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The Internal–External Zone Boundary in the eastern Betic Cordillera, SE Spain

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Abstract—The Internal–External Zone Boundary (IEZB) in the Betic Cordillera of southern Spain separates the Internal Zone, which was deformed and in part metamorphosed before the early Miocene, from the External Zone, which consists of the cover rocks of the Iberian margin shortened to form a thin-skinned fold and thrust belt in early to middle Miocene times. Over much of its length the IEZB is roughly linear, trending approximately 070°. It has been referred to as a major dextral strike-slip zone, and has even been considered as the northern boundary of a westward moving Alboran microplate.

Field and kinematic data from the eastern Betic Cordillera show that the IEZB crops out over a 60 km distance as a gently-dipping thrust with displacement to the southeast or south-southeast, oblique to its regional trend. There is no evidence of dextral strike-slip movement along the boundary. New micropalaeontological studies of calcareous nannoplankton indicate that the Oligo-Miocene basin along the boundary was the site of continuous deposition until the beginning of the middle Miocene. The thrust at the IEZB cuts early Miocene rocks, and is overlapped by the middle Miocene. Thrusting therefore occurred in this time interval. The IEZB is therefore unlikely to have been the dextral boundary of a westward-moving Alboran microplate at this time, but was more likely to have been the locus of NW-directed dextrally oblique convergence.

INTRODUCTION

THE Betic Cordillera of southern Spain, in common with many other mountain belts in the Alpine system, is divided into Internal and External Zones. The boundary between them is a fundamental tectonic contact that extends the entire length of the Cordillera. Palaeozoic, Mesozoic and locally Tertiary rocks of the Internal Zone were all affected by Alpine events before the early Miocene, and they show a complex and varied metamorphic history with grades including glaucophane schist-; eclogite-, greenschist- and amphibolite-facies in the lower tectonic complexes (Nevado-Filábride and Alpujárride) to anchizone or less in the uppermost (Malaguide) Complex (Egeler & Simon 1969, Torres-Roldán 1979). The External Zone, by contrast, is an essentially unmetamorphosed thin-skinned fold and thrust belt, formed by shortening of the Mesozoic and Tertiary cover of the original south Iberian margin (García-Hernandez *et al.* 1980). Deformation began in the southern, more basinal, part of the External Zone (Subbetic) in the Burdigalian; and continued into the late Miocene in the northern, shelf facies, part of the thrust belt (Prebetic) (Hermes 1978, Banks & Warburton 1991). The underlying Variscan basement is apparently not involved in the deformation.

In the eastern Betic Cordillera the Internal–External Zone Boundary (IEZB) is a major topographic lineament striking ENE that can be recognized from the north side of the Sierra Espuña in the east, through the Velez Rubio corridor to the Sierra Arana, east of Granada (Fig. 1). East of the Sierra Espuña the contact disappears under the late Miocene and younger sediments of the Fortuna Basin, and may be cut by the Crevillente fault (Fig. 1), a Neogene fault with a postulated large dextral strike-slip displacement (De Smet 1984a,b). In the Sierra Espuña–Velez Rubio region the IEZB juxtaposes rocks of the uppermost, non-metamorphic, Malaguide complex of the Internal Zone against Subbetic rocks of the External Zone. These two regions show significantly different stratigraphic, palaeogeographic and tectonic characteristics. The purpose of this paper is to present structural and biostratigraphic data from the Sierra Espuña and the Velez Rubio corridor constraining the character and timing of deformation along this critical boundary.

RECENT IDEAS ON THE NATURE OF THE IEZB

Most workers agree that the IEZB forms the northern boundary of an Alboran domain or microplate, comprising the Internal Zones of the Betic Cordillera and the Rif mountains of Morocco, together with much of what now underlies the Alboran Sea between Iberia and North

Africa. Many believe that the main ENE-trending section of the IEZB has functioned essentially as a major dextral strike-slip fault zone (e.g. Paquet 1974, Hermes 1978, Durand-Delga 1980, Sanz de Galdeano 1983, Leblanc & Oliver 1984, Bouillin *et al.* 1986, Martín-Algarra 1987). According to Leblanc & Oliver (1984), for example, the westward motion of the Alboran domain produced transpressive dextral motion along the IEZB in Spain and sinistral motion along the equivalent boundary (the Jebha Fault) in Morocco. De Smet (1984a,b) postulated that the Internal Zone obliquely approached and made contact with the External Zone along the IEZB, but that further movement was accommodated within the External Zone by strike-slip faulting along the Crevillente Fault, generating a compressive flower structure. Frizon de Lamotte *et al.* (1989) believe that the IEZB forms a major lateral branch line where westward-directed thrusts in the Internal Zone are turned into a steep orientation and cut off against the roof-thrust of the Subbetic system. Hence they also view this zone as a dextral strike-slip fault, but formed as a lateral structure in a westward-moving thrust system.

Most of the authors cited above do not estimate the amount of strike-slip displacement along the IEZB. The palaeogeographic reconstructions in Bouillin *et al.* (1986) require displacements of several hundred kilometers along the IEZB to bring the Internal Zone into its present position. Sanz de Galdeano (1983) stated that there must be approximately 300 km translation along the zone.

An alternative view of the IEZB is suggested by relationships in the Sierra Arana, where García-Dueñas & Navarro-Vilá (1976) show that Alpujárride and Malaguide rocks of the Internal Zone lie above Subbetic limestones of the External Zone along a low-angle thrust, a relationship subsequently modified by SE-directed folding and imbrication. Banks & Warburton (1991) interpret the boundary in the eastern Betic Cordillera as a SE-directed backthrust, but they do not attribute any great significance to it, because they regard the Malaguide as a continuation of the Subbetic sequence to the north. In the westernmost part of the Betic Cordillera, the Internal Zone is generally interpreted to overlie the Subbetic, but Balanyá & García-Dueñas (1987) suggest that the boundary has been reactivated in this region as a low-angle normal fault.

There is no consensus about the timing of movements along the IEZB. Paquet (1974) envisaged displacement of the Internal Zone along the boundary from ENE to WSW during the Eocene. In his opinion the late Eocene and younger rocks in the Sierra Espuña area (his '*zone limite*') suture the boundary. Durand-Delga (1980) believed that motion on the IEZB was complete by Burdigalian (early Miocene) time, as in the eastern Betic

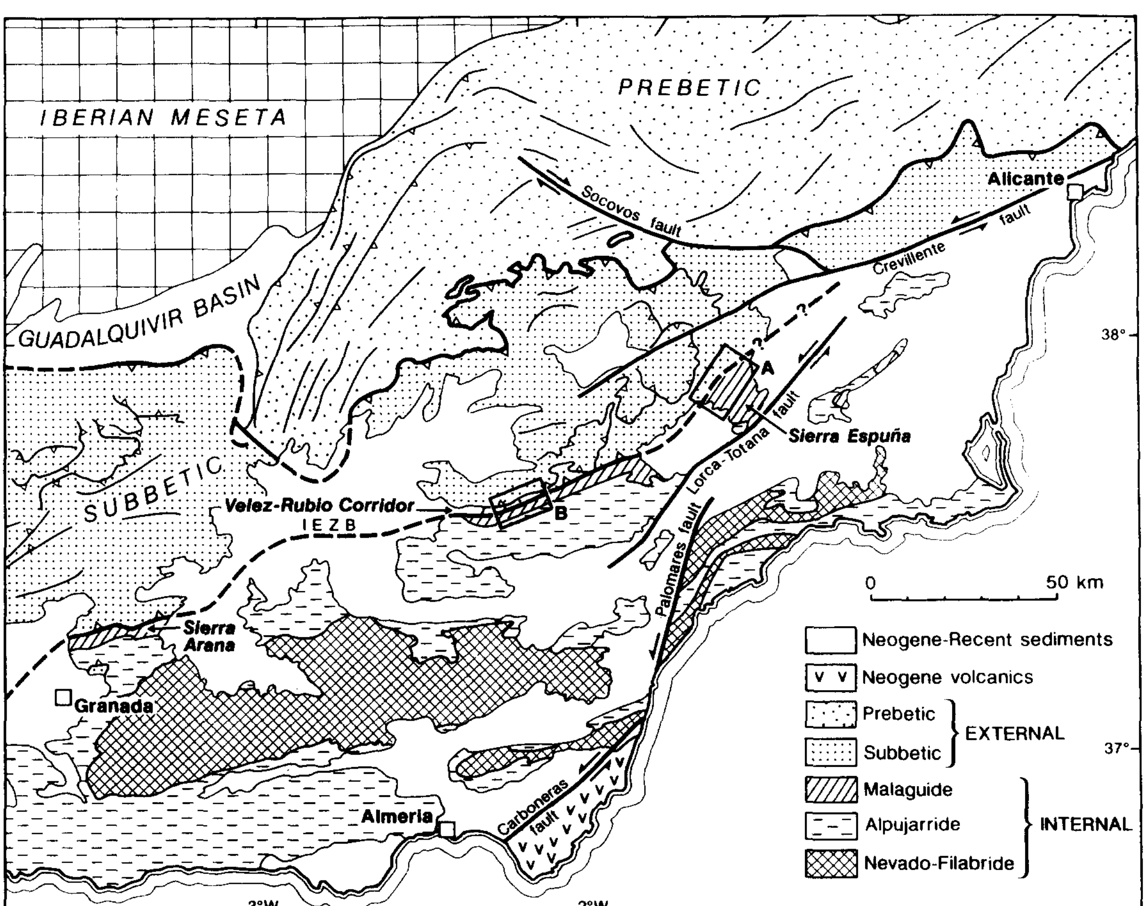

Fig. 1. Geological map of the eastern Betic Cordillera showing the location of the Internal–External Zone Boundary (IEZB). Box A is study area on the north side of the Sierra Espuña and is location of Figs. 3 and 6. Box B is study area in the Velez Rubio corridor and location of Fig. 9.

