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Geology and tectonics of the Betic Zone, SE Spain

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ABSTRACT

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The southern margin of the Iberian continent comprises a complex geological history, which can be traced back into the Proterozoic when a gneissic basement was formed. A cover of Nevado–Filabride rocks was deposited onto this Proterozoic basement during the Palaeozoic. The Permian intrusion of the Bédar granite suggests that the Palaeozoic sedimentary rocks now exposed in the Betic Cordilleras have been affected by the Hercynian orogeny, similar to that of Palaeozoic rocks in the adjoining Ossa Morena zone. The Hercynian basement was probably covered by Alpujarride carbonates in the Mesozoic, when rifting opened Tethys and intruded 140 Ma old basic dyke swarms.

The Palaeozoic basement of the Betic Cordilleras comprises a suture of Late Cretaceous subduction (85–80 Ma B.P.). This can be inferred from (HP,LT)-metamorphic facies obducted in coeval thrust slices of Nevado–Filabride rocks. The Nevado–Filabride nappes include Mesozoic carbonates transformed into marbles (Macael). The suture was probably annnealed after the closure of Tethys.

The Mesozoic boundary between Africa and Europe and the rifting in Tethys ceased during the Palaeogene when the Atlas Mountains were formed (36 Ma B.P.). Subsequently, relative movements between Africa and Europe in the Neogene were accommodated by strike–slip faults. These define the Iberian microplate moving eastward by Atlantic spreading and possibly subducting under Italy.

Neogene deformation of Iberia is most intensive in southeastern Spain due to emplacement of a mantle diapir under the Alboran area. This occurred early in the Neogene (Aquitanian) and caused nappe spreading involving Alpujarride and Malaguide rocks. The subsequent Neogene history has now been studied in great detail and is reviewed here. It involves clear examples of calc-alkaline volcanism – most probably not related to subduction, seismic motion along strike–slip faults within the Iberian microplate, and syntectonic deposition in intermontane Neogene basins.

1. INTRODUCTION

The Iberian continent occupies a key position in reconstructions of the relative motion of the African and European tectonic plates. The southern margin of Iberia comprises the Betic Cordilleras, a Neogene fold belt which continues uplifting at the present time. This fold belt has been interpreted by some workers as a straight forward product of the collision between Africa and Spain (cf. Dewey et al., 1973). Field data of the area accumulated over the past century suggest a rather complex geological history, which deserves special attention.

Since systematic mapping was begun after the devastating earthquake of Málaga in 1884 (Missión D'Andalousie, 1889), southern Spain has been investigated by several generations of geoscientists of various nationalities. Consequently, the available information is dispersed over a variety of sources and languages – only relatively few results have been published in English. This situation led to the false impression that the geology of SE Spain had been poorly studied. For example, geological knowledge of the Betic Cordilleras would be "still in the classic form of standing on mountain tops waving one's arms in grand demonstrations of major structural units", according to the prestigeous work "The Geology of Europe" (Ager, 1980: p. 365). Many important studies have appeared only in Spanish, French, Dutch, German, or Italian journals. The majority of detailed mapping data has not been published at all but is recorded in a vast number of M.Sc.- and Ph.D.-theses of which copies are preserved in relatively few European universities. The diverse literature dealing with the area is also confused and sometimes contradictory.

The present study outlines the geological data available and attempts to reconstruct the geodynamic history. Discussed are: the geological setting (section 2); physiography (section 3); Betic units (section 4); Neogene sedimentary cover (section 5); Neogene volcanic rocks (section 6); seismic faults (section 7); geodynamic history (section 8) and recommendations for future research (section 9). This synthetic review uses both published and unpublished sources in a variety of European languages.

This review was also inspired by the concern that an important body of information on the geology of southeastern Spain will become less accessible due to the recent upheaval of the geology department of the Municipal University of Amsterdam. The 1988-closure, finalized after earlier cuts in the Dutch science budget (cf. Weijermars, 1983), terminated a long tradition of geological mapping in southeastern Spain by the M.Sc.- and Ph.D.-students of Amsterdam's Municipal University. Projects in the Betic Zone s.s. were initiated by the late professor H.A. Brouwer in the period 1920–1936 and resumed since 1958 by the late professor C.G. Egeler (structural geology) and professors emeriti W.P. de Roever (petrology and mineralogy), H.J. Mac Gillavry (palaeontology), and J.J. Hermes (stratigraphy). Nonetheless, this review emphasizes that the present state of knowledge on the geology of the Betic Cordilleras results from detailed field studies by an international group of geoscientists.

2. GEOLOGICAL SETTING

The Betic Cordilleras (Fig. 1a) are traditionally divided into an External Zone in the north and an Internal (or Betic) Zone in the south (Fallot, 1948). The External Zone is

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Ruud Weijermars, a native of Amsterdam (34), was awarded a B.Sc. and M.Sc. in Structural Geology (Holland) and Ph.D. in Geodynamics (Sweden). Working as a teaching assistant at Amsterdam University, he was partially responsible for the organisation of annual field courses in structural mapping held in various areas in Spain and Germany (1979–1983). His M.Sc.-thesis (1981) covered an area in the Betic Zone, which has captured his interest ever since. Weijermars continued working in the Betics after moving to Sweden, culminating in a 4 month field survey in 1988.

More recently, Dr. Weijermars has taken up new investigations into rock rheology and the ways in which laboratory modelling may advance our understanding of rock structures and geodynamic processes. He gained experience in scale modelling of tectonic processes working in the tectonic laboratories of Uppsala University, Sweden (1983–1988), the Swiss Federal Institute of Technology, Zürich (1988–1989) and the University of Texas at Austin, USA (1990).

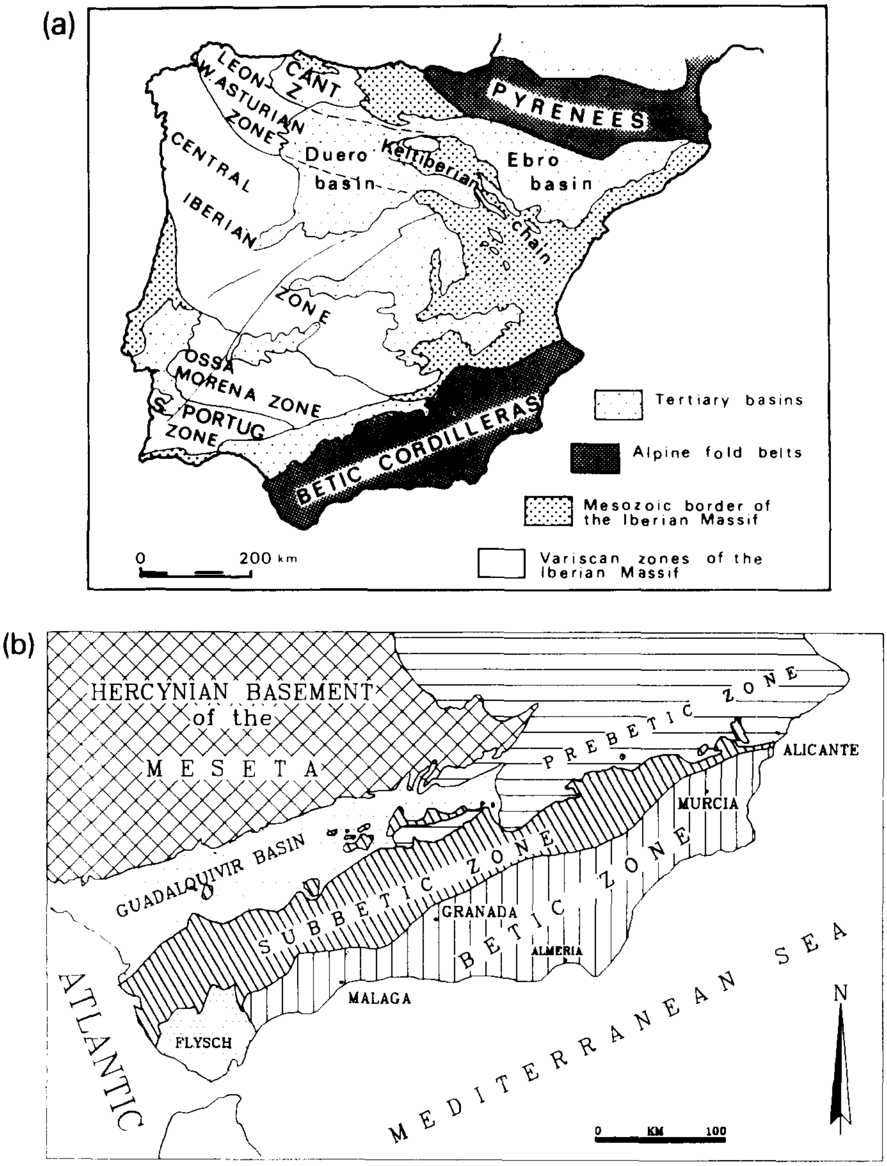


Fig. 1. (a) Tectonic map of the Iberian Peninsula showing the location of the Betic Cordilleras in SE Spain and how it is overprinting the Variscan Zones of the Iberian Massif (after Rondeel et al., 1984, fig. 1). (b) Structural sketchmap of the major tectonic provinces of the Betic Cordilleras. Major provinces are: the Internal or Betic Zone s.s., Subbetic and Prebetic Zones (from Dronkert, 1985, fig. 1.8).

subdivided into a Prebetic and Subbetic district (Fig. 1b), only comprising non-metamorphic sedimentary rocks of essentially Mesozoic and Tertiary age. The Internal or Betic Zone s.s. comprises mainly Triassic and older clastic metasedimentary rocks, exposed in thrust nappes (Egeler and Simon, 1969a,b; Rondeel and Simon, 1974; Kampschuur and Rondeel, 1975; Egeler and Fontboté, 1976).

The Prebetic Zone is generally regarded as the autochthonous to parautochthonous cover of the Spanish Meseta (Fallot, 1948). More specifically, it comprises a Mesozoic to Tertiary platform and shelf sequence which has been deformed by thin-skinned compressional tectonics (García-Hernandez et al., 1980). The Subbetic Zone separates the Prebetic foreland fold-belt in the north from the metamorphic Betic Zone in the south (Fig. 1b). Subbetic sedimentary rocks include Cretaceous to Early Tertiary deep-water sequences, associated with minor basaltic volcanic rocks, and may have been deposited in a mid-Jurassic rifting basin (Hermes, 1978). Subbetic sedimentary rocks were thrust onto the Prebetic in Tortonian times (Jerez-Mir, 1973; Hoedemaeker, 1973; Azéma, 1977; García-Hernandez et al., 1980).

