde southwestward beyond the underlying marls and fine calcarenites, interpreted as pro-delta deposits, onto the marls, planktonic foraminiferal and fine-sandy turbidites of the Espejos formation (Figure 6A). The latter formation in turn rests unconformably upon imbricated units of the Malaguide Complex, Ciudad Granada formation included (Figure 6B).

Bernabeles area

The Middle Miocene sediments of the Bernabeles area straddle the boundary between the Internal and External Zones northwest of the Espuña (Figures 8A, B). On the west side they onlap onto folded and imbricated Subbetic Cretaceous to Oligocene limestones, on the east side they rest with an angular unconformity on either the Pliego or the Ciudad Granada formation. There is a certain west-east asymmetry between the stratigraphic development in the lower part of the deposits.

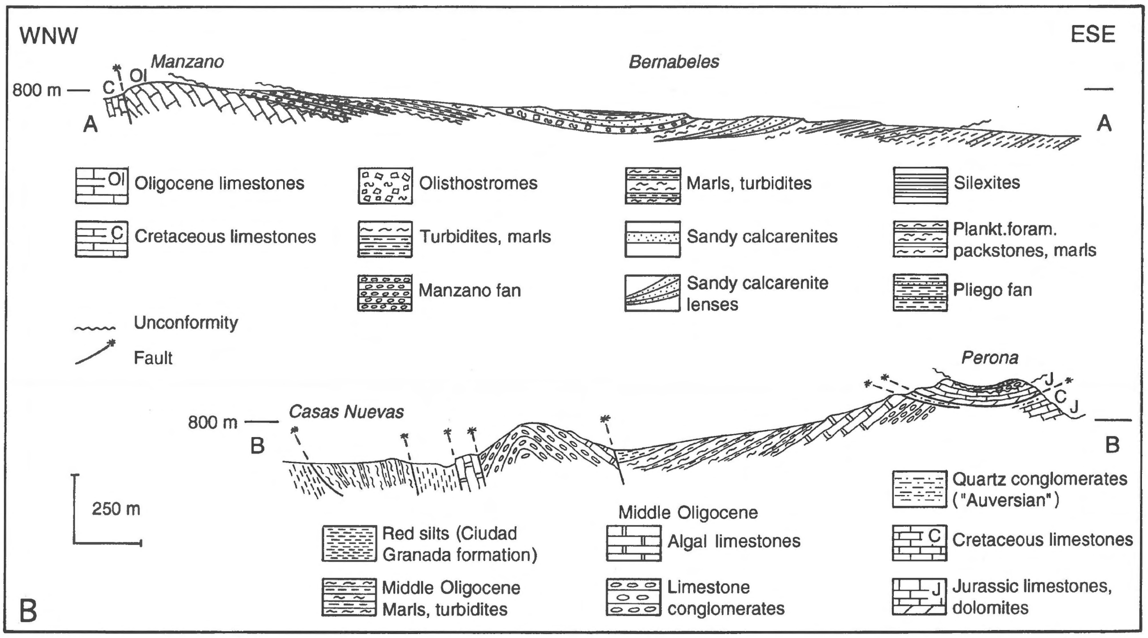

Figure 8B. Cross sections of the Bernabeles area and northwestern part of the Espuña (for location, see Figure 8A). Cross section AA shows the asymmetry of the Middle Miocene Bernabeles deposits (description in the text). Cross section BB shows that main thrusting post-dates Oligocene sedimentation: the imbricated unit of the Perona, comprising Jurassic dolomites and limestones, covered with an erosional unconformity by Middle Oligocene limestone conglomerates, is thrust northwestward over similar Middle Oligocene limestone conglomerates and algal limestones. Near Casas Nuevas, steeply dipping imbricated units consist of Middle Oligocene yellow marls and turbidites and red silts of the (here) Upper Oligocene Ciudad Granada fm.