Cordillera (Sierra Espuña area) Burdigalian rocks are transgressive over the contact. According to Leblanc & Oliver (1984), the IEZB was active from the Eocene–Oligocene to the lower Miocene. In the upper Miocene the boundary of the Alboran Block shifted north to the Crevillente Fault. Leblanc (1990) modifies this theory and states that the Alboran block only moved independently between Europe and Africa until the early Burdigalian. After that time the IEZB was sealed and further movements in the region were transferred onto the sinistral Alhama–Carboneras–Nekkor Line (Fig. 1): the 'trans-Alboran lithospheric shear zone' in the sense of de Larouzière *et al.* (1988). Martín-Algarra *et al.* (1988) state that stratigraphic relations with Miocene sediments indicate that the movements on steep strike-slip structures on the north side of the Sierra Arana are of Burdigalian and Langhian (middle Miocene) age. Sanz de Galdeano (1990) also believes that the contact was sutured by Langhian times.

Interpretations of the IEZB as a major dextral strike-slip zone are largely based on stratigraphic or palaeogeographic relationships. Very little kinematic or structural data has been collected from the faults to test these interpretations. In this paper we present field and kinematic data which indicate that the boundary in the eastern Betic Cordillera is a gently-dipping thrust zone, active in early to middle Miocene time, with transport directions to the southeast or south-southeast, oblique to the general strike of the contact.

STRATIGRAPHIC CONTRASTS ACROSS THE IEZB

The IEZB separates domains with significantly different stratigraphic and palaeogeographic histories (Fig. 2). In the eastern Betics the oldest known rocks of the Malaguide complex are Silurian to Carboniferous green–grey conglomerate, greywacke and interbedded shale and limestone of the Piar Formation, which have been well described from the Velez Rubio corridor by Geel (1973). Palaeozoic rocks of the Malaguide occur locally within the Sierra Espuña at the base of thrust slices. These are overlain by continental red beds, evaporites and dolostone of the Permo-Triassic.

By far the best development of the Jurassic to Tertiary rocks of the Malaguide Complex in the Betic Cordillera is exposed in the Sierra Espuña. There, Liassic dolostone passes up into a thick sequence of Middle and Upper Jurassic limestone, which is dominantly oolitic. The Cretaceous is more varied and in places is missing or incomplete. Locally, it contains sandy glauconitic horizons. The top of the Cretaceous consists of fine-grained limestone with silica nodules. The Palaeocene is missing and the Cretaceous is directly overlain by lower Eocene limestone with *Alveolina*, followed by an Eocene shallow-shelf nummulitic limestone sequence with sporadic clastic influxes derived from the Malaguide Palaeozoic basement. The uppermost Eocene and the Oligocene locally lie unconformably on older rocks down to the Jurassic, and this unconformity marks a fossilized thrust front formed at this time (Lonergan 1993). A thick sequence of Oligocene conglomerate then passes up into lower Miocene basinal deposits. The Miocene rocks are particularly distinctive because of the presence of metamorphic clasts, derived from the metamorphic complexes of the Internal Zone.

In the Velez Rubio corridor the Mesozoic to Eocene Malaguide rocks have similar lithofacies as in the Sierra Espuña, but are considerably reduced in thickness. The Oligo-Miocene section is disrupted by faulting, but in general the rocks are also of a basinal facies with detritus of Internal Zone provenance (Geel 1973).

The Mesozoic and Tertiary stratigraphy of the adjacent Subbetic is very different to that of the Malaguide (Paquet 1969, Hermes 1978). In the Subbetic no rocks older than the Triassic are ever exposed. Continental red beds, shallow marine dolostone and evaporites of a 'Germanic'-type lithofacies make up the Triassic sequence (Hermes 1978, García-Hernandez *et al.* 1980, Fontboté & Vera 1983). Shallow-water carbonates were widespread in the Early Jurassic and are now exposed as dolostone. In the central parts of the Subbetic a marly facies with pillow lavas was deposited in the Middle and Late Jurassic, but in the southern Subbetic shallow-water limestone was deposited on block highs.

The Cretaceous of the Subbetic is very different to that of the Malaguide, being dominated by thick (up to 800 m) sequences of marly limestone with pelagic faunas. The *couches rouges* facies pink marls are distinctive in the Upper Cretaceous and in some sections they extend across the Cretaceous–Tertiary boundary. The Palaeogene was again dominated by pelagic marl, in contrast to the shallow-water fossiliferous limestone of the Malaguide sequence. In the Oligo-Miocene, minor clastic detritus first appeared in the form of carbonate conglomerates (Instituto Geologico y Minero de España 1:50,000 Coy sheet). Marl and calcarenitic turbidites became widespread in the Burdigalian and Langhian. The Tertiary rocks of the Subbetic contain no siliciclastic detritus, however, in contrast to those of similar age in the Malaguide. The first sedimentary links between the Malaguide Complex and the Subbetic are observed on the north side of the Sierra Espuña where Langhian rocks at the top of the Malaguide sequence onlap onto Subbetic Aquitanian rocks.

OLIGO-MIOCENE SEDIMENTS ALONG THE IEZB

The Oligocene and Miocene rocks along the IEZB are particularly significant in that they span the period when the Internal and External Zones were juxtaposed, and there has been considerable disagreement about their stratigraphic relations.

Oligo-Miocene rocks of the Malaguide are exposed along the IEZB on the north side of the Sierra Espuña, in the Pliego valley. According to Durand Delga (1980, personal communication 1990), Mäkel (1985) and

Martín-Algarra (1987) there is biostratigraphic evidence for a significant stratigraphic break within the sequence, separating the red–brown Amalaya Formation (Oligocene–Aquitanian) from the overlying grey–green Bernabeles Formation (Fig. 3). If this is true, there may have been a phase of motion between the Internal and External Zones that has been masked by the subsequent deposition of the Bernabeles Formation. Hence it is important to re-examine the stratigraphic relationships of the Oligo-Miocene rocks and to date the sequences accurately.

The Amalaya Formation is made up of red, yellow and grey marl interbedded with coarse sandstone and polymict conglomerate. Sedimentary structures in the sandstone and conglomerate beds indicate that they have been deposited by mass gravity flows. Planktonic faunas are present in the marls. These features suggest that the rocks of the Amalaya Formation were deposited in a basinal environment.

Many aspects of the overlying Bernabeles Formation are similar. It consists of grey graded sandstone, siltstone and polymict conglomerate interbedded with grey marl containing planktonic faunas. Some coarser calcarenitic horizons, about 20 m thick, can be traced laterally for several kms (Fig. 4). A large slump crops out on the Bullas road in a 30–40 m wide zone and it can be traced along strike for 0.5 km (Fig. 4). It contains evidence for intraformational deformation: folded and bedded sequences are arranged in a chaotic fashion, and there is a block of Malaguide Permo-Triassic purple shale 1 m in diameter. Based on present locations of the Malaguide red beds in the Sierra Espuña and the size of the Tertiary

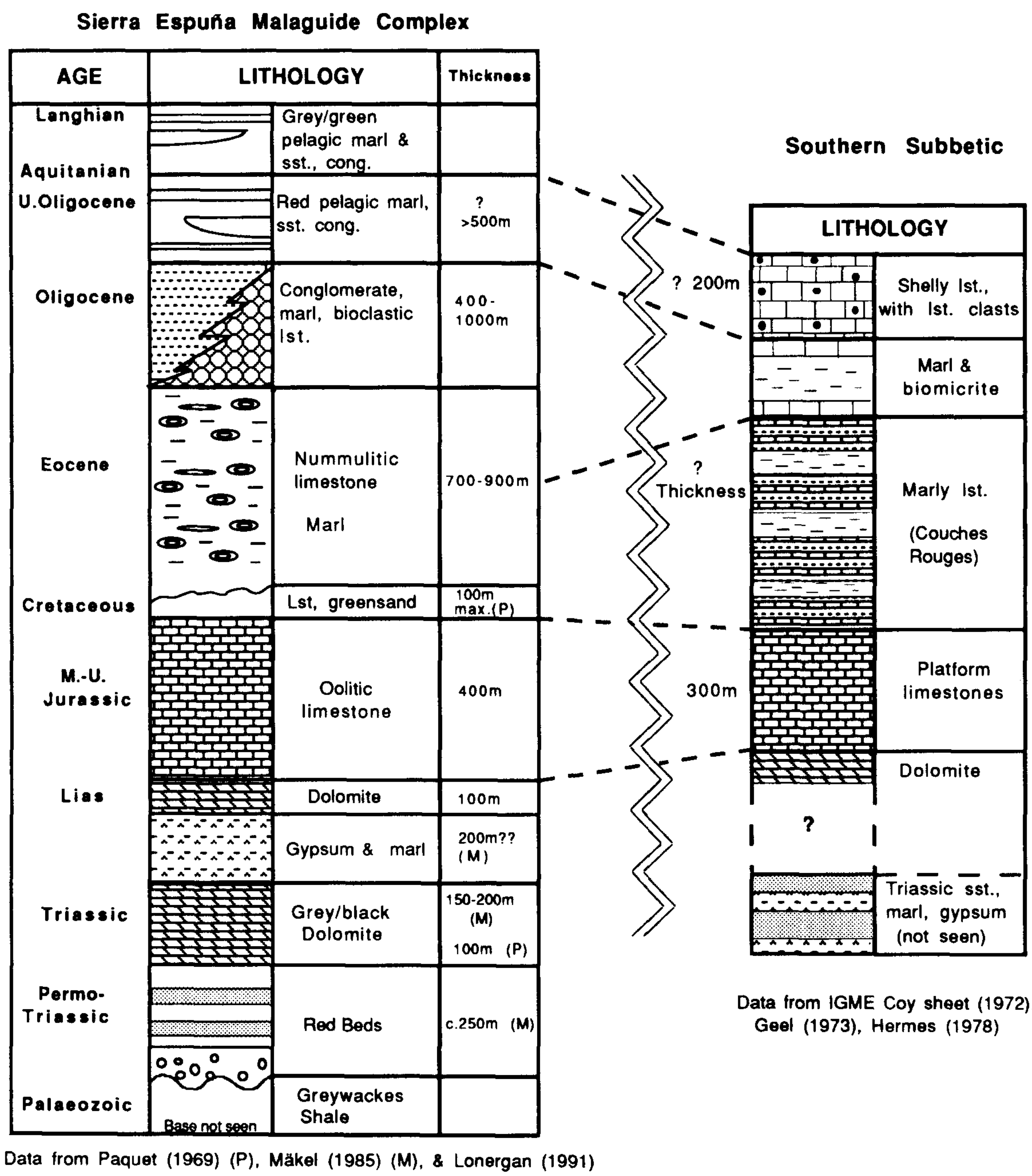

Fig. 2. Schematic stratigraphic columns for the Subbetic to the north of the IEZB and the Malaguide complex to the south.