The area reviewed here lies entirely within the Internal Zone of the Betic Cordilleras. However, the geological map of Fig. 2 illustrates the correspondence between the Internal Zones of the Betic and Rif orogens and groups the nappes into various tectonic complexes. The Nevado–Filabride Complex, only exposed in the eroded core of the Betic mountain range (Fig. 2), is overlain by the so-called Higher Betic nappes. The Higher Betic nappes comprise, in ascending order, the Alpujarride, Almagride and Malaguide nappe complexes (Fig. 2). It is worth noting that Alpujarride fold nappes in the westernmost part of the Betic Zone comprise the Ronda peridotite massif (Fig. 2), the world's largest exposure of upper mantle rocks (300 km^2). The Ronda massif, now commonly regarded as a thrust sheet itself (Westerhof, 1977; Lundeen, 1978; Tubía and Cuevas, 1986, 1987), is only minimally altered by serpentinisation and therefore probably one of the few well preserved pieces of mantle rock available (e.g. Reisberg and Zindler, 1986; Suen and Frey, 1987; Reisberg et al., 1989).

The geology of the Betic Zone can best be studied in the province of Almería (Fig. 3a), because the garigue to semi-desert vegetation leaves the rock formations extremely well exposed. The area comprises metamorphic basement inliers covered by Neogene sedimentary rocks and is transected by a volcanic ridge

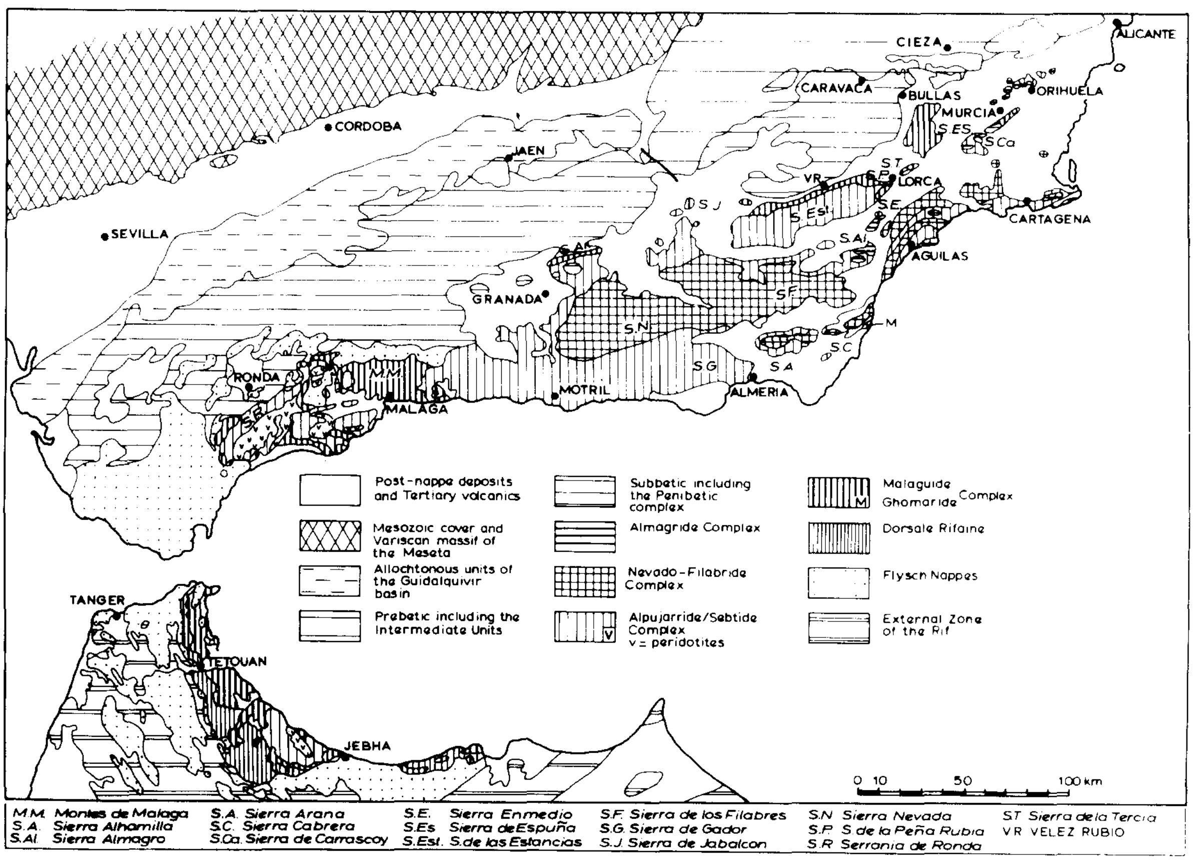


Fig. 2. Geotectonic map of the Betic–Rif orogen peripheral to the Alboran Basin (from Mäkel, 1985, fig. 1–2).

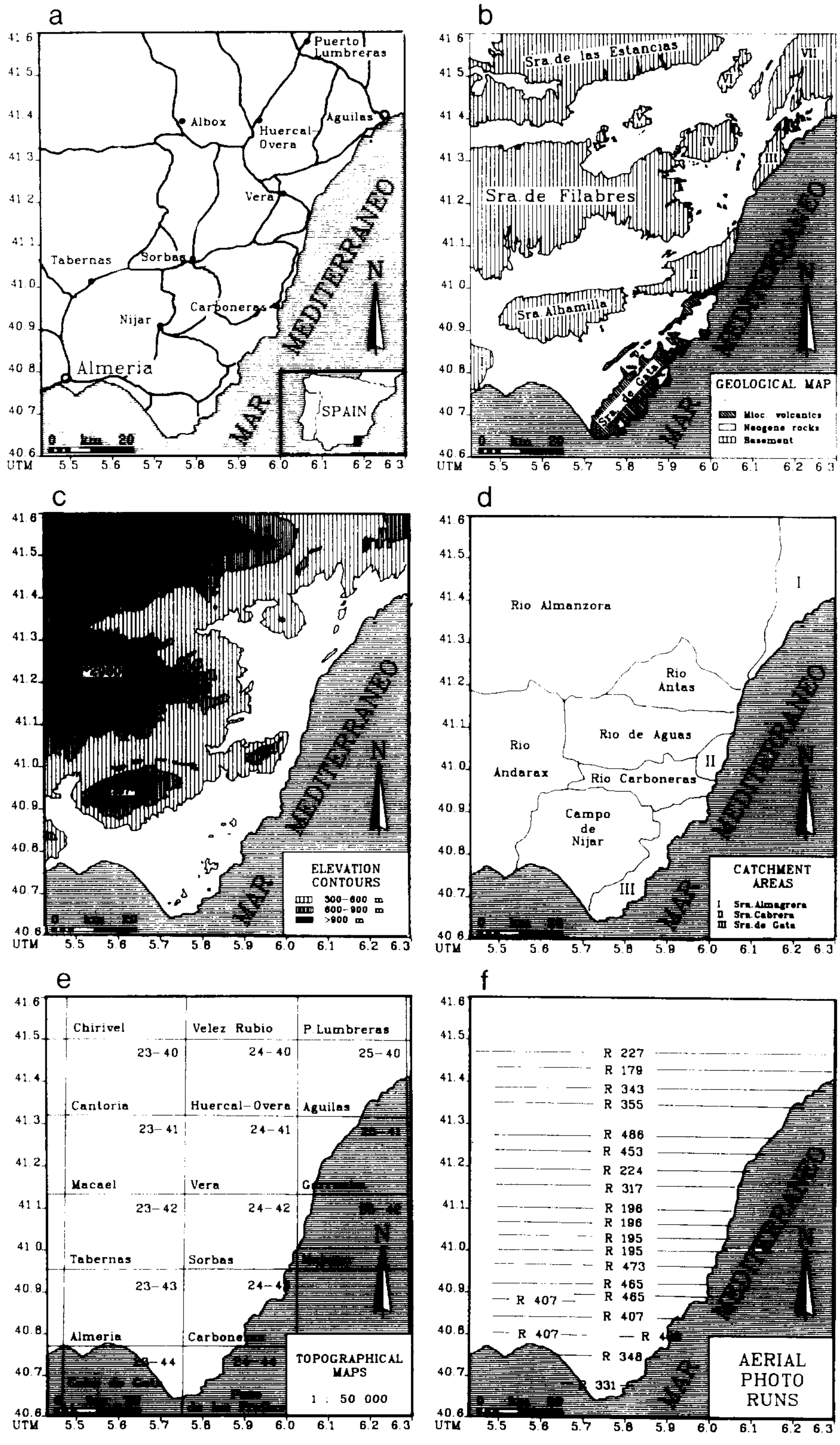


Fig. 3. Physiographic maps of Almería province in SE Spain showing: (a) geographical locations, (b) geology, (c) morphology, (d) water catchment areas, (e) index of available geological and topographical maps, and (f) index of runs covered by aerial photographs (after Dronkert, 1985, figs. 1.1 and 1.3–1.7).

(Fig. 3b). The geomorphology of the province of Almería is to a great extent controlled by geological structures. Highs in the regional relief (600–2000 m; Fig. 3c) coincide with axial culminations in the inliers of refolded basement rocks of the Sierras de los Filabres, Alhamilla and Estancias (Fig. 3b). The topographic lows (< 600 m) are occupied by the Neogene sedimentary rocks which form intermontane basins between the basement highs (Fig. 3b and c). The southeastern coastline is entirely determined by the strike of the ridge of calc-alkaline volcanic rocks of the Sierra de Gata (Fig. 3b). The Neogene plains are the most accessible, due to a rapidly expanding network of both metalled and non-metalled roads, whilst the mountaineous basement areas and volcanic ridge of the Sierra de Gata are only transected by a few roads.

The province of Almería is covered by 1 : 50 000 topographic and geological map sheets outlined in Fig. 3e. The topographic maps (Mapa Militar de España) are reliable, but the maps of the Geological Survey (Mapa Geológico Nacional) pertinent to the area of interest have not yet incorporated improved mapping data available from recent academic studies. Detailed geological maps from academic studies ranging in scale from 1 : 50 000 down to 1 : 10 000 have been com-