On the west side, on the eastern flank of El Manzano, the base is formed by an up to 100-m-thick formation consisting of bedded pebbly bioclastic packstones, laminated bioclastic packstones, clast-supported, pebble and boulder-bearing mass flows, mud-supported pebble and boulder-bearing debris flows, and subordinate marls. Normal and inverse grading, channelling, intraformational unconformities and upwards thickening and coarsening as well as upwards thinning and fining sequences have been observed. The transport direction is towards the southeast. The badly rounded clasts have been derived from the Subbetic: e.g. Triassic black dolomites, predominant near the base, Jurassic limestones, Cretaceous pelagic limestones, *Lepidocyclina* limestones, and diabase. The beds contain also reworked larger Foraminifera, e.g. *Nummulites, Eulepidina, Miogypsinoides* and *Miogypsina*. The fan-shaped geometry and characteristics of mass gravity flows, indicate that this basal formation is a submarine fan composed of debris flows. The intraformational unconformities are probably induced by synsedimentary tectonics.

According to Paquet (1969), IGME (1972) and Lonergan et al. (1994) the above fan deposits are Aquitanian and belong to the Subbetic. However, these sediments onlap onto tectonized Subbetic rocks, including Aquitanian limestones, and show the same gentle eastward dip as the overlying uppermost Burdigalian sediments (Bernabeles formation of Lonergan 1991). Given the fact that the Subbetic became tectonized during the Burdigalian, the sediments of the Manzano fan have to be younger, that is of (?middle) late Burdigalian age.

The fan of El Manzano is overlain by a 15-m-thick, pebble and cobble-bearing, mud-supported debris flow containing clasts of Subbetic limestones, Malaguide Palaeozoic greywackes and Alpujarride phyllites and quartzites. Next follows a 120-m-thick, upwards thinning and fining turbidite-marl sequence. The turbidites, sandy planktonic foraminiferal packstones, show several of the Bouma divisions and flute casts. The turbidite-marl sequence is topped by an up to 90-m-thick olisthostrome, a slump-debris flow deposit, in which slump overfolds of sandy planktonic foraminiferal packstones, and pebbles and blocks (up to 125 cm long) of Subbetic, Malaguide and Alpujarride derivation, float in a sandy to marly matrix.

The slump overfolds indicate a slope inclined to the southeast.

On the east side, the base of the Bernabeles deposits is formed by light-coloured planktonic foraminiferal turbidites, silexites, and relatively sandy turbidites rich in Alpujarride phyllites. Intercalated large lenticular bodies (up to 25 m thick) cut erosively into underlying marls. They are made up of slightly sandy bioclastic packstones. Small and large-scale channelling is common. We interpret these bodies to represent stacked, laterally migrating channels. This more marly and silexitic development is topped by the same olisthostrome recognized on the west side of the basin fill. However, on the east side this level is thinner (c. 40 m), slump overfolds are absent, and the clasts are smaller.

From the olisthostrome upwards, the stratigraphy shows a uniform development: 15 m of coarse bioclastic packstones representing liquefied and/or grain-flow deposits, are succeeded by marls (up to 80 m thick) with intercalations of bioclastic turbidites, and at the top silexites.

In view of the above-described sedimentological characteristics and asymmetric stratigraphy, the entire succession of the Bernabeles area can be interpreted as a submarine slope fan with a more distal facies in the south and east and a more proximal facies in the north and west.

The deposits below the olisthostromal level can be assigned to the uppermost part of the N7 biozone below the first appearance datum (FAD) of *Praeorbulina*, on account of the presence of *Globigerinoides altiaperturus, Gd. bisphericus* and *Gd. diminutus*. Sediments above the olisthostrome contain the first *Praeorbulina transitoria* together with *Globigerinoides bisphericus* and can be assigned to the lower part of N8, below the FAD of *Praeorbulina glomerosa*. The youngest sediments did not yield a determinable fauna. Martín Algarra (1987) reports the presence of orbuliniform Foraminifera (either *Praeorbulina glomerosa* or *Orbulina suturalis*). The conclusion is therefore warranted that the sediments of the Bernabeles area are of latest Burdigalian to Langhian and perhaps early Serravallian age (late N7, N8 and N9?).