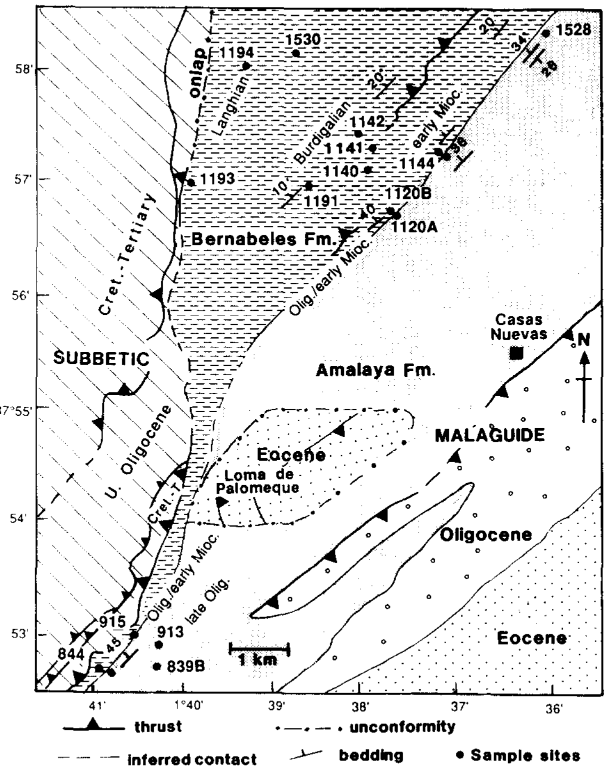

Fig. 3. Nannoplankton sampling sites located on a simplified geological map showing the mapped contact between the Bernabeles and Amalaya Formations in the Pliego valley.

basin, blocks like this must have travelled distances on the order of tens of kilometres.

The clastic detritus within both formations is derived from the Internal Zone, with an increase in metamorphic clasts up-section in the Bernabeles Formation (Lonergan 1991). The sedimentary features described above suggest a similar depositional environment for the two formations. The contact as mapped on the ground follows the strike of bedding within the two formations (Fig. 3). The exposure is poor because much of the Pliego valley is covered with fields, but dry stream sections allow the contact to be traced. In one such stream valley (sample site 1144, Fig. 3), a transitional contact between the two formations is exposed. Varicoloured red and brown marls and sandstones of the Amalaya Formation pass up into the grey marls and sandstones of the Bernabeles Formation without a break. There is no indication of any unconformity. In view of this evidence it seems unlikely that there can be a major stratigraphic interruption between the two formations with uplift, deformation and erosion.

Biostratigraphy of the Miocene sediments

It is now more than 20 years since the original biostratigraphical work was done on the Tertiary sequences of the Pliego valley by Paquet (1969) and since then the stratigraphic resolution of this era has been improved. Paquet (1969, p. 141 and fig. 6.1) describes in detail a section from the Oligocene up to the Subbetic contact in the southwestern end of the Pliego valley. From the stream sections leading up to the exposed thrust at the base of the Subbetic (sites 844 and 915, Fig. 3) Paquet obtained upper Oligocene planktonic foraminifera assemblages from red marl (Amalaya Formation), and Aquitanian assemblages with some reworked Cretaceous components from the green marl of the Bernabeles Formation. He states that the grey marl in the Pliego valley to the east is transgressive on both upper Oligocene red marl of the Malaguide (his '*zone limite*') and on the Subbetic Aquitanian limestone, and he gives an early Miocene age for these sediments based on earlier work by Fallot (1945).

More recently, Martín-Algarra (1987, p. 803) proposed that an upper Aquitanian biozone is missing between the Bernabeles and Amalaya Formations. Planktonic foraminifera from 12 sites from Paquet's original section in the western end of the Pliego valley give upper Burdigalian (zone N8 of Blow 1969) and Langhian (*Glomerosa curva* zone of Molina 1979) ages (Martín-Algarra 1987, p. 825 and fig. 177). This contrasts with work by Rivière *et al.* (1980) who also resampled Paquet's section in the west of the Pliego valley. They obtained late Oligocene–Aquitanian ages from both planktonic foraminifera and nannoplankton (zone NN1) in the red marls and Burdigalian ages (zone NN2/3) from assemblages in the grey–green marls with no evidence for a stratigraphic gap.

Martín-Algarra (1987) also distinguishes two different formations within the Bernabeles Formation. South of Loma de Palomeque, he considers the green marls as belonging to the Burdigalian Grupo Viñuela—a group that he recognizes the length of the Betic Cordillera as transgressive on the Internal Zone. North of Loma de Palomeque, he places the grey marls and calcarenites into a middle Miocene Formation (see Martín-Algarra 1987, fig. 171).

To re-evaluate and reconcile the discrepancies in previous work with the lack of a mappable unconformity in these sediments, we resampled Paquet's original section (sample sites 844 and 915, Fig. 3) and then sampled as close as possible to the contact between the Amalaya and Bernabeles Formations to the northeast (sample sites 1120, 1144 and 1528). Samples were also collected from lower in the Amalaya Formation, and higher up section in the Bernabeles Formation in order to determine the youngest age for these sediments, and hence constrain the timing of deformation. Planktonic foraminifera are sparse and poorly preserved, so nannoplankton assemblages were studied instead. The sample sites are shown on the map in Fig. 3, and the results in Fig. 5. These suggest the following conclusions. (a) Samples from within the Amalaya Formation (sites 839B and 913) contain Late Oligocene or Oligo-Miocene boundary assemblages. (b) Oligo-Miocene boundary assemblages are present in red and grey marls on either side of the length of the mapped contact (sites 1528, 1120, 844, 1144, 1120 and 915) and there is no evidence for missing biozones. (c) The Burdigalian dates for sites 1191, 1140, 1141 and 1142, and the Langhian dates for sites 1193 and 1194 are consistent with stratigraphic younging up-section, and hence there is no biostratigraphical evidence for any unconformities within the Bernabeles Formation.

From the biostratigraphical study, it appears that there has been continuous sedimentation from late Oligocene through to Langhian times in the Tertiary basin. There also must have been considerable convergence between the Internal and External Zones in this same time span because towards its top the Bernabeles Formation onlaps onto Subbetic rocks (Figs. 3 and 4). Prior to the Langhian, there is no record of sedimentary links between the Subbetic and Malaguide domains.

KINEMATICS OF DEFORMATION ON THE IEZB

Sierra Espuña

At the southwest end of the Pliego valley, upper Cretaceous to lower Eocene marly limestones of the Subbetic are in thrust contact with lower Miocene rocks (Bernabeles Formation) of the Internal Zone sequence. The thrust dips gently to the northwest and can be traced around the hillside approximately parallel to the contours for 3 km. A stream section (at A, Fig. 6) exposes a 15 m thick fault zone through the thrust with spectacular fault fabrics developed in the Cretaceous–Tertiary marly limestones. Lenses of limestone are separated by marly gouge zones. Curved shear planes form anastomosing patterns isolating lenses of rock. Identical tectonite fabrics are described by Casas & Sàbat (1987) from thrust zones in Mallorca involving very similar Cretaceous marly limestones. Well developed calcite crystal fibres on gouge surfaces and shear planes indicate the movement direction. The geometry of the fault fabrics clearly suggests a thrust sense towards the southeast.

To the northeast, along the strike of the contact, the thrust is no longer exposed. At locality D, Fig. 6, the contact with lower Miocene Subbetic rocks is steep, but the fault itself is not exposed. Kinematic data collected on minor faults in the Aquitanian Subbetic limestones at the contact indicate antithetic motion to the northwest with a component of oblique slip. Further north, the faulted contact disappears and middle Miocene rocks of the Bernabeles Formation can be seen onlapping clearly on to Subbetic Aquitanian limestones, as is shown on the 1:50,000 geological map (Instituto Geologico y Minero de España 1972) (at E, Fig. 6).