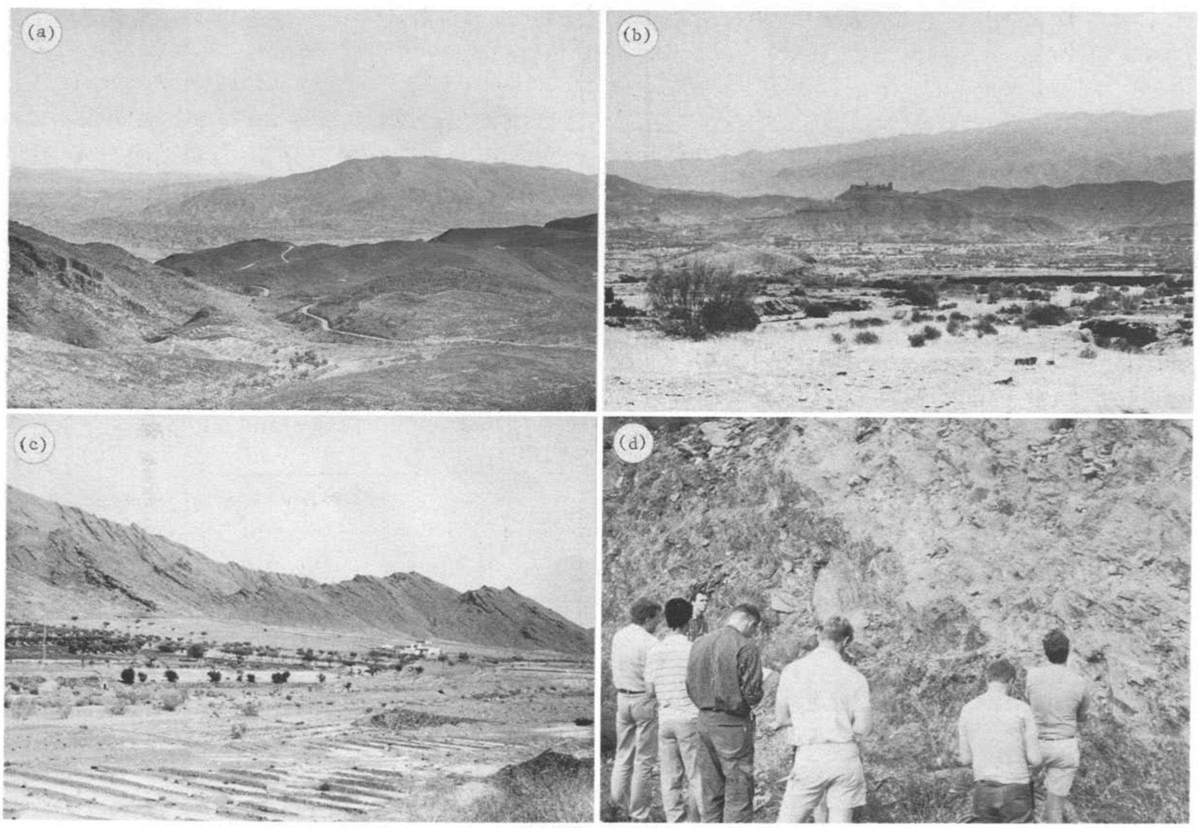


Plate 1. Overview. (a) Oblique view of the regional anticlinal defining the Sierra Alhamilla. Looking towards the NE from the Sierra de Gádor. The summit of the Alhamilla is 1387 m high and the doubly plunging anticlinal seen is about 25 km long. The Tabernas Basin is situated to the north of the Sierra Alhamilla, left part of the picture. (b) Central plain of the Tabernas Basin. Looking towards the SE from El Buho towards Tabernas Castle (el. 491 m), which is at 2 km from the point of view. (c) South margin of the Sierra de los Filabres meeting plains of the Tabernas Basin. Looking east from the road between Tabernas and Senés. Main foliation in the black Tahal schists of the Lubrín unit dips 40° south and partly defines the anticlinal of the Sierra de los Filabres. (d) Contact between Nevada and Tahal schists demarcated by light coloured, yellowish weathering zone in top of the Nevada schist, exposed in road-cut between Senés and Tahal, Sierra de los Filabres. Field party comprises students of the University of Uppsala.

piled in previous studies (i.e. Vissers, 1981; Bordet, 1985; Weijermars et al., 1985; Weijermars, 1987a; De la Chapelle, 1988). A larger compilation covering the southern half of Fig. 3b is given in Fig. 18 (see later). Figure 3f shows the index map for low altitude aerial photographs available from the Instituto Geológico y Minero de España (IGME), Madrid. There are also road maps available from many different manufacturers, but none of the existing maps are up-to-date. The best road map is provided by Michelin's 1 : 400 000 sheet no. 446.

The geology of the area outlined here is extremely well suited for organizing field courses and excursions. A variety of geological features is illustrated in the selected photographs of Plates 1–12. Plates 1a and b give scenic views of the regional anticline in the Sierra Alhamilla and the intermontane Tabernas Basin. Field aspects of the Nevado–Filabride basement, Higher Betic nappes, Neogene sedimentary rocks, Neogene volcanic rocks and strike–slip faults are illustrated in Plates 1–12. These will be discussed in detail below. An outline of a 2-week excursion to southeastern Spain, based upon experience with field parties of up to 25 Ph.D.-students and staff members from Uppsala University, is available upon request.

3. PHYSIOGRAPHY

A brief physiographic outline seems justified since it plays such an important role in maintaining the magnificent present-day exposure and as it is relevant for planning future field surveys. Almería province is drained by five large river beds, i.e. the Ríos de Almanzora, Antas, Aguas, Carboneras, and Andarax (Fig. 3d). These are usually devoid of any water, but may instantaneously be filled with very dangerous mudstreams when the occasional precipitation occurs (torrential rains, whirlwinds or hailshowers with lumps up to fist-size). For example, catastrophic floods in 1973 caused about 100 casualties and washed away several roads, bridges and cars (Dronkert, 1985). The modern fluvial systems would have been established during the Pleistocene when the ancestral Río Carboneras lost part of its drainage network to the Río de Aguas (Harvey and Wells, 1987; see also Harvey, 1984a,b, 1987a,b). Torrential rains supply only 18 cm precipitation/year averaged over the past 25 years and it usually falls in only a few days, leaving the rest of the year completely dry (Rodriguez et al., 1987). Normally, up to 95% of the total precipitation is lost by evaporisation so that groundwater replenishment cannot be meteorologically maintained.

The climate in Almería province varies from garigue via semi-desert to that of a desert environment. The most convenient seasonal climate for field trips occurs in the months of March, April, May, October, and November with diurnally averaged coastal temperatures of 14.2°, 16°, 19°, 19.8°, and 15.3°C and precipitation still remaining at the negligibly low levels of 10 to 30 mm per month (Table 1). The climate in the mountains may be considerably cooler than at the coast. This is because the temperature decreases with 0.37°C and 0.76°C/100 m at altitudes above 0 and 400 m, respectively (Carulla Gratacos, 1977). For example, the theoretical temperature on top of the 1 km

TABLE 1

Climatic data of field area in SE Spain

Month:	Jan	Feb	March	Apr	May	June	July	Aug	Sept	Oct	Nov	Dec
Air temp. (°C)	12.1	12.6	14.2	16	19	21.9	25	25.6	23.3	19.8	15.3	12.8
Sea temp. (°C)	15	14	15	15	17	19	21	23	22	20	17	15
Rain (mm)	29.7	18.4	22.1	12.7	9.3	3.2	0.3	0.4	11.1	21.1	28.2	21.2

Data according to Rodriguez et al. (1987) and Michelin Map no. 446.

high Sierra Alhamilla is 6 °C cooler than that at coastal level and it is 15 °C lower near the astronomic observatory of Calar Alto at 2.1 km height in the Sierra de los Filabres (Fig. 3b and c).

Easter usually provides the best opportunity for excursions to the area, but alternatively the climate around Christmas is still moderate and suitable (Table 1). The diurnally averaged temperature at coastal level is 25 °C in the summer (July–August, with daily peaks of 40 °C), 12 °C in the winter (December–February, with rare lows near freezing point), 16 °C in April and 20 °C in October (Rodriguez et al., 1987). The summer months, with daily peak temperatures of 40 °C, are inconveniently hot and prevailing SSW winds bring humid sea air that hazes the views of the landscape in the summer. Occasionally, snow may cover the mountainous areas above 900 m in the winter (Fig. 3c). The storm frequency is the lowest of the Iberic Peninsula.

4. BETIC UNITS

Figure 4 shows the principal geographical distribution of Betic units, i.e. the metamorphic basement inliers of the eastern Betic Zone. The deepest parts are the so-called Nevado–Filabrides, abundantly exposed in the core of an eroded E–W striking anticlinorium that controls the morphology of the Sierra Nevada and Sierra de los Filabres. Nevado–Filabride rocks are also exposed in the eroded core of the Alhamilla–Cabrera

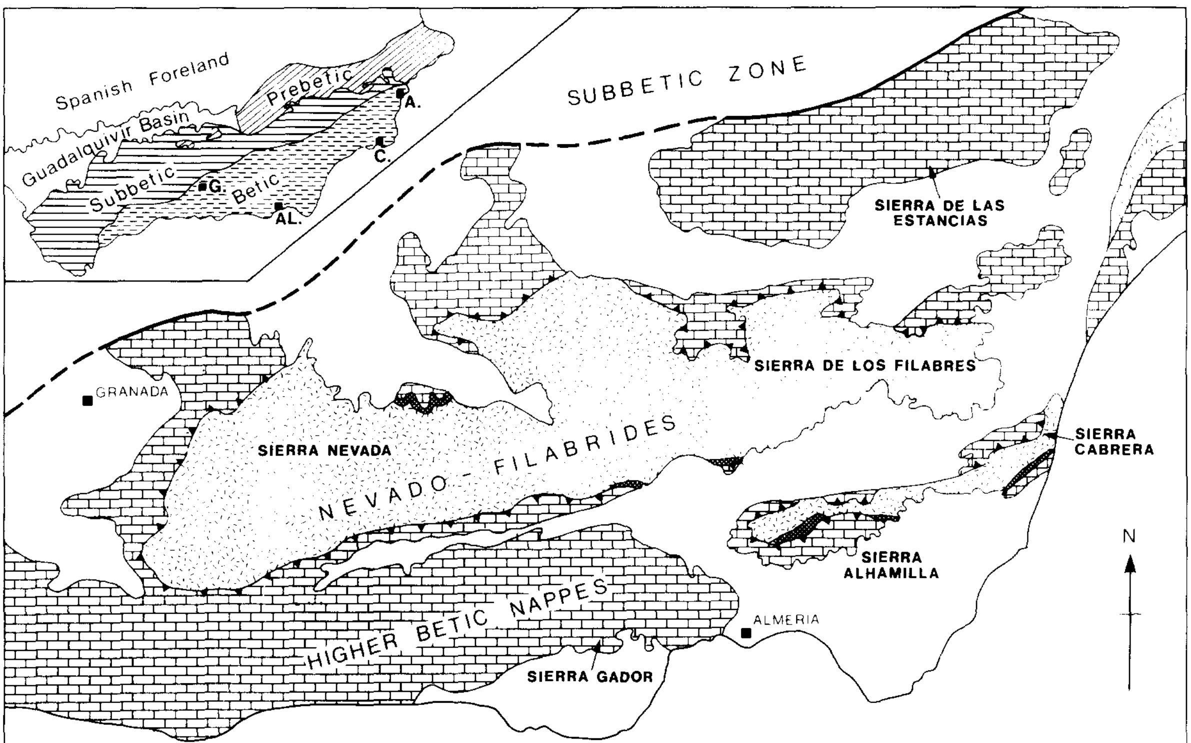


Fig. 4. Distribution of the metamorphic basement inliers of the Betic Zone s.s. in the Betic Cordilleras. The Nevado–Filabride core of the basement is exposed in the eroded regional anticlinorium controlling the topographical strike of the Sierra Nevada and Sierra de los Filabres. Nevado–Filabride units are also exposed in the eroded anticlinal that controls the strike of the Alhamilla–Cabrera Mountain Range. The Nevado–Filabrides are overlain by Higher Betic nappes and the tectonic contact is in many locations demarcated by mylonites and cataclasites in up to 400 m thick zones. The anticlinal of the Sierra de Gádor is not eroded deep enough to disclose its core of Nevado–Filabride basement (from Platt et al., 1983, fig. 1).

anticlinorium (Fig. 4). The contact between the Nevado–Filabrides and the overlying Higher Betic nappes is, in many places, demarcated by mylonitic and cataclastic footwall rocks (Fig. 4). This 0–400 m thick mylonitic carpet has been termed the Betic Movement Zone (Platt and Vissers, 1980; Platt et al., 1984; Platt and Behrmann, 1986; Behrmann, 1987). The regional map patterns of Fig. 4 suggests that the minimum displacement of the Higher Betic nappes over the Betic Movement Zone is in the order of 50–80 km. This assumes nappe transport towards the N–NW as suggested by Platt et al. (1983). However, deformation within the Betic Movement Zone is complex and several components of movement have been distinguished using quartz C-fabric analyses (Galindo-Zalvidar et al., 1989; Zevenhuizen, 1989). The Nevado–Filabrides and Higher Betic nappes will be discussed separately, in sections 4.1 and 4.2, respectively. The state of knowledge on the tectonic subdivision of the Betic units in the 1960's has been outlined by Egeler and Simon (1969a,b).