According to Lonergan et al. (1994), her Bernabeles Formation follows, without a break in sedimentation, upon sediments of her Amalaya (= Pliego) Formation; they would, younging northward, onlap onto Subbetic rocks. This interpretation, however, appears to result from erroneous sampling and a misinterpretation of nannoplankton data (Geel 1996). Considering the stacked channels described above to be tectonically imbricated units, Lonergan et al. (1994) concluded that an important SE-directed thrust zone within the Bernabeles deposits causes duplication of calcarenite bands and up to 50% of tectonic shortening. The alleged tectonic duplication of calcarenite bands concerns two sedimentologically different units which can both be followed around the southern closure of the syncline (shown in Figure 8A).

Martín-Martín (1996) held the Bernabeles formation to be deposited on a shallowing-upward mixed siliciclastic-carbonate platform not extending far beyond the present outcrop. The main argument for a shallow-water environment would be the presence of shallow nannoplankton species, a not very reliable criterion in turbidite deposits. Given the remarkable similarity in development of the Campico and Bernabeles deposits, especially the occurrence of slump-debris flows at the same stratigraphic level, it is more plausible that the Campico and Bernabeles slope deposits and submarine fans were deposited in the same basin extending from at least the Pantano of Lorca in the southwest to Mula in the northeast.

Discussion and conclusions

It is now generally accepted that the Internal Zones of the Spanish Betic Cordilleras and the Moroccan Rif, together the Betic-Rif Internal Zone, once formed part of a West Mediterranean continental segment in eastward continuation of Iberia. The northern part of this segment, the North Sardinian block, comprising Sardinia, Corsica and Calabria, rotated anticlock-wise during the Early Miocene, with concomitant opening of the Provençal Basin on its west side and formation of the Apennine orogen on its east side (Alvarez et al. 1974; Rehault et al. 1984). The southern part, called Alboran microplate, or more aptly South Sardinian block (cf. Sanz de Galdeano 1990), disintegrated and its fragments migrated and collided with the African plate in Sicily (Peloritan mountains) and Algeria (Kabylies), and with both the African and European plates in southern Spain and Morocco (Betic-Rif Internal Zone). In various proposed models, the forming of the southwestern Mediterranean basins (Alboran Sea-Algerian Basin) is thought to be connected in some way or other with this dispersion and migration (see review in Sanz de Galdeano 1997). A discussion of the geodynamic merits of the proposed models is outside the scope of the present paper. We will restrict ourselves to the presentation of data constraining the chronology of the events (Figure 9).

In this context, it is apparent that understanding the geology of the Vélez Rubio Corridor and Espuña area is important because here the contact between the Internal and External Zones is located. The former, allochthonous zone was once situated several hundreds of kilometers eastward, the latter represents the southern, passive margin of Iberia.

In the first part of the paper, we presented the results of our research and analysis of the Vélez Rubio Corridor and Espuña sediments. In the following paragraphs we will reconstruct the successive stages in basin development (Figure 9). In this paper we concentrate on the chronology of events in the Corridor and Espuña areas. A comparison with sediments from more external (North Subbetic and Prebetic) Zones and from elsewhere in the Western Mediterranean, and a discussion of their West Mediterranean plate-tectonic context will be given in a separate paper (Geel & Roep 1998).

Early and early Late Oligocene

During this time span, the Southern Subbetic formed the southern, more basinal part of a platform-slope-basin configuration of the Iberian passive margin. The Malaguides of the Vélez Rubio Corridor and Espuña area belonged to the South Sardinian block. In Eocene times, a shallow carbonate platform stretched from the Malaguides now found in the Vélez Rubio Corridor to those now in the Espuña. In Oligocene times the Malaguides of the Vélez Rubio Corridor were uplifted and eroded whereas the Espuña Malaguides subsided. The Espuña site at first shows a transgressive shallow shelf sequence, subsequently followed by fan deltas made up of limestone conglomerates. Given the nature of the detritus and the direction of prograding and supply, the site of the Espuña area, the easternmost Malaguides, forming part of the South Sardinian block, was still juxtaposed to or rather connected with the North Sardinian block (Sardinia; Geel 1996).