Thrust-related deformation does not die out but is transferred onto calcarenitic horizons within the Bernabeles Formation of the Internal Zone sequence. A series of thrusts in one such horizon is exposed in a quarry at B,

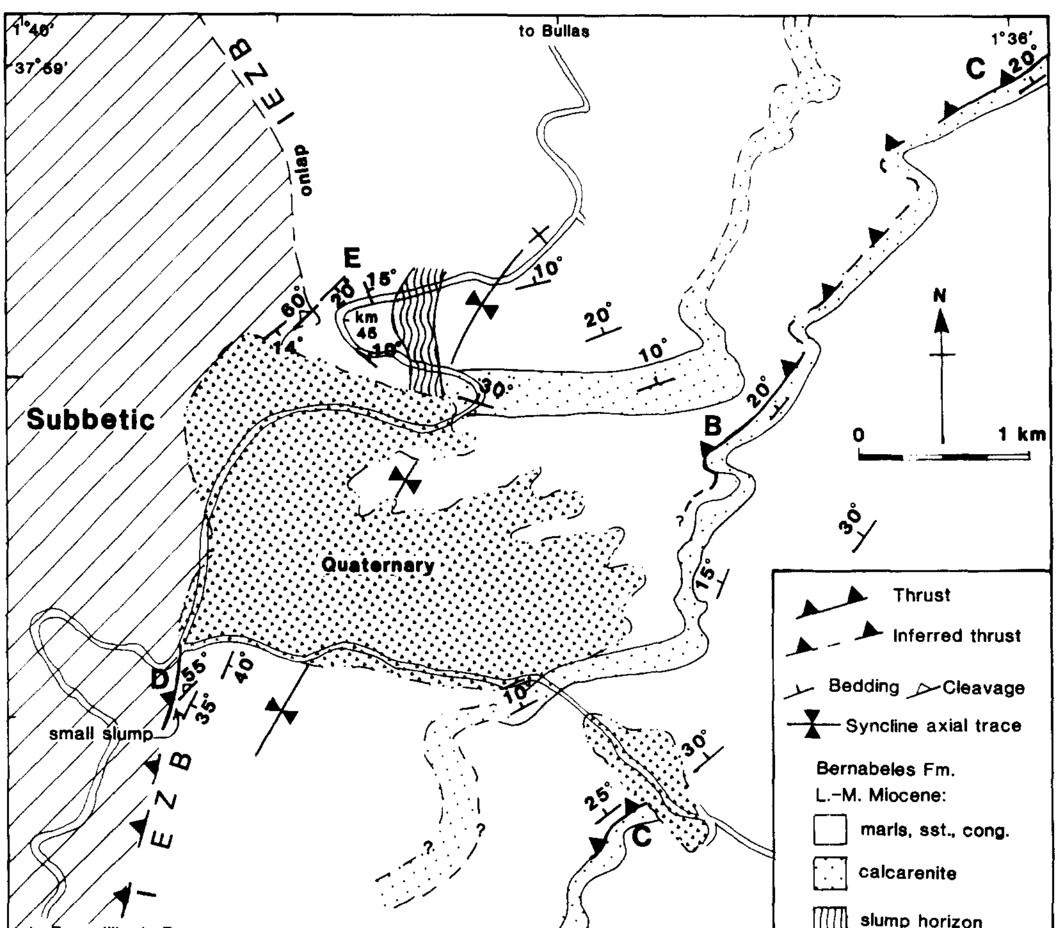

Fig. 4. Geological map of the area where the IEZB between the Subbetic and the Miocene Bernabeles Formation ceases to be a fault contact and the Miocene sediments onlap on to the Subbetic. Letters B–E are the localities where the kinematic data shown in Fig. 6 were collected.

SAMPLE	Coccolithus pelagicus	Cyclicargolithus abisectus	Cyclicargolithus floridanus	Helicosphaera intermedia	Reticulofenestra scissura	Sphenolithus sp.?	Zygrhablithus bijugatus	Discoaster deflandreii	Helicosphaera perch-nielseniae	Reticulofenestra minuta	Sphenolithus conicus	Sphenolithus moriformis	Helicosphaera euphratis	Helicosphaera recta	Helicosphaera truempyi	Pontosphaera enormis	Reticulofenestra scrippsae	Sphenolithus distentus	Pontosphaera multipora	Discoaster variabilis	Helicosphaera carteri	Helicosphaera mediterranea	Braarudosphaera bigelowii	Sphenolithus dissimilis	Helicosphaera ampliaperta	Reticulofenestra perplexa	Retic. pseudoumbilicus/gelidus	Sphenolithus heteromorphus	Sphenolithus belemnos	Calcidiscus macintyreii	Cretaceous sp. (reworked)	Eocene/Oligocene sp. (reworked)	AGE & Fm.
1193C	x	x	x	x	-	-	-	-	x	-	-	-	-	-	-	-	-	-	-	-	x	-	-	-	x	-	-	x	-	-	x	-	
1193A	x	x	-	-	x	-	-	-	-	-	-	x	-	-	-	-	-	-	-	-	x	-	-	-	x	x	-	-	-	x	x	x	**Top Bernabeles Fm.**
1194A	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	-	-	x	-	-	x	-	x	-	-	Langhian
1194	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	-	-	x	-	-	x	-	x	-	-	
1530	x	x	-	-	-	-	-	-	-	-	x	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	-	x	-	-	x	x	Burdigalian/Langhian
1141	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	x	-	-	-	-	x	-	-	x	x	
1140	x	-	x	-	-	-	-	x	-	-	-	-	-	-	-	-	-	-	x	-	x	-	-	-	x	x	x	x	-	-	x	x	
1142	x	-	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	-	x	-	-	x	-	**Bernabeles Fm.**
1191	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	x	x	x	-	-	x	-	Burdigalian
1191B	x	x	x	x	-	-	-	-	-	-	-	-	-	x	-	-	-	-	x	x	x	-	-	-	-	-	-	-	x	-	-	-	
1191C	x	-	x	-	-	-	-	-	x	x	-	-	-	-	-	-	-	-	-	x	-	-	-	-	-	-	-	-	-	-	-	-	
915D3	x	x	x	-	-	-	-	x	-	x	-	-	x	x	-	x	-	-	-	-	-	-	x	x	-	-	-	-	-	-	x	-	
915B3	x	-	x	-	-	-	-	-	-	-	-	-	x	-	-	-	x	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	
844B	x	x	x	-	-	-	-	-	-	x	-	x	x	-	-	-	-	-	x	-	x	-	x	-	-	-	-	-	-	-	-	-	
1120B	x	x	x	-	-	-	x	x	-	x	x	-	x	x	-	x	-	x	-	-	-	-	-	-	-	-	-	-	-	-	x	-	**Amalaya/Bernabeles**
1144C	x	x	x	x	-	-	-	x	-	-	-	-	-	x	-	-	-	-	x	-	x	-	-	-	-	-	-	-	-	-	x	-	**boundary**
1144B	x	x	-	-	-	-	-	-	x	-	x	-	-	-	-	-	-	-	x	x	-	-	-	-	-	-	-	-	-	-	-	x	L. Olig./E. Mioc.
1144A	x	x	x	x	-	-	-	x	x	-	x	-	-	x	-	-	-	-	x	x	x	x	-	-	-	-	-	-	-	-	x	x	
844A	x	x	x	-	x	-	-	x	-	x	-	-	-	-	-	-	x	-	x	-	-	-	-	-	-	-	-	-	-	-	-	-	
1120A	x	x	x	x	x	-	x	x	-	x	x	-	x	x	x	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	x	x	
1528	x	x	x	x	-	-	-	x	x	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	
913	x	x	x	x	-	-	-	x	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	x	-	**Amalaya Fm.**
839B	x	x	x	x	x	x	x	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	L.Olig.
RANGE	Paleog.-Rec.	NP24/Mioc.		?Eoc./NN5	U. Olig.		Eoc./NN1-	Olig./NN4			NN1/3	Eoc./NN9	Eoc./NN4	NP24/NN1	NN1	NP25		NP23/24				NP25/NN5		NP24/NN2	NN2/4			NN4/5	NN3	NN4/NN19			x present - absent

Fig. 5. Nannoplankton species lists for samples from the Amalaya and Bernabeles Formations. For sample locations see map, Fig. 3. Note that samples within each of the stratigraphic divisions indicated were collected along strike from one another and do not represent sequences in stratigraphic order. NP = Nannoplankton Palaeogene; NN = Nannoplankton Neogene; biozones according to Martini (1971) and Okada & Bukry (1980).

Fig. 6. There has been significant bedding-plane slip with local imbrication. Duplexes are developed with horses merging into the roof thrust along the base of a marl horizon (Fig. 7) and planar backthrusts cut the curved imbricate slices. The dominant transport direction was to the southeast with antithetic backthrusts to the northwest. At a mesoscale the rock is criss-crossed by synthetic and antithetic minor faults along which stepped calcite fibres have grown, giving a clear indication of the sense of shear. The kinematic data from this duplex structure are summarized in Fig. 6. Some of the NW-dipping minor faults occur in sets within beds and although crystal fibres have developed along the faults there are no obvious offsets on the beds. The fractures appear to be incipient horses or a spaced fracture cleavage. It is difficult to make any accurate estimate of the amount of shortening being taken up along this calcarenite band. A 50 m long stack of imbricate thrust slices is exposed in the quarry parallel to the transport direction. The calcarenite band is about 6 m thick. Elsewhere along its strike it is no more than 3–4 m thick, so there may have been shortening of up to 40–50% in the imbricate zone. Kinematic data from the same stratigraphic horizon and other similar horizons (C, Fig. 6) indicates the same SE-directed thrust-transport direction and a lesser component of northwest antithetic motion.

At locality E (Fig. 6), exposures in the sides of a dry river gorge clearly show the marl, sandstone and conglomerate sequence of the Bernabeles Formation onlapping onto Subbetic Aquitanian limestones (see the geological map in Fig. 4). The contact dips at less than 20° to the southeast. There is no difficulty in distinguishing the two formations. The Subbetic rocks are shelly conglomeratic limestones with no siliciclastic components, whereas the Bernabeles Formation rocks contain metamorphic and clastic components derived from the Internal Zone. A strong SE-vergent stylolitic cleavage is developed in the Subbetic rocks only. There has been bedding-plane slip in the Bernabeles Formation with well developed calcite fibre lineations on conglomerate and coarse calcarenite beds indicating NW-directed sense of shear. Four hundred metres east of the onlap at the intersection of the dry river bed and the Bullas road (km 45, Fig. 4) individual coarse micaceous sandstone beds are imbricated. Calcite fibre lineations from the imbricate surfaces and antithetic Riedel fractures indicate a thrusting direction to the southeast. The kinematic data from this area are summarized on Fig. 6, E.