4.1. Nevado–Filabrides

The nature of the Nevado–Filabride nappe complex will be discussed on the basis of exposures in the Sierra de los Filabres (Fig. 5). The term Nevado–Filabride Complex was introduced by Egeler (1963) to describe all metamorphic rocks exposed underneath the Higher Betic nappes. It included what had been referred to as two different tectonic units since Brouwer (1926a,b): i.e. "crystalline schists of the Sierra Nevada" and the "Mischungszone". These are now commonly referred to as the Nevado-Lubrín unit (Nijhuis, 1964) and the Higher Nevado–Filabride nappes, respectively.

A typical tectonic subdivision of the Nevado–Filabrides is illustrated in Fig. 6. The bulk of the Sierra de los Filabres is occupied by the *Nevado-Lubrín unit*, comprising graphitic, black, garnet-bearing "Nevada" micaschists and overlying, garnet-bearing, albite-chlorite, mica schists of Tahal (Fig. 6). Although these two lithologies look very similar, the Tahal schists are lighter-coloured than the underlying Nevada schists s.s. Their contact is thought to be a stratigraphic unconformity on the basis of (1) horizons of basal conglomerate near the base of the Tahal schist sequence, and (2) the presence of a continuous zone of several meters thickness, comprising yellowish, chloritoid-garnet, mica schist in the top of the Nevada schist sequence s.s. (Linthout and Vissers, 1979). The yellowish zone of micaschist is thought to be a metamorphosed palaeo-weathering surface (K. Linthout, pers. commun., 10-03-1989). Plates 1c and d, and 2 illustrate field aspects of Nevada schists and Tahal schists, constituting the lower and upper half of the Palaeozoic and Permo-Triassic Nevado-Lubrín unit, respectively.

The Tahal schists of the Nevado-Lubrín unit are capped by a zone of up to 75 m thick brecciated marble (rauhwackes; Leine, 1966, 1968, 1971), which is a cataclastic fault gouge underlying the base of the set of overriding *Higher Nevado–Filabride nappes* comprised in Brouwer's (1926a,b) "Mischungszone". The carbonate breccia contains fragments of both the underlying and overlying rocks and individual fragments may range up to meter-size.

Workers of the Granada-school have employed the terms "Veleta" and "Mulhacen nappes" to subdivide the Nevado–Filabrides (Puga, 1976). However, this terminology is somewhat confusing since it remains unclear as to what are the criteria for their distinction. For example, the boundary between the Veleta and Mulhacen complexes could coincide with the weathering zone at the base of the Tahal schists, according to maps of Puga et al. (1988a) and Portugal Ferreira et al. (1988). However, other workers suggest that the top of the pile of Veleta nappes is exposed only further to the west, i.e. much deeper than the weathering zone (García-Dueñas et al., 1988). An attempt to correlate the terminology of the Granada-school with that employed by others would be useful but is

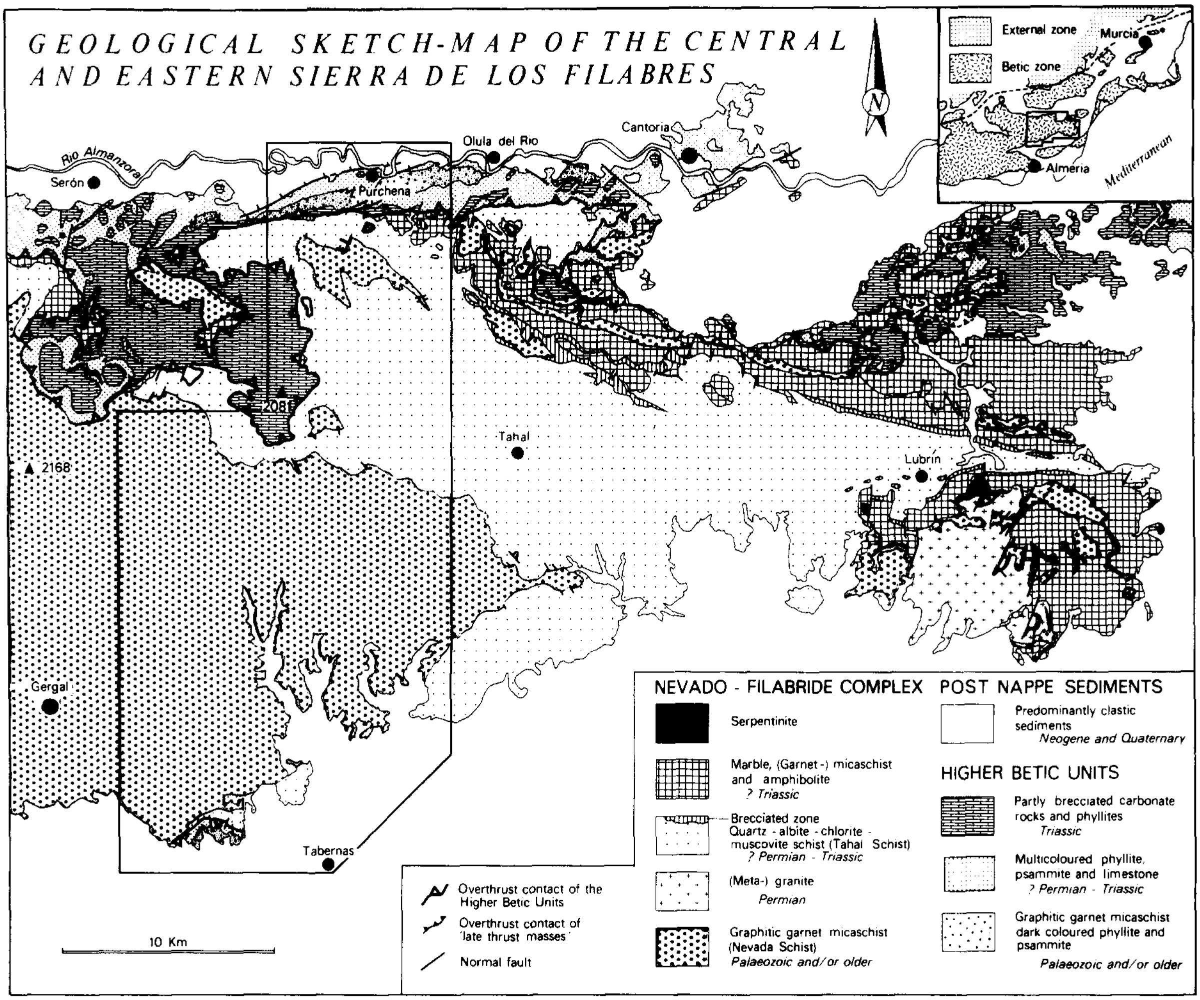


Fig. 5. Geotectonic map of the eastern Sierra de los Filabres compiled by Vissers (1981) on the basis of 50 years of intensive mapping and field surveys by various research groups. Vissers (1981) himself studied in detail the area outlined in the central Sierra de los Filabres between Purchena and Tabernas. The tectono-metamorphic data discussed in Figs. 8–10 are based on observations in this area and Vissers' (1981) microstructural interpretation of about 800 thin sections. Altogether, the Sierra de los Filabres has been covered by at least 5000 thin sections, conserved by the Geological Museum of the Municipal University of Amsterdam. (Major contributions to the knowledge of Nevado–Filabrides in the area mapped were made by De Sierra, 1915; Brouwer and Zeijlmans van Emmichoven, 1924; Zeijlmans van Emmichoven, 1925; Brouwer, 1926a,b; Zermatten, 1929; Brouwer and Jansen, 1933; Jansen, 1936; Patijn, 1937; Fallot, 1948; Fallot et al., 1961; De Roever et al., 1961; Leine and Egeler, 1962; Egeler, 1963; De Roever and Nijhuis, 1963; Nijhuis, 1964; Bicker, 1966; Leine, 1966, 1968, 1971; Voet, 1967; Helmers and Voet, 1967; Linthout and Westra, 1968; Egeler and Simon, 1969a,b; Westra, 1970; Langenberg, 1972; Kampschuur et al., 1973; Kampschuur, 1975; Vissers, 1977a,b, 1981; Linthout and Vissers, 1979.)

beyond the scope of this study. Such a correlation would certainly help to solve the continuing debate concerning a possible pre-Alpine metamorphic event in certain parts of the Nevado–Filabrides (e.g. Puga and Diaz de Frederico, 1976; Vissers, 1977b; Van den Eeckhout and Konert, 1983; Gomez-Pugnaire and Franz, 1988).