Late Late Oligocene to earliest Aquitanian

This period is characterized by the deposition of the Ciudad Granada formation. The Late Oligocene transgression and drowning of part of the Malaguides, accompanied by a change in source area of detritus, from extra-Malaguide to intra-Malaguide, and the deposition of the Ciudad Granada formation in grabens, suggests that the distribution of this formation was controlled by extensional tectonics. Extension in the Malaguides was coeval both with West Mediterranean rifting and with a phase of extension and subsequent reheating in deeper Betic units (De Jong 1991). This late phase of metamorphism in the deeper Betic units was preceded by an earlier phase of nappe forming and metamorphism during Cretaceous subduction (Biermann 1995 and references therein). The submarine fans of the partly coeval Pliego formation were deposited into a more strongly subsiding area, probably the first sign of the imminent separation of the North and South Sardinian blocks. However, this disintegration had not yet far advanced, as can be concluded from the presence of clasts of deeper Malaguide crystalline basement in the Ciudad Granada formation near Chirivel. Given the fact that a crystalline basement is unknown in the Spanish Malaguides but does occur in their Kabylian equivalent in Algeria, it can be concluded that the Algerian Basin, now separating the Spanish Malaguides from the Kabylies, had not yet opened (Geel 1996). During the same time, slope sedimentation continued in the Southern Subbetic.

Aquitanian

After their deposition and before the Burdigalian, the Ciudad Granada and Pliego formations were strongly tectonized into imbricated units, together with subjacent Malaguide sediments. Consequently, an important phase of deformation affected the Malaguide Complex during the Aquitanian. This phase can be correlated with the phase of inversion tectonics in deeper Betic units (De Jong 1991). Given their distribution and detritus, the allochthonous Aquitanian Solana submarine fans were deposited outside but close to the Malaguides. Because slope sedimentation, with southwest-ward supply, continued in the Subbetic, the Internal and External Zones must still have been wide apart (cf. Geel et al. 1992; Geel 1995). All data combined suggest that disintegration of the South Sardinian block and subsequent southwestward movement, thrusting and uplift of the Betic Internal Zone fragment started during the Aquitanian.

Aquitanian-Burdigalian boundary

The first indication for initial disruption of the Subbetic slope is in the late Aquitanian to early Burdigalian green marls with silexites of the Upper Green Pelite. A nearness of External and Internal Zones, i.e. of Subbetic and Malaguides, is suggested by the occurrence of comparable facies in association with the