Paquet (1969) recognized that the Subbetic was thrust over Miocene rocks of his '*zone limite*' (see Paquet 1969, fig. 102, p. 237 and plate 1) but he interpreted the thrusting as being part of the Miocene deformation of the Subbetic, and suggested that the fundamental contact between the Malaguide rocks and the Subbetic is hidden by the '*zone limite*' sediments. If there were strike-slip displacements between the Malaguide and Subbetic domains prior to the Oligo-Miocene one would expect to find faults exhibiting strike-slip kinematics in the older Tertiary rocks on the north side of the Sierra Espuña. However only one strike-slip fault of any significance was found, exposed in the middle of the Pliego valley in Oligocene marls, in a road cut along the Casa Nuevas road, 2 km southwest of Casa Nuevas (Grid Ref: 06223 41970 UTM; 7827 3704 L). Shear surfaces with fibre lineations in the fault zone indicate that it has been reactivated a number of times. The clearest, biggest surfaces are a set of NNE-trending sinistral faults (Fig. 8a). This sinistral trend is also observed in Neogene rocks over large areas of the southeastern Betics (e.g. de Larouzière *et al.* 1988); hence, these are probably late structures post-dating the contact of the Internal Zone

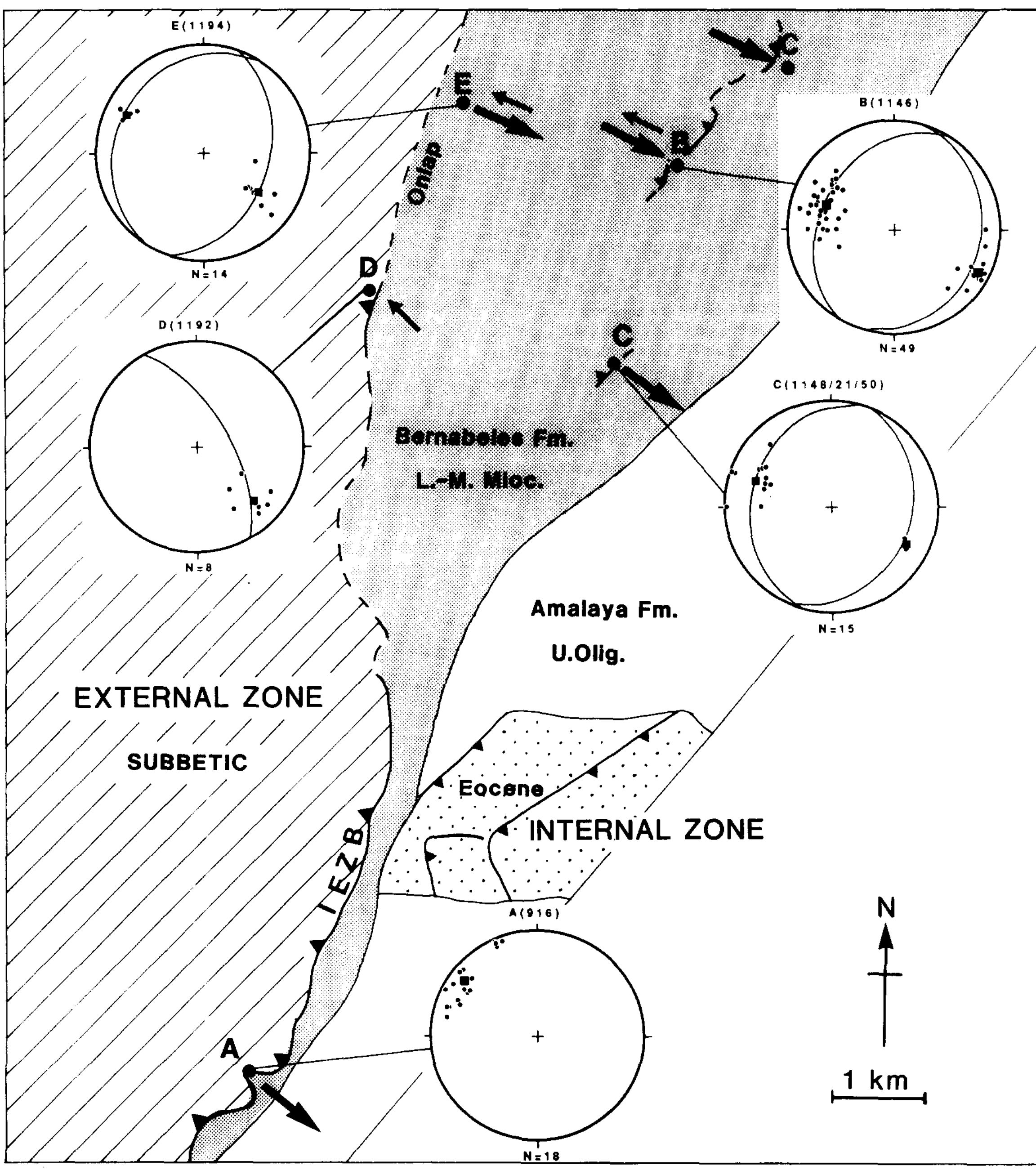

Fig. 6. Kinematic data from the IEZB on the north side of the Sierra Espuña, in the Pliego valley. Crystal fibre lineations are shown on stereographic projections. The mean of the data cluster is shown as a black square. The mean of the fault planes on which lineations were measured is shown as a great circle. Big arrow is the main thrust transport azimuth and the small arrow is the antithetic (backthrust) azimuth. For location of map see Fig. 1.

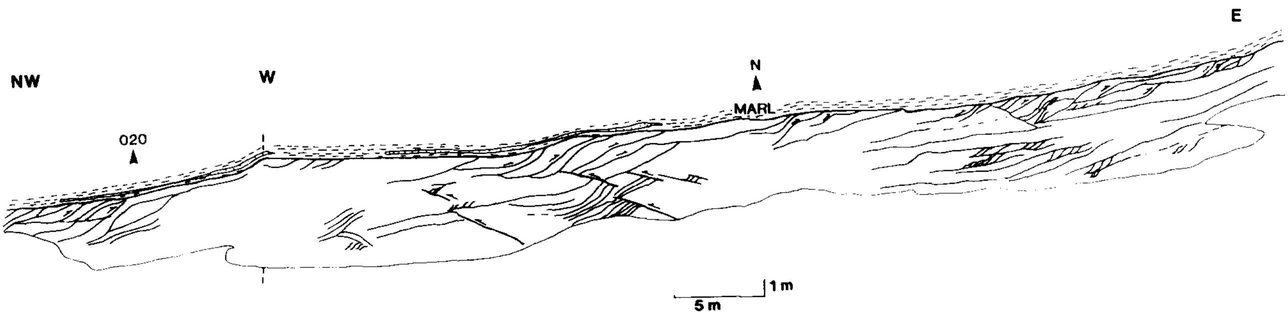

Fig. 7. Line drawing of a quarry face in the Bernabeles Formation where the data of locality B (Fig. 6) were collected. The quarry is in a calcarenite horizon within a marly sequence. There has been significant bedding-plane slip with local imbrication. Duplexes are developed with horses merging into the roof thrust along the base of a marl horizon. Planar backthrusts cut the curved imbricate slices. The dominant thrusting was to the southeast with antithetic backthrust motion to the northwest.

with the External Zone. A set of NE-trending dextral surfaces are also present (Fig. 8b). The lineations suggest an oblique extensional component to the movement. A third set of surfaces (Fig. 8c) may indicate a component of thrusting to the southeast, based on two sense of shear indicators, which is in accordance with the rest of the data from the IEZB presented above.

The dominant structures of the IEZB in the Sierra Espuña area are consistent with SE-directed thrusting along this contact, and there is very little evidence for NE–SW-trending dextral motion.

Velez Rubio corridor

The Velez-Rubio corridor (Figs. 1 and 9) is a pronounced ENE-trending linear morphological feature extending from Fuensanta in the east, where it disappears beneath Tortonian and younger sediments of the Lorca Basin, to Chirivel in the west, where it is masked

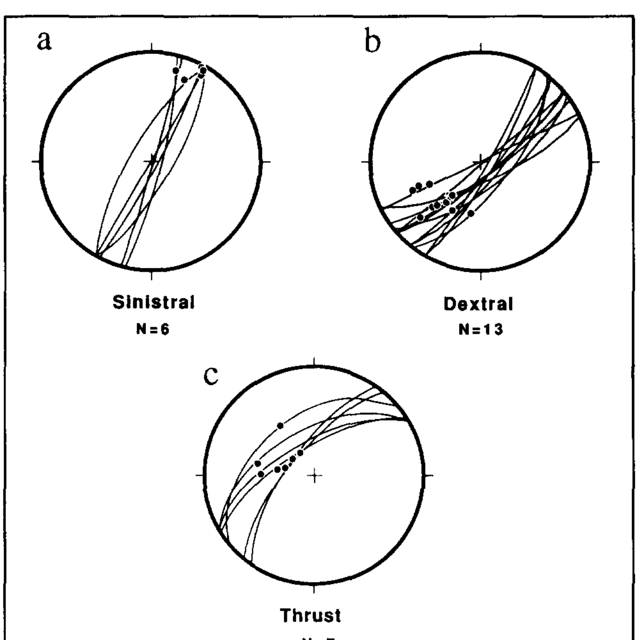

Fig. 8. Kinematic data from a fault zone exposed in Oligocene rocks 2 km southwest of Casas Nuevas. The fault zone has been reactivated at least three times. The dominant fracture surfaces form a NNE-trending sinistral set (a). A dextral extensional set trending NE–SW also occurs (b) and there is also a possible SE-directed thrust set (c).

by Pliocene and Quaternary fill of the Baza basin. The linearity of this zone is probably one of the factors that stimulated speculation about a possible strike-slip character for the IEZB. In fact, the corridor is occupied by several structural features with differing kinematic significances.