The Nevado-Lubrín or Lower Nevado-Filabride unit may not need to be interpreted as a *nappe*, as no conclusive arguments seem to exist against an autochthonous interpreta-

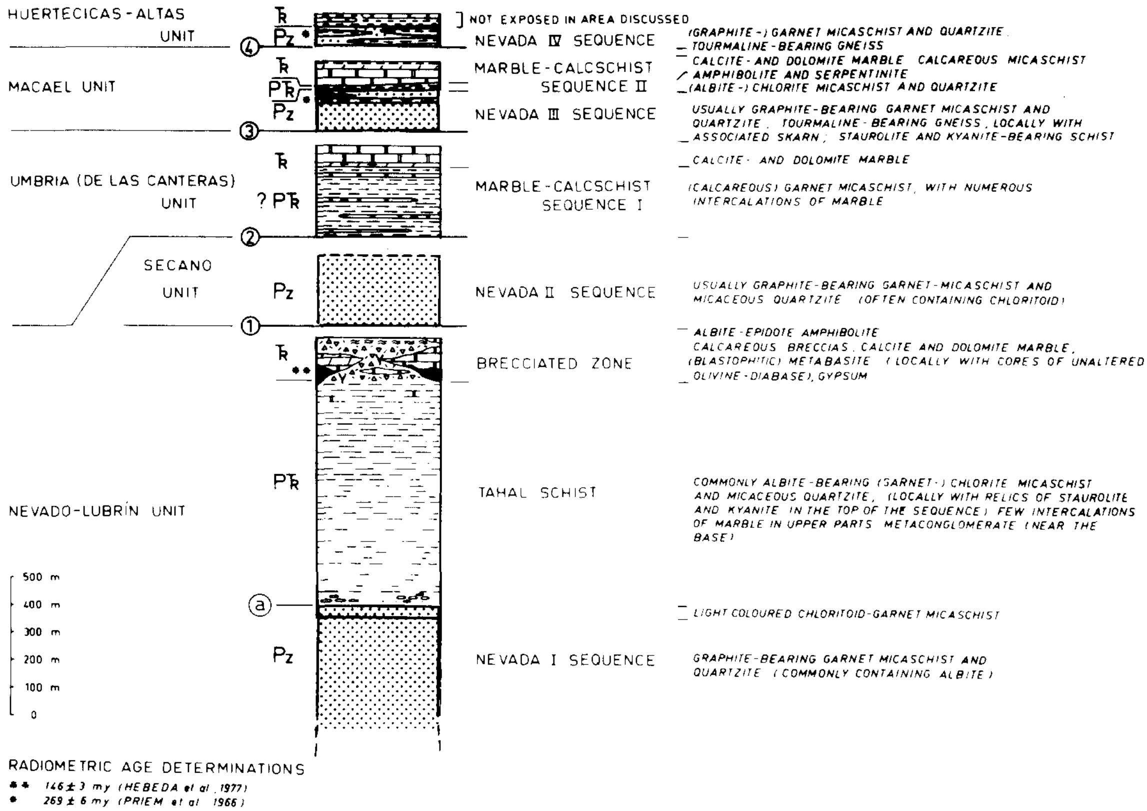


Fig. 6. Typical tectono-stratigraphic section of the Nevado–Filabride complex, showing the Nevado-Lubrín unit, possibly autochthonous, and four overlying Higher Nevado–Filabride nappes of the "Mischungszone". The thrust planes 1–4 correspond to those outlined in Figs. 7a and 7b. These lithological columns only give tectonic thicknesses. The contact labelled "*a*", between Nevada and Tahal schists is a supposed erosional unconformity locally marked by a basal conglomerate. The provisional ages indicated are *Pz* = Palaeozoic, *PTr* = Permo-Triassic and *Tr* = Triassic. These are only suggestions based on the correspondence of the gross-lithological subdivision with similar sequence dated in other regions. The 269 Ma radiometric age indicated in the Macael unit is the intrusion age of granite exposed in the correlative Bédar unit of the southeast Filabres (Priem et al., 1966). The 146 Ma radiometric age is unaltered olivine gabbro in the core of a metabasitic dyke (Hebeda et al., 1977) (after Linthout and Vissers, 1979, fig. 3).

tion of these rocks. In fact, it may may be connected to the Hercynian forelands of the Spanish (Fallot, 1948) and Moroccan Mesetas, and may thus be of Palaeozoic age on the basis of lithological resemblances. No fossils are preserved in the rocks of the Nevado-Lubrín unit due to polyphase isoclinal folding and crystalloblastesis under greenschist facies conditions (up to glaucophane facies in the top of the Tahal sequence, see later). Crinoids in dark marble horizons of some Higher Nevado–Filabride nappes yield ages of Eifelian or Middle Devonian, i.e, 390–380 Ma B.P. (Lafuste and Pavillon, 1976).

The Higher Nevado–Filabride units are all proven nappes comprising a variety of marbles, gneisses, schists, amphibolites and serpentinites (Fig. 6). At least four individual thrust sheets have been mapped along the northern margin of the Sierra de los Filabres, i.e. the nappes of Secano, Umbria, Macael and Huertecicas (Fig. 7a and b). The map pattern of Fig. 5 can be used to suggest that the Umbría, Macael and Huertecicas nappes correlate with the Chive, Bédar and Almocaizar nappes distinguished in the SW flank of the Sierra de los Filabres (Nijhuis, 1964; Helmers and Voet, 1967; Linthout and

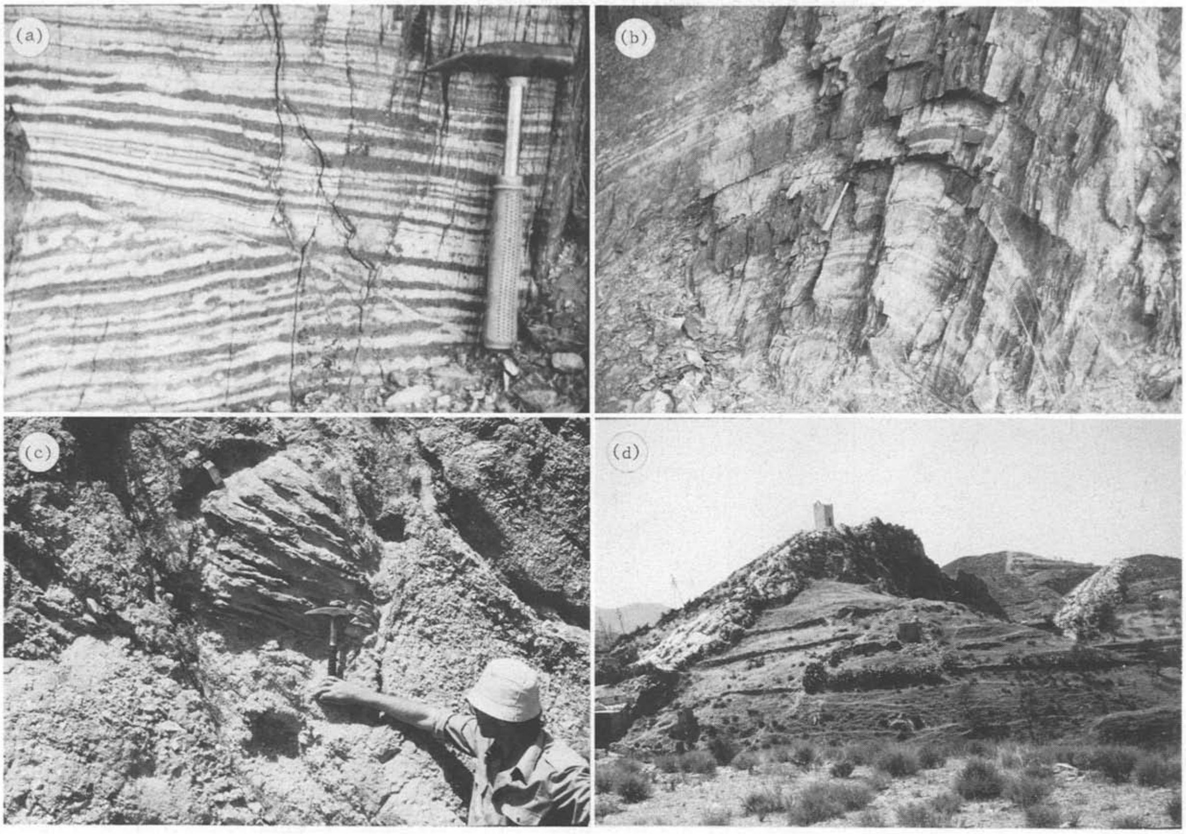


Plate 2. Nevado–Filabrides, Lubrín unit. (a) Bedding preserved in Tahal schist, in road-cut along road between Senés and Tahal, Sierra de los Filabres. Discontinuity in sedimentary bands is either due to faulting or may be primary cross-bedding. (b) Tahal schist, near location of Plate 2a. Bedding may be transposed, but is outlined by refraction of the steep young phase foliation. (c) Rauhwacke or brecciated marble demarcating the boundary between the Higher Nevado–Filabrides and the Lubrín unit in many places. Exposure is along the road between Tahal and Macaél, Sierra de los Filabres. Hammer is supported by Willem Zevenhuizen. (d) Band of brecciated marble defining shallow hills directly east of Lubrín, Sierra de los Filabres. The marble is surrounded by Tahal schists and the regional map pattern (Fig. 5) suggests this is a boudinaged fragment isolated by a minor fold from the main horizon.

Vissers, 1979). Similarly, lithological resemblances can be used to suggest that the Macael unit is correlatable to Platt and Behrmann's (1986) Castro unit in the Sierra Alhamilla and Westra's (1969) Arto Complex in the Sierra Cabrera. A check of this regional correlation of Higher Nevado–Filabride nappe units could be attempted by isotopic dating and whole-rock chemical analysis of gneiss and metagranites exposed in the tectonic units of Bédar (Nijhuis, 1964), Macael (Voet, 1967), Arto (Westra, 1969), Castro (Platt and Behrmann, 1986) and gneisses in a similar unit exposed north of Níjar along the road to Lucaiñena de las Torres. These correlations suggest minimum displacements of 15–40 km for the Higher Nevado–Filabride nappes. Field aspects of some Higher Nevado–Filabride rocks exposed in the Sierra de los Filabres are shown in Plate 3 and that of the Sierra Alhamilla in Plate 4.

The Bédar nappe in the southeast Sierra de los Filabres, sandwiched between the Chive-Macael and Almoicazar-Huertecicas nappes, has been intruded by tourmaline-granites before its tectonic emplacement. The Rb–Sr isochron of the Bédar granite suggests that intrusion took place in the Early Permian 269 ± 6 Ma B.P. (Priem et al., 1966). The subsequent emplacement of the Bédar nappe