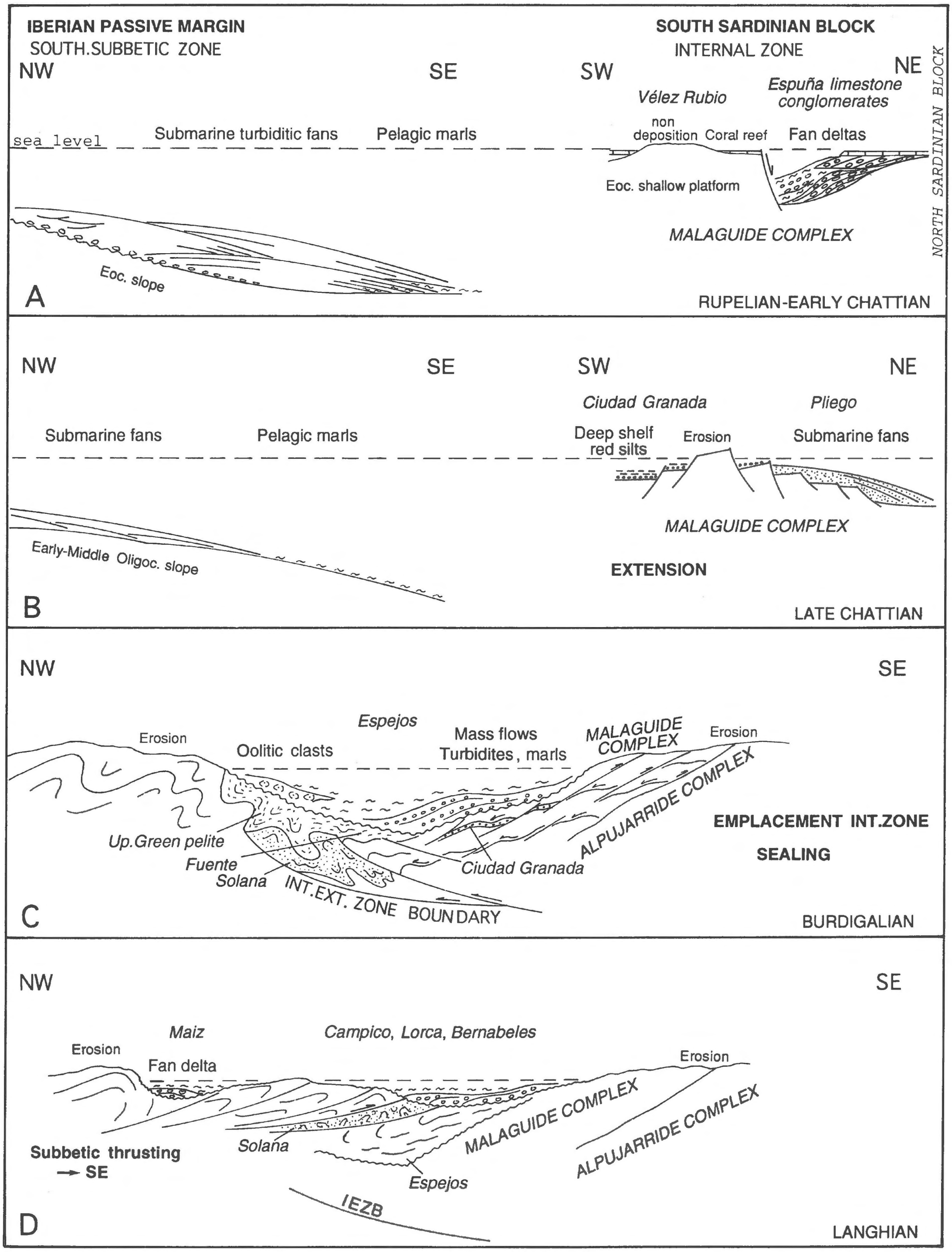

Figure 9. Cartoons illustrating the tectono-sedimentary history of the Vélez Rubio Corridor and Espuña area during Oligocene to Middle Miocene times (not to scale), showing that the Internal-External Zone boundary (IEZB) became successively covered under the Burdigalian Espejos fm and under Subbetic thrust masses. The Subbetic thrusting is sealed by Middle Miocene deposits. The original distance between the Iberian South Subbetic slope and the Malaguide Complex, belonging to the South Sardinian block, was c. 400 km.

Malaguides: the Fuente formation and member C of the Frac formation (cf. MacGillavry et al. 1963/64).

Burdigalian

The Subbetic slope became completely destroyed, by folding and uplift, during the Burdigalian. This is thought to be due to collision of the Internal Zone with Iberia (Hermes 1985; Martín Algarra et al. 1988; Sanz de Galdeano & Vera 1992). On the suture, a deep basin (c. 1000 m deep) was formed and filled by sediments of the Espejos formation with detritus from both the Internal Zone (Malaguide and Alpujarride components) and the External Zone. This detritus testifies firstly to erosional unroofing of deeper Betic units and secondly to juxtaposition of the Internal and External Zones in this time. The occurrence of Subbetic oolitic limestone clasts is important because this facies is restricted in the Subbetic to the Sierras de Maria, Maimón, Gigante and Pericay (Figure 1). Large-scale lateral movements at the site of the Vélez Rubio Corridor during Burdigalian to Middle Miocene times, as suggested by Hermes (1984, 1985), are therefore unlikely. The idea, expressed by Martín-Martín (1996) and Guerrera et al. (1997), that their El Niño formation (our Espejos formation) predates the collision of the Internal and External Zones, is not supported by the detritus in this hardly tectonized formation.