(1) The northern boundary of the corridor is marked by a major S- to SE-directed backthrust zone that places massive Jurassic limestone of the internal Subbetic over Cretaceous to early Miocene pink, white and green pelagic marl and limestone (Taibena Formation, Geel 1973), which have been transported in turn over Oligo-Miocene sediments (Fig. 10). Both the outcrop pattern and measured slip surfaces (Fig. 11) in the thrust zone suggest that it dips gently north. The Cretaceous to early Miocene marl–limestone sequence has taken up a substantial part of the displacement in this thrust zone, and is intensely deformed, with spectacularly developed gouge fabrics and shear bands marked by calcite fibres. Kinematic data are abundant and for the most part clearly indicate SSE- to SE-directed motion on gently N- to NW-dipping surfaces (Figs. 11d, f, h & j). Some areas show evidence for dextral motion on W- to WNW-trending surfaces (Figs. 11b & e) or sinistral strike-slip on N- to ENE-trending surfaces (Figs. 11c & j). Location *b* lies on a WNW-trending fault that dextrally offsets the thrust zone by about 500 m (Fig. 9), and clearly post-dates it. Several such dextral faults can be distinguished in the area of Fig. 10.

(2) The central part of the corridor is occupied by ?Oligocene to middle Miocene sediments, distinguished as the Solana and Espejos formations by Geel (1973). These comprise grey, green, and white marls and siltstones with interbedded turbiditic sandstones and conglomerates. They are intensely folded and faulted, and are poorly exposed, and we have not distinguished the two formations: they are shown as 'Intermediate Zone' on the map and cross-section. The clastic beds are rich in detritus derived from the Malaguide, particularly from the Palaeozoic Piar formation (Geel 1973). In places, the early Miocene Espejos Formation lies in stratigraphic contact with Eocene rocks of the Malaguide along the southern margin of the outcrop belt (e.g. just south of the road 1–2 km west of Velez Rubio). The sequence as a whole can probably be correlated broadly

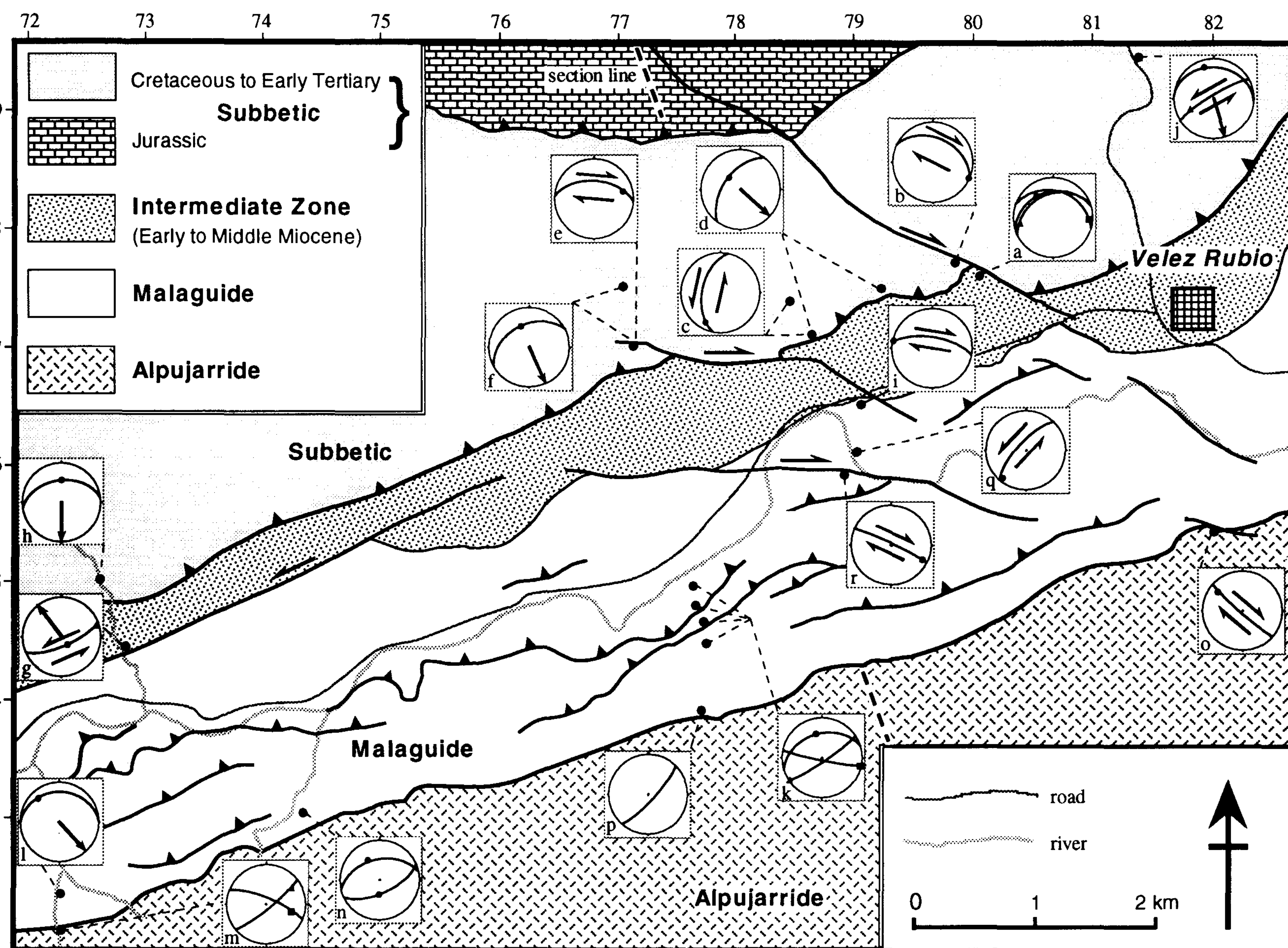

Fig. 9. Map of the Velez Rubio corridor (based in part on Geel 1973), with summary kinematic data. The mean fault plane (great circle), mean lineation (point symbol) and sense of shear are shown from individual locations.

with the Amalaya and Bernabeles Formations of the Pliego valley, described above. They were presumably deposited in a trough along the northern margin of the Internal Zone before or during the start of deformation in the Subbetic.

Kinematic data are difficult to collect in the clayey marls of the Miocene sequence. An intensive study in a moderately well-exposed section near the northern boundary of the Miocene tract (location *a* in Fig. 9, 1.5 km west of Velez Rubio) shows that the dominant fracture surfaces dip moderately north. Striations on these surfaces indicate dextral, sinistral and reverse senses of motion (Fig. 11a), and because of this overprinting, the slip sense on many surfaces could not be identified. The importance of dextral slip at this locality is clearly related to the nearby cross-cutting dextral fault, described above.

The southern boundary of the Miocene outcrop belt in the western part of Fig. 9 is an ENE-trending steep fault zone about 100 m wide occupied by slices of Malaguide Palaeozoic, Triassic, Jurassic and Eocene rocks, as well as Miocene. Kinematic data from the fault zone indicate both reverse and sinistral strike slip motion for this fault zone (Figs. 8c and 11g). The fault can be traced for nearly 3 km, but disappears eastwards within poorly exposed Miocene. In view of its width, it is likely to have a significant displacement, and is probably at least partly responsible for the linear character of the corridor.

(3) The Malaguide rocks in the Velez Rubio region form a narrow strip 1–2 km wide along the southern margin of the corridor, contributing to its distinctive character. The internal structure of the Malaguide is complicated. Folds in the Palaeozoic Piar Formation are upright but downward facing over a large area (Borradaile 1976); which does not appear to be true for the younger formations. This may indicate a Variscan phase of folding. Palaeozoic and Triassic rock units have been imbricated together by Alpine thrusting. The Jurassic and younger carbonate rocks have been detached on the Trias and folded and imbricated independently of the underlying rocks. The Eocene nummulitic limestone, although folded, appears to lie stratigraphically on all underlying units down to the Trias, so some of the Alpine deformation may have occurred before the end of the Eocene, as in the Sierra Espuña. The overall dip of the Malaguide rocks is moderately to steeply north. The contact with the underlying Alpujárride Complex, which is regionally subhorizontal within the Internal Zone as a whole, here dips very steeply north. The narrowness and linearity of the Malaguide belt is at least partly the result of this steep dip, and the structural thickness of the Malaguide Complex is probably over 1000 m (Fig. 10). Towards Fuensanta in the east, the dip of the basal contact decreases, and the outcrop width of the Malaguide correspondingly increases.

In some areas within the Malaguide, the orientations of fracture surfaces and lineations are so variable as to be uninterpretable, and this is likely to reflect the superposition of several phases of deformation. Kinematic data from some zones of fairly clearcut young faulting, however, show patterns consistent with those discussed above for the central part of the corridor. These include conjugate sets of reverse faults, slipping south on moderately N-dipping surfaces and north on steeply S-dipping surfaces (Figs. 11k, l & n); and apparently conjugate sets of sinistral and dextral strike-slip faults (Fig. 11i, k, m, o, q & r). Dextral faulting at locations *r* and *o* is clearly related to mappable WNW-trending faults that dextrally offset geological boundaries by distances of 100–1000 m (Fig. 9).