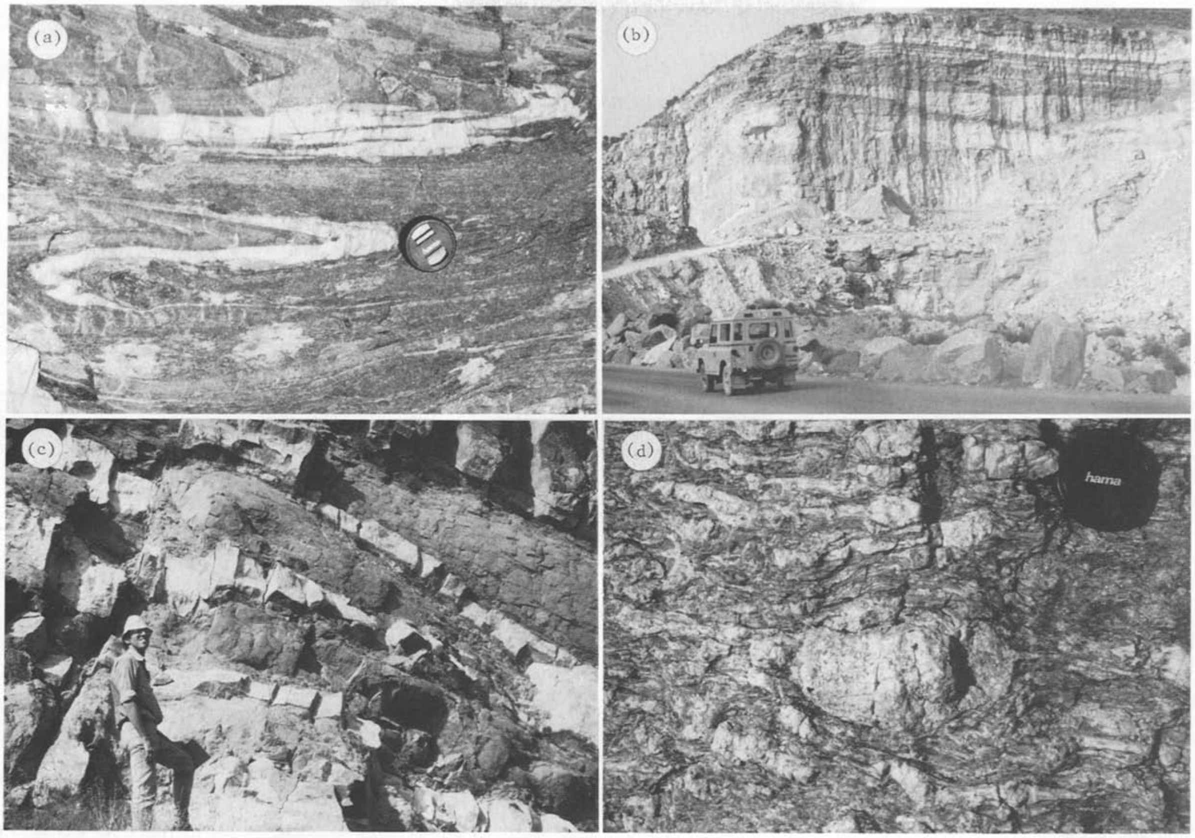


Plate 3. Higher Nevado–Filabrides. (a) Macaél unit. Isoclinally folded quartzite horizon in the base of the Macaél nappe, exposed along road between Macaél and Laroya, N-slope Sierra del los Filabres. (b) Macaél unit. Marble quarry of Sanchez Lopez near Macaél, Sierra de los Filabres. The horizon of pure white marble is only a few meters thick, but is transposed by isoclinal recumbent folding associated with emplacement of the Macaél nappe (W.A. Zevenhuizen, pers. commun., 06-08-1988). (c) Chive unit. Green amphibolite schist comprising bands of isoclinally folded marbles, La Colmenica along road Lubrín-Ulella del Campo, 4 km SE of Lubrín, Sierra de los Filabres. Geologist is Willem Zevenhuizen. (d) Bédar unit. Detail of feldspar augen in strongly sheared portion of the Bedár granite. Exposure near km 3-mark along road between Lubrín and El Chive, Sierra de Bedár, E-flank of the Sierra de los Filabres.

has largely deformed the Bédar granite into tourmaline-gneiss with large feldspar augen. Strain analysis of shear zones in a single outcrop of Bédar granite yielded minimum strain values of 1.5 (Borradaile, 1976).

A microstructural study of omphacite (a metamorphic clinopyroxene) in eclogitic hedenbergite skarns and pyroxenites exposed in schists underneath the Bédar granite (Fig. 5) indicates that eclogitisation took place at 12 kbar and 500 °C (Van Roermund, 1984). Unaltered olivine diabase in the core of metabasite in the top of the Nevado-Lubrín unit close to the Bédar nappe yielded a Rb–Sr whole-rock age of 146 ± 3 Ma B.P. (Hebeda et al., 1977). This diabase could be a remnant of a basic dyke swarm weakening the continental crust in connection with spreading of Tethys in the Jurassic (Weijermars, 1989b). The K–Ar whole-rock dating on the ophiolitic Cobdar metabasites yielded ages of 174 ± 4 Ma and 164 ± 4 Ma B.P. (Portugal et al., 1988; Puga et al., 1989). It puts a wide but useful lower age limit on the metamorphism and subsequent emplacement of the Higher Nevado–Filabride nappes onto the Nevado-Lubrín unit.

The tectono-metamorphic history of the

(a)

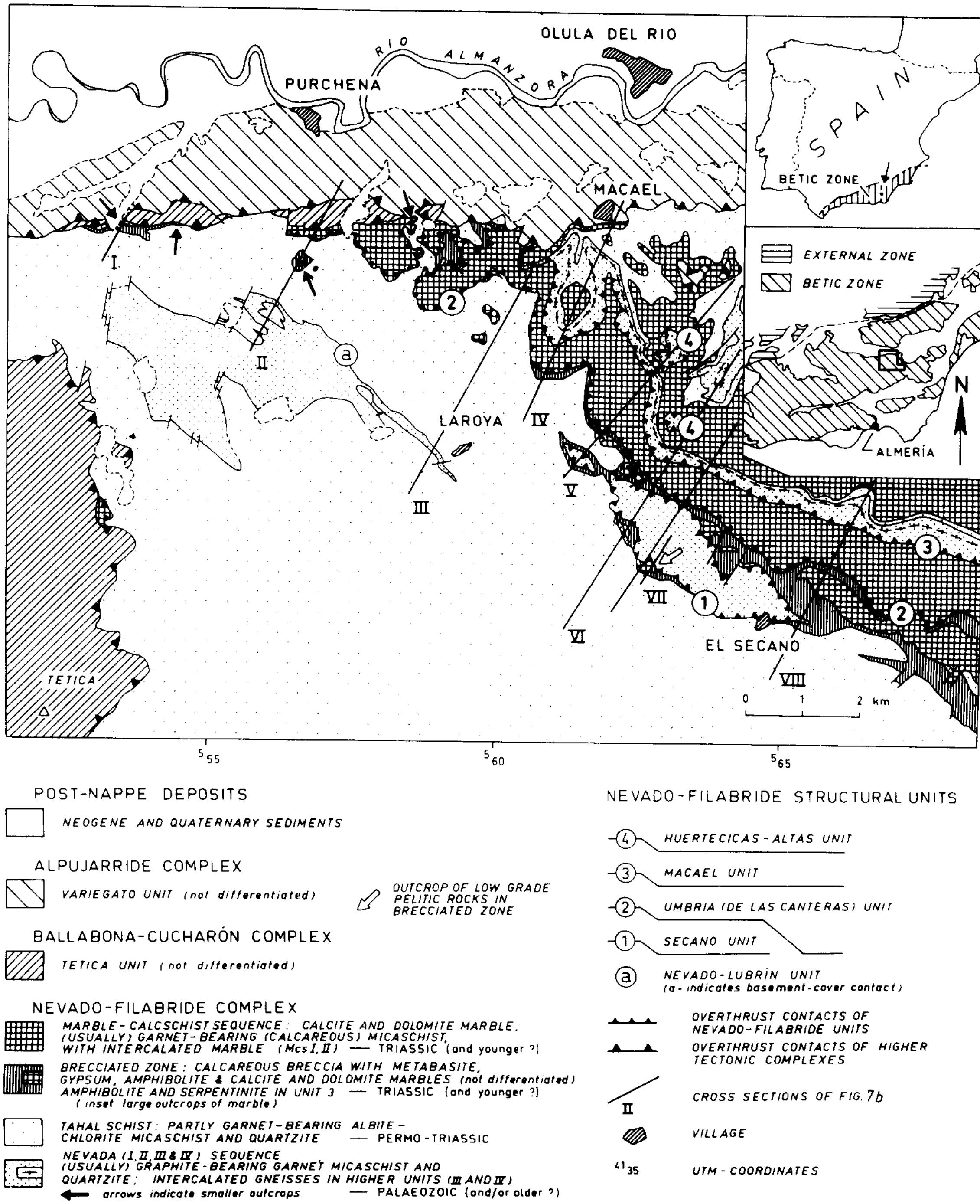


Fig. 7a. Detailed tectonic map of the northern margin of the Sierra de los Filabres near Purchena and Olula del Río (Fig. 5). The four Higher Nevado–Filabride nappes on top of the Nevado-Lubrín unit are underlain by thrusts labelled *1–4*, corresponding to those indicated in Figs. 6 and 7. Cross-sections of Fig. 7b are taken along lines labelled *I–VII*. The famous marbles of Macael are quarried in both the Umbría and Macael nappes (from Linthout and Vissers, 1979, fig. 1).

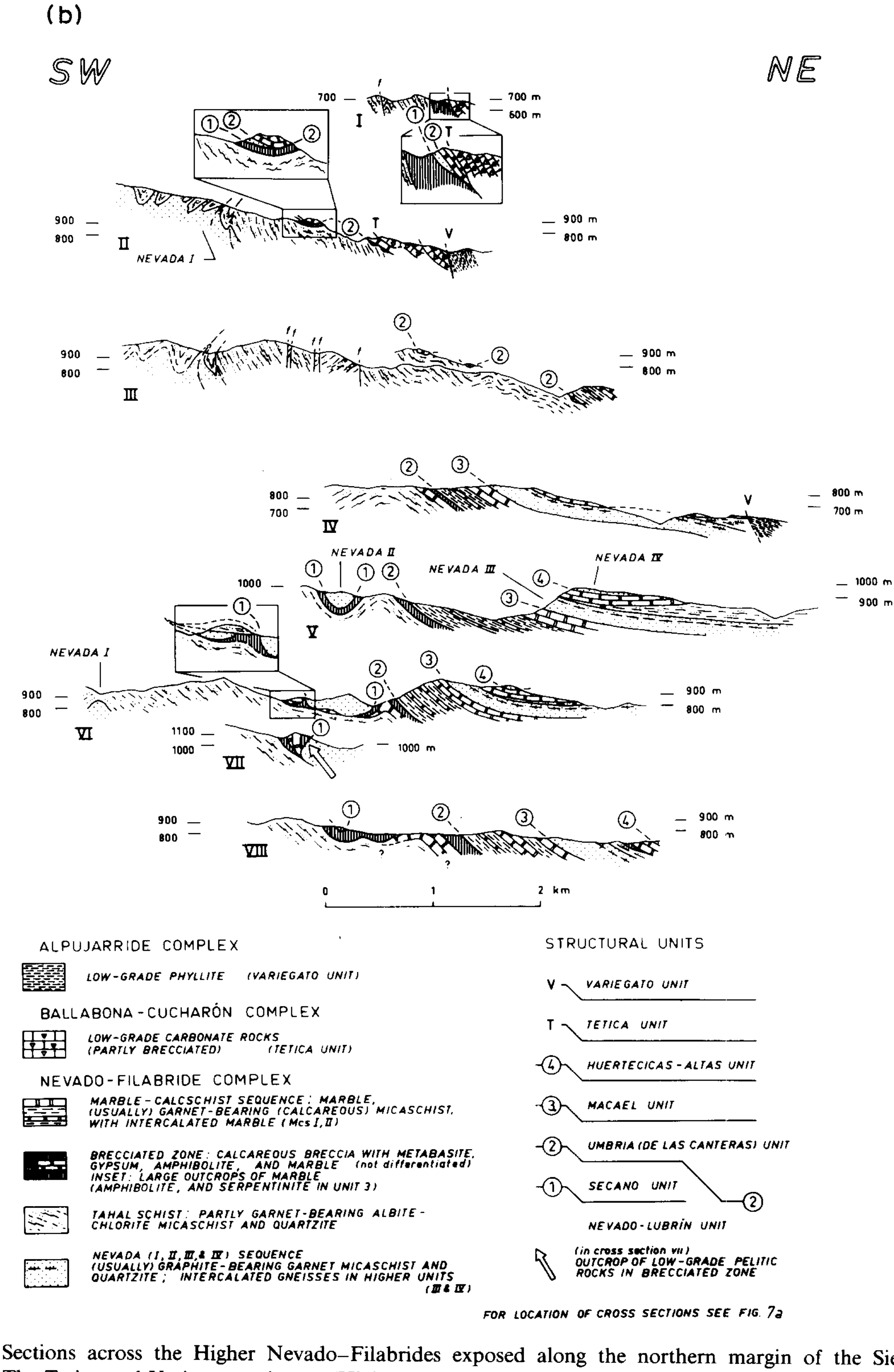


Fig. 7b. Sections across the Higher Nevado–Filabrides exposed along the northern margin of the Sierra de los Filabres. The Tetica and Variegato units are (Higher) Betic nappes, overlying the Nevado–Filabrides (from Linthout and Vissers, 1979, fig. 2).