Latest Burdigalian to Middle Miocene

Sedimentation in the Espejos basin was interrupted in the late Burdigalian by southward thrusting of the Subbetic. The southward thrust is sealed by latest Burdigalian to Langhian sediments, deposited in a new suite of basins in the areas of Cañada del Maiz, Campico de Flores-Pantano of Lorca, and Bernabeles. These basins show rapid shoaling in the Serravallian. In the Pantano of Lorca area, the moving-to-the-right of the source of a southward outbuilding delta indicates uplift and the presence of laterally moving blocks to the north of this area. This suggests active strike-slip faulting during the Serravallian.

Late Miocene

The genesis and evolution of the Late Miocene Lorca Basin and comparable basins elsewhere is generally attributed to Tortonian strike-slip faulting (e.g. Montenat et al. 1987).

Internal-External Zone Boundary, review of interpretations

The importance of the Internal-External Zone Boundary in the Vélez Rubio Corridor and Espuña area was also realized by British workers (Lonergan et al. 1994). Unfortunately, they assumed that this boundary is a tangible surface. They stated that there are several interpretations of this boundary (dextral strike-slip fault or low-angle northward or southward thrust) but that very little kinematic or structural data had been collected from the fault system to test these interpretations. There is an obvious reason why there are few kinematic data: the Internal-External Zone Boundary is hidden below Subbetic rocks and sealing sediments (Figure 9). Lonergan et al. (1994) considered the conspicuous late Burdigalian southward thrust, that postdates the suturing (Geel 1973 and references therein), to be the boundary and confirmed by kinematic data that this contact is a gently north-dipping thrust zone. Furthermore, Lonergan et al. (1994) did not differentiate the 'Oligo-Miocene' sediments of the Vélez Rubio Corridor. They thought that 'the sequence as a whole can probably be correlated broadly with sediments of the Pliego Valley [the latter valley being the Espuña area]'. They considered these sediments to represent a continuous Oligocene to Langhian Malaguide section. Due to this approach, they seem to have failed to notice the thrust contact between the allochthonous Aquitanian Solana formation and the authochthonous Burdigalian Espejos formation. They also overlooked the Upper Oligocene to lower Aquitanian Ciudad Granada formation and its tectonic position, and thus underestimated the significance of the angular unconformity below the Espejos formation. Their idea that there was continuous sedimentation from the Oligocene through to the Langhian in the Malaguide realm after Eocene emplacement, and that deposition occurred in one gradually deepening Oligo-Miocene foreland basin adjacent to a deforming Internal Zone orogen, is therefore at variance with the tectono-stratigraphic record.

Our results invalidate equally the interpretations of Hermes (1984, 1985), who regarded the Vélez Rubio Corridor as a major strike-slip zone, operative during the Burdigalian to Serravallian, along which whole mountain ranges were transported eastward (from Gibraltar) to their present position in the Corridor.

Though many of the conclusions of Martín-Martín (1996) accord well with our results, some of his ideas are not substantiated by our findings. Contrary to his views, our results suggest a) that the Middle Oligocene to Aquitanian sediments of the Bosque, Ciudad Granada and Pliego formations were deposited in different successive basins, b) that thrusting in the Malaguides started during the Aquitanian, and c) that the Burdigalian Espejos (El Niño) formation postdates the collision of the Internal and External Zones.

Acknowledgements

We made use of the unpublished field reports, kept in archives of the Vrije Universiteit and prepared by P. van Rooyen, 1967, B. van Helden, 1967, C.G. van Houten, 1967, L. Dekker, 1968, J. Bresser, 1969, M. Snijdoodt, 1972, R. Wittink, 1972, St. de Clercq, 1973, T. Romein, 1974, L.P.A. Geerlings, 1978, A.J. Speelman, 1983, W. Meyboom, 1985, and T. Steens, 1986. We wish to thank all these former undergraduates, regrettably Roel Wittink and Tineke Steens posthumously, for their efforts in the field and for the use of their data. Frans Bianchi is thanked for the fruitful discussions and pleasant cooperation in the field. Jan Manuputty and Anne Fortuin are thanked for the determination of planktonic foraminifera. The constructive critical comments by Cees Biermann are gratefully acknowledged. We also wish to thank O.J. Simon for his critical review of the paper. This is contribution no. 980602 of the Netherlands Research School of Sedimentary Geology (NSG).

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