(4) The Malaguide–Alpujarride boundary is itself an important tectonic lineament that runs parallel to the Velez Rubio corridor for about 25 km. Over most of this distance it has a dip of 60–90°N, flattening to horizontal at the eastern end of the corridor near Fuensanta. It is marked by a nearly continuous layer of calc–mylonite 1–100 m wide derived from Alpujarride carbonate rocks, and on the north side of the contact by a layer of brittle fault gouge 10 m or more thick, derived from sediments of the Piar formation. The mylonite has a strong regionally subhorizontal stretching lineation, and ductile sense-of-shear indicators give a clear dextral sense. Kinematic data from the gouge zone show evidence for both sinistral and reverse slip parallel to the boundary, as well as dextral slip on cross-cutting WNW-trending surfaces (Figs. 11m, n & o). In one location the slip data are so variable as to be uninterpretable (Fig. 11p). It seems likely that this major mechanical boundary has been reactivated in each of the several events that have affected the Velez Rubio corridor.

The present orientation of the mylonite zone along the Malaguide–Alpujárride contact is a result of later steepening, and in its original orientation it was probably a gently-dipping surface with a top-to-the-northeast

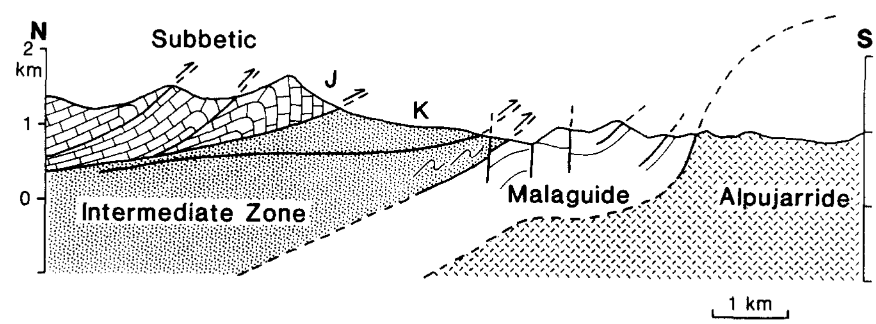

Fig. 10. Schematic section across the Velez Rubio corridor, to show the relationships among the major tectonic units. Section line is located on Fig. 9, but is extended a short distance to the north and south of the map.

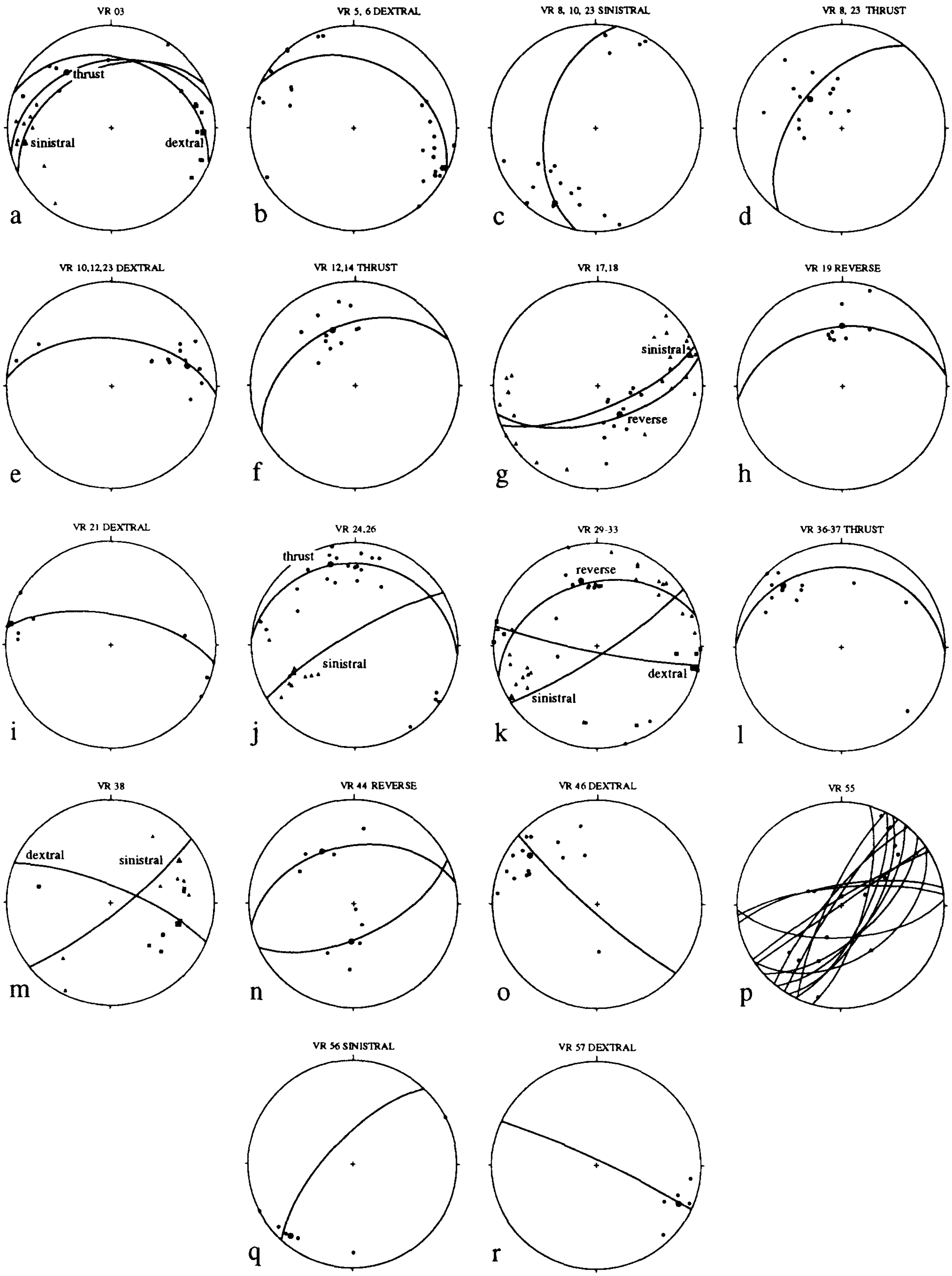

Fig. 11. Equal-area plots of kinematic data from the Velez Rubio corridor. Mean fault planes are shown as great circles, lineations as point symbols, with the mean lineation as a larger symbol. Plots *a–r* are keyed to the map (Fig. 9), and are referred to in the text. VR numbers at the top of each plot refer to locality numbers in J. P. Platt's notebook. Some data sets were collated from several adjacent localities.

sense of shear. It can be linked to a pervasive deformation within the Alpujárride Complex throughout the Sierra de las Estancias, and is an important element in the tectonic evolution of the Internal Zone. This deformation and its significance will be discussed elsewhere: it is not directly pertinent to the Velez Rubio corridor and the IEZB. The steepening of the contact, and of the Malaguide complex as a whole, into parallelism with the corridor, is pertinent, however. This steep zone may be analogous to the steep limb of the major N-vergent fold in the Sierra Espuña described by Lonergan (1993), and like that fold may be a compressional feature related to shortening across the IEZB. In any event, the present outcrop location of the IEZB, and the linearity of the Velez Rubio corridor, appear to be largely a result of the existence of this steep zone.

TIMING OF DEFORMATION ALONG THE IEZB

In the Sierra Espuña SE-directed thrusting affects the Langhian parts of the Bernabeles Formation (E, Fig. 6). Regionally, upper Miocene (Tortonian) rocks are unconformable on deformed Internal Zone rocks. Hence the observable thrusting can be constrained to be of middle Miocene age. It is possible, however, that the thrusting may have started earlier, during the Burdigalian to Langhian deposition of the Bernabeles Formation. The large slumps present in the Bernabeles Formation may have been triggered by faulting, or initiated on slopes formed due to faulting, in an active tectonic environment. Continuous convergence between the Internal and External Zones during early to middle Miocene time is therefore possible.

In the Velez Rubio corridor SSE- to SE-directed thrusting probably also took place in Burdigalian to Langhian time, as it affects early to middle Miocene rocks. Deformation related to this thrusting extends across the corridor and affects the Malaguide rocks on the south side. The W- to NW-trending dextral faults that can be mapped in the corridor (Fig. 9) clearly cut and hence post-date the thrusts. The ENE-trending sinistral fault along the corridor is apparently unaffected by the thrusts and therefore probably also postdates them. The relative timing of the two sets of strike-slip faults, however, is not clear from local map or field data. The main ENE-trending sinistral fault zone is parallel to the Crevillente fault, which was active in a sinistral sense in post-Messinian time (Bousquet & Montenat 1974), and thus may form part of the N- to ENE-trending set of sinistral faults that affected the eastern Betics in the later Neogene (de Larouzière *et al.* 1988). It seems plausible that there was some transfer of sinistral displacement from the Crevillente fault southwards onto the IEZB along the Velez Rubio corridor in Pliocene time.

The W- to NW-trending dextral faults in the corridor are not in an appropriate orientation to be antithetic to ENE-trending sinistral set. One possible interpretation is that they are synthetic Riedel shears to a cryptic dextral fault parallel to the corridor. There is no supporting evidence for this hypothesis, and it seems unlikely that the evidence for the Riedel shears would be preserved while the main fault zone was not. Relatively minor dextral faults of late Neogene age with west to northwest trends are common throughout the Betic Cordillera, and their occurrence in the Velez Rubio corridor may have nothing to do with the corridor itself.