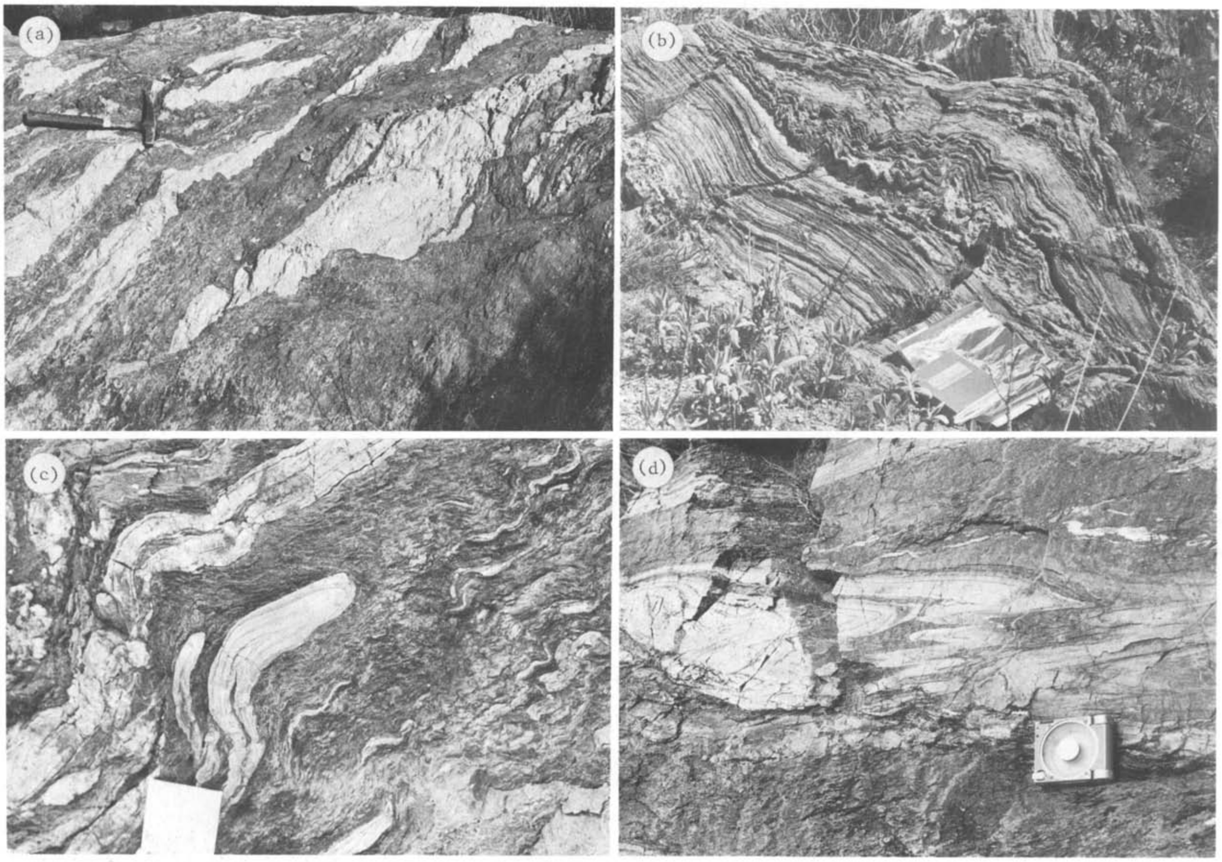


Plate 4. Castro Zone, Sierra Alhamilla. (a) Contact Castro Zone/Alhamilla unit. Nevada-type black-grey mica-schists of the Alhamilla unit are separated from the overlying Castro unit by a tectonic mélange. Slivers are leucocratic schists within grey micaschists. E-branch of Rambla de la Gallarda (exposure 368 of Weijermars, 1981). (b) Castro Zone. Brown-white marble with minor folds refolding isoclinal fold defined by dark pelitic bands. Looking N, near road directly north of La Mina (exposure 228 of Weijermars, 1981). (c) Castro Zone. Rootless intrafolial fold in quartzitic layer refolded by younger phase. Barranco Inferno. (d) Castro Zone. Disharmonic similar folds of quartzitic bands in micaschist matrix. Rambla Honda.

Nevado–Filabrides, studied in detail by Vissers (1981), will be summarized here. Five phases of folding can be distinguished, labelled D_1–D_5, as summarized in Fig. 8a. The significance of D_5–D_4 folding on outcrop scale is of relatively minor importance, but causes the important fold closure of the large anticlinorium visible on the regional map pattern of Fig. 6. This regional anticlinorium is also apparent from the enveloping surface of the dominant schistosity (S_2) in the Nevado-Lubrín unit (see Fig. 18 and Vissers, 1981). The Nevado–Filabride updoming and the associated minor T_c-thrusting towards the south (Fig. 8b) occurred during the Younger Neogene [1] (e.g. Völk and Rondeel, 1964), about 15–7 Ma B.P., as the Sierra de los Filabres supplied detritus to Tortonian sedimentary rocks of the Tabernas Basin (Weijermars et al., 1985).

Open to tight D_3-folding (Fig. 8a) is related to the northward emplacement of the Al-

[1] The term "Younger Neogene" has a specific meaning (see section 5) and is widely used in the geological literature on southeastern Spain. Use of "Late Neogene" instead of "Younger Neogene" as recommended for stratigraphic time by the International Subcommission on Stratigraphic Classification (Hedberg, 1976) is therefore inappropriate in this particular sense.

pujarride or higher Betic units during thrusting T_B (Fig. 8b). This emplacement can be interpreted as having taken place between 20 and 25 Ma B.P., according to a tectonic scenarios discussed in detail in section 8 (see also Weijermars, 1985a,b). D_2 and D_1-folding are isoclinal, causing tectonic thinning and repetition of lithologies associated with complex fold interference patterns. However, the emplacement of the Higher Nevado–Filabride ("Mischungszone") nappes by thrusting T_A occurred after D_2-folding ceased (Fig. 8a and b), so the nappe contacts appear relatively undistorted on the map of Fig. 6.

The plurifacial metamorphic history of the Nevado–Filabride Complex shows an interesting retrograde development from high-pressure low-temperature (HP-LT) towards low-pressure low-temperature (LP-LT) environment (Vissers, 1981). The relative timing of the metamorphism and deformational structures in the Nevado-Lubrín unit and the Higher Nevado-Filabrides is outlined in Fig. 9a and b, respectively. Glaucophane-eclogite facies metamorphism (indicated by the glaucophane-, epidote-, chlorite-, actinolite-assemblage of porphyroblasts) occurred pre- to syn-kinematic with D_1 in the top of the Tahal schists of the Nevado-Lubrín unit (Fig. 9a) and throughout the Higher Nevado–Filabrides (Fig. 9b). This metamorphism has been recognized previously in both the Sierra de los Filabres (De Roever and Nijhuis, 1963; Nijhuis, 1964; Langenberg, 1972) and the Sierra Nevada (Puga, 1971; Diaz de Federico et al., 1979).

However, it was first interpreted as a subduction feature by Vissers (1981) using hints of Torres-Roldán (1979). The extraction and upheaval of the glaucophane schists from the

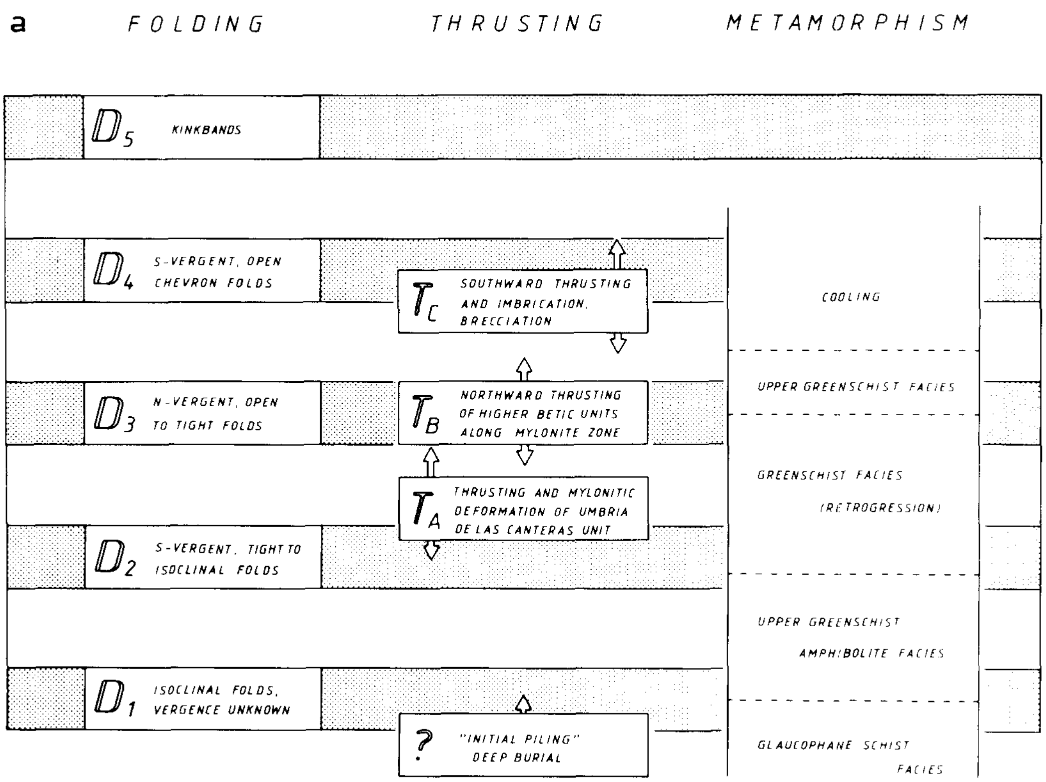


Fig. 8a. Tectono-metamorphic scheme showing the relative timing of nappe emplacement by thrusting phases (T_A–T_C) with respect to folding phases (D_1–D_5) and the development of metamorphic facies. Uncertainty in the age of thrusting events T_A–T_C is indicated by the range of the arrows outlined. All data refer to the Nevado-Lubrín unit of the central Sierra de los Filabres. Details of the tectonics and metamorphic facies development are outlined in Figs. 9a and 9b, respectively (after Vissers, 1981, fig. 49).