DISCUSSION

The kinematic data presented in this paper indicate that the dominant observable structures along the IEZB are related to SSE- to SE-directed thrusting, oblique to the strike of the contact. This transport direction is recognizable over a 60 km length of the IEZB in the eastern Betics. Strike-slip structures post-date the thrusting, and are not in an orientation compatible with a dextral strike-slip boundary.

The relationships with Miocene sediments in the vicinity of the Sierra Espuña show that the observable thrust deformation along the IEZB occurred in early to middle Miocene times. This suggests that the boundary was the locus for oblique convergence between the Internal and External Zones at this time. This convergence affected much of the Subbetic, causing extensive folding and thrusting, and then migrated progressively northwestwards into the Prebetic.

Recent palaeomagnetic work by Osete *et al.* (1989), Platzman (1992) and Allerton *et al.* (in press) in the Subbetic and in the Sierra Espuña has identified large clockwise rotations about vertical axes. These rotations are consistent with the dextrally oblique convergence suggested by the kinematic data, but they raise the possibility that the kinematic data themselves have been rotated. We cannot eliminate this possibility. The kinematic data vary by less than 30° over 60 km, however; whereas the observed rotations vary from 0° to more than 180° over distances of the order of 20 km (Allerton *et al.* in press). For this reason we think it is likely that the IEZB is a master fault separating domains within which fault slices were rotated, but that the IEZB itself and the kinematic data have not been affected by the differential rotations observed palaeomagnetically.

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REFERENCES

Allerton, S., Lonergan, L., Platt, J. P., Platzman, E. S. & McClelland, E. In press. Palaeomagnetic rotations in the eastern Betic Cordillera, southern Spain. *Earth Planet. Sci. Lett.*

Balanyá, J. C. & García-Dueñas, V., 1987. Les directions structurales dans le Domaine d'Alborán de part et d'autre du Détroit de Gibraltar. *C. r. Acad. Sci., Paris* **304**, 929–932.

Banks, C. J. & Warburton, J. 1991 Mid-crustal detachment in the Betic system of southeast Spain. *Tectonophysics* **191**, 275–289.

Blow, W. H. 1969. Late Middle Eocene to Recent planktonic foraminiferal biostratigraphy. *Proc. Ist Int. Congr. Planktonic Microfossils* **1**, 199–422.

Borradaile, G. J. 1976. "Structural facing" (Shackleton's Rule) and the Palaeozoic rocks of the Malaguide Complex near Vélez Rubio, SE Spain. *Verh. K. Ned. Akad. Wetenschapp.* **79**, 330–336.

Bouillin, J. P., Durand-Delga, M. & Oliver, P., 1986. Betic, Rifian and Tyrrhenian Arcs: Distinctive features, genesis and development stages. In: *The Origin of Arcs* (edited by Wezel, F. C.). Elsevier, Amsterdam, 281–304.

Bousquet, J. C. & Montenat, C. 1974. Tectonique-Présence de décrochements nord-est-sud-ouest plio-quaternaires, dans les Cordillères bétiques orientales (Espagne). Extension et signification générale. *C. r. Acad. Sci., Paris* **278**, 2617–2620.

Casas, J. M. & Sàbat, F., 1987. An example of three-dimensional analysis of thrust related tectonites. *J. Struct. Geol.* **9**, 647–657.

de Larouzière, F. D., Bolze, J. Bordet, P., Hernandez, J., Montenat, C. & Ott d'Estevou, P., 1988. The Betic segment of the lithospheric trans-Alboran shear zone during the Late Miocene. *Tectonophysics* **152**, 41–52.

De Smet, M. E. M. 1984a. Wrenching in the External Zone of the Betic Cordilleras, southern Spain. *Tectonophysics* **107**, 57–79.

De Smet, M. E. M. 1984b. *Investigations of the Crevillente Fault Zone and its Role in the Tectogenesis of the Betic Cordilleras, Southern Spain.* Free University Press, Amsterdam.

Durand-Delga, M. 1980. La Méditerraneé occidental: étapes de sa genèse et problèmes structuraux liés à celle-ci. *Mém. Soc. géol. Fr.* **10**, 203–224.

Egeler, C. G. & Simon, O. J. 1969. Sur la tectonique de la Zone Bétique (Cordillères Bétiques, Espagne). *Verh. K. Ned. Akad. Wetenschapp.* **25**.

Fallot, P. 1945. Estudios geologicos en la zona subbètica entre Alicante y el rio Guadiana. *Menor. Mem. Inst Lucas Mallada C.S.I.C.*

Fontboté, J. M. & Vera, J. A. 1983. La Cordillera Betica. In: *Geologia de España* (edited by Comba, J. A.). *Libro Jubilar J. M. Rios* **2**, 205–412.

Frizon de Lamotte, D., Guézou, J.-C. & Albertini, M.-A. 1989. Deformation related to Miocene westward translation in the core of the Betic zone. Implications on the tectonic interpretation of the Betic orogen (Spain). *Geodinamica Acta* **3**, 267–281.

García-Dueñas, V. & Navarro-Vilá, F. 1976. Alpujarrides, Malaguides et autres unités allochtons au Nord de la Sierra Nevada (Cordillères Bétiques, Andalousie). *Bull. Soc. géol. Fr.* **18**, 641–648.

García-Hernandez, M., Lopez-Garrido, A. C., Rivas, P., Sanz de Galdeano, C. & Vera, J. A. 1980. Mesozoic palaeogeographic evolution of the external zones of the Betic Cordillera. *Geologie Mijnb.* **59**, 155–168.

Geel, T. 1973. The geology of the Betic of Malaga, the Subbetic and the zone between these two units in the Velez Rubio area (southern Spain). *Geophys. Un. Am. Pap. Geol. Ser.* 1, **5**.

Hermes, J. J. 1978. The stratigraphy of the Subbetic and southern Prebetic of the Velez Rubio–Caravaca area and its bearing on transcurrent faulting in the Betic Cordilleras of southern Spain. *Verh. K. Ned. Akad. Wetenschapp.* **B81**.

Instituto Geologico y Minero de España. 1972. Mapa Geologico de España, 1:50,000. Coy (No. 932).

Leblanc, D. 1990. Tectonic adaptation of the External Zones around the curved core of an orogen: the Gibraltar Arc. *J. Struct. Geol.* **12**, 1013–1018.

Leblanc, D. & Olivier, P. 1984. Role of strike-slip faults in the Betic-Rifian orogeny. *Tectonophysics* **101**, 345–355.

Lonergan, L. 1991. The structural evolution of the Sierra Espuña, Betic Cordillera, SE Spain. Unpublished D.Phil. thesis, University of Oxford.

Lonergan, L. 1993. Timing and kinematics of deformation in the Malaguide Complex, Internal Zone of the Betic Cordillera, SE Spain. *Tectonics* **12**, 460–476.

Mäkel, G. H., 1985. The Geology of the Malaguide complex and its bearing on the geodynamic evolution of the Betic Rif orogen (Southern Spain and Northern Morocco). *Geophys. Un. Am. Pap. Geol. Ser.* 1, **22**.

Martín-Algarra, A., 1987. Evolucion geologica Alpina del contacto entre las zonas internas y las zonas externas de la Cordillera Betica. Unpublished Ph. D. thesis, University of Granada.

Martín-Algarra, A., Sanz de Galdeano, C. & Estévez, A. 1988. L'évolution sédimentaire miocène de la région au nord de la Sierra Arana (Cordillères bétiques) et sa relation avec la mise en place du bloc d'Alboran. *Bull. Soc. géol. Fr.* **8**, 119–127.

Martini, E. 1971. Standard Tertiary and Quaternary calcareous nannoplankton zonation. In: *Proceedings of II Planktonic Conference, Roma* 1970, *Volume* 2 (edited by Farinacci, A.), 739–785.

Molina, E. 1979. Oligo-Mioceno inferior por medio de forameniferos planctonicos en el sector central de las Cordilleras Béticas. Unpublished thesis, University of Granada.

Okada, H. & Bukry, D. 1980. Supplementary modification and introduction of code numbers to the low latitude coccolith biostratigraphic zonation (Bukry 1973, 1975). *Mar. Micropalaeont.* **5**, 321–325.

Osete, M. L., Freeman, R. & Vegas, R. 1989. Palaeomagnetic evidence for block rotations and distributed deformation of the Iberia–African plate boundary. In: *Paleomagnetic Rotations and Continental Deformation* (edited by Kissel, C. & Laj, C.). NATO ASI Series. Kluwer, Dordrecht, 381-391.

Paquet, J. 1969. Etude Géologique de l'Ouest de la Province de Murcia. *Mém. Soc. géol. Fr.* **48**.

Paquet, J. 1974. Tectonique éocène dans les Cordillères Bétiques: Vers une nouvelle conception de la paleogéographie en Mediterranée occidentale. *Bull. Soc. géol. Fr.* **16**, 58–71.

Platzman, E. S. 1992. Paleomagnetic rotations and the kinematics of the Gibraltar arc. *Geology* **20**, 311–314.

Rivière, M., Bourgois, J. & Feinberg, H. 1980. Evolution de la zone bétique au Miocène inférieur: asynchronisme tectonique entre l'Est et l'Ouest (Cordillères bétiques Espagne). *C. r. somm. Soc. géol. Fr.* **1**, 21–24.

Sanz de Galdeano, C. 1983. Los accidentes y fracturas principales de las Cordilleras Béticas. *Estudios Geol.* **39**, 157–165.

Sanz de Galdeano, C. 1990. Geologic evolution of the Betic Cordilleras in the western Mediterranean, Miocene to present. *Tectonophysics* **172**, 107–119.

Torres-Roldán, R. 1979. The tectonic subdivision of the Betic Zone (Betic Cordilleras, Southern Spain): Its significance and one possible geotectonic scenario for the westernmost Alpine belt. *Am. J. Sci.* **279**, 19–51.