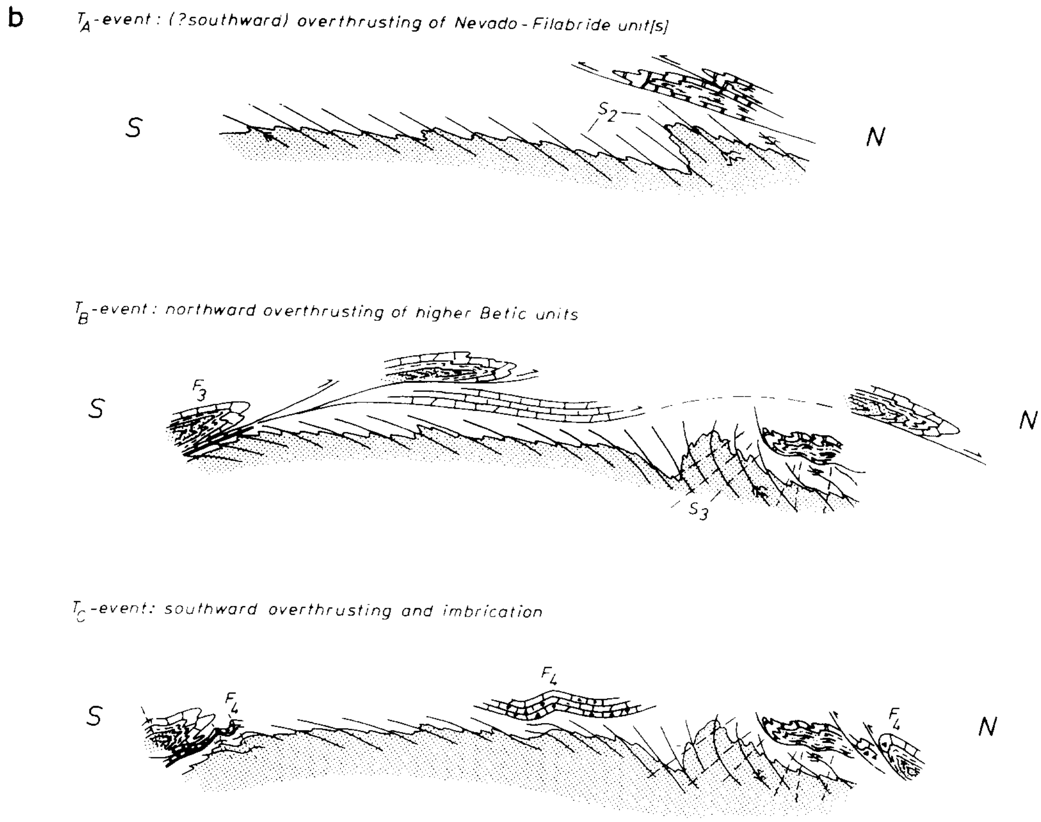
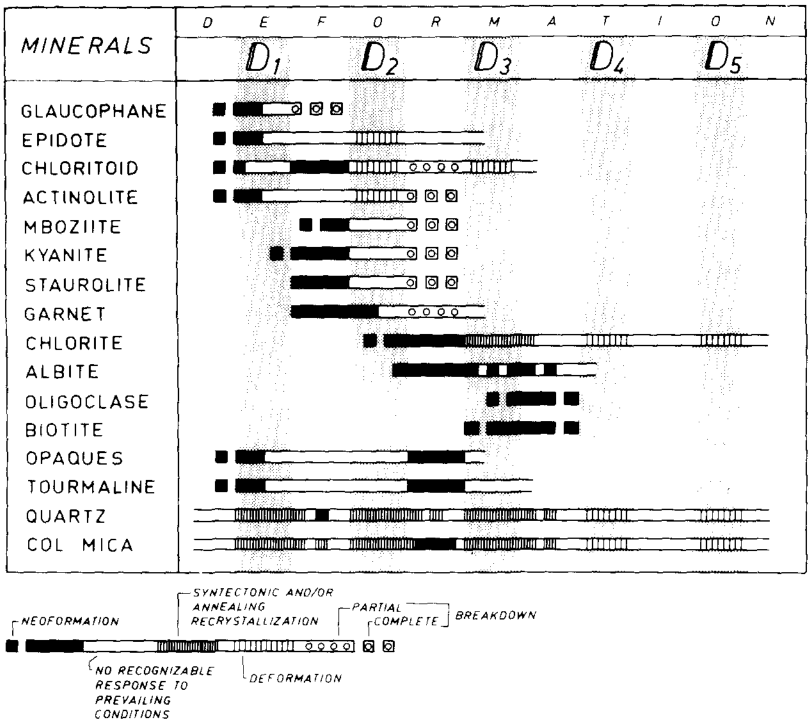


Fig. 8b. Schematics of the structural developments associated with the three thrusting events T_A–T_C referred to in Fig. 8a. T_A may have been established already 80–85 Ma B.P. (De Jong, 1987), T_B about 25–20 Ma B.P. (Weijermars, 1985a,b), and T_C is associated with Neogene updoming leading to surfacing of the Nevado-Filabrides between 12 and 7 Ma B.P. (Weijermars et al., 1985) (sketch after Vissers, 1981, fig. 56).

subduction zone can be inferred from the general trend of retrogressive metamorphism. More specifically, Fig. 9a and b illustrate that both the lower and upper Nevado–Filabrides evolve metamorphically from glaucophane schist facies (pre–syn D_1) via amphibolite–upper greenschist facies (post D_1–pre D_2) to lower greenschist facies (syn D_2–pre D_3) with a final peak of upper greenschist facies (syn D3). It is worth noting that some Tahal schists contain porphyroblasts of alkali-calciferous amphiboles with low silica and high aluminium content. This particular mineral named mboziite, later renamed as "taramite", was first described as a metamorphic variant of this rare amphibole by Linthout and Kieft (1970).

Figure 10 shows the tectono-metamorphic observations in a P,T,t-diagram and summarizes its geodynamic implications (Vissers, 1981). The lithostatic pressure increases about 25 MPa/km or 0.25 kbar/km and the average hypothetical geothermal gradient is 30°C/km depth. The odd combination of high pressure (> 12 kbar) with relatively low temperatures (< 600°C) can only be explained by subduction environments where rapid sinking of cold lithosphere can maintain low geothermal gradients of the order of 10°C/km at depth. A possible mechanism for obduction of Betic HP-LT rocks has been proposed by Platt (1986, 1987). Recent studies of the metamorphic history in the Higher Nevado–Filabride nappes suggest that the glaucophane-eclogite facies metamorphism at 10 kbar and 350°–425°C (De Jong and Bakker, 1988) essentially occurred as long ago as 80–85 Ma B.P. (De Jong, 1987). This age is a



b

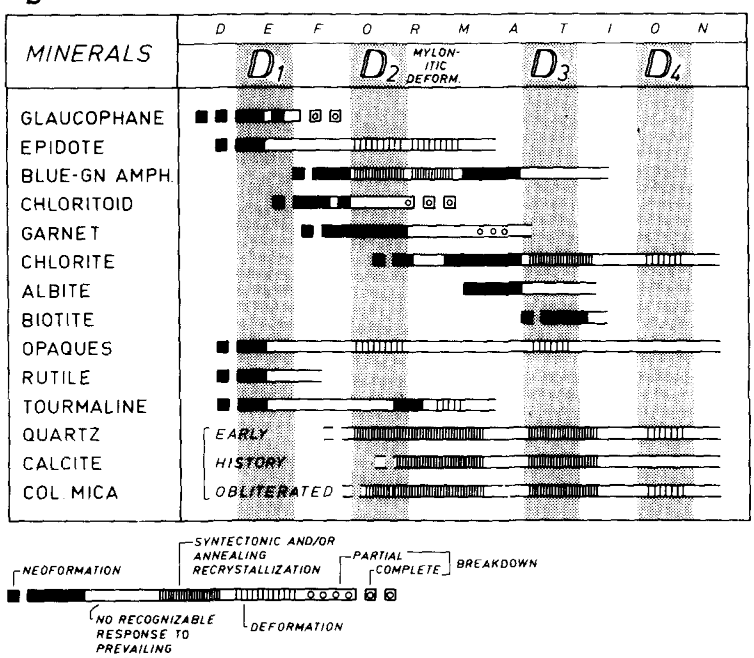


Fig. 9. Tectono-metamorphic maps showing details of the relative timing of the deformation and porphyroblastesis in (a) the Nevado-Lubrín unit, and (b) the Umbría nappe of the Higher Nevado–Filabrides. Four different generations of mineral assemblages characteristic for a particular metamorphic facies can be distinguished, i.e. glaucophane schist facies (glaucophane, epidote, chloritoid*, andalusite*), upper greenschist-amphibolite facies (blue-green amphibole, chloritoid, garnet, mboziite*, kyanite*, staurolite*), lower greenschist facies (albite, chlorite), and upper greenschist facies (biotite, oligoclase*). The minerals marked with an asterisk are only present in the top of the Nevado-Lubrín unit and are lacking in the Umbría nappe (from Vissers, 1981, figs. 37 and 42, respectively).

based on radiometric dating of metamorphic tourmalines syngenetic with D_1 in the sheared metagranite of Bédar (De Jong, 1987). This suggests that the continental collision implied by the HP-LT rocks would be substantially older than the 20–25 Ma nappe-shedding event from the bulge in the crust of the Alboran area (see section 8).

If the Nevado-Lubrín unit is autochthonous, the next question arising is: What was the palaeogeographic position of the Higher Nevado–Filabride nappes before their emplacement? Recent investigations suggest that they were involved in W- to NW-directed transport (De Jong and Bakker, 1988) and this would mean a palaeogeographic realm east of the Nevado-Lubrín rocks. In their initial geographic position, possibly also overlying Palaeozoic basement, rocks of the Higher Nevado–Filabrides were intruded by the Bédar granite 269 ± 6 Ma B.P. (Priem et al., 1966). Consequently, the initial sedimentary sequence of the Higher Nevado–Filabrides predate the Jurassic and could well be of Permo-Triassic age as commonly assumed on the basis of their lithological facies. The pre-Neogene palaeogeography of the W-Mediterranean has recently been discussed in detail by Mäkel (1988).

The Nevado–Filabride Complex yields two economically important rock types: (1) high quality marbles in the Higher Nevado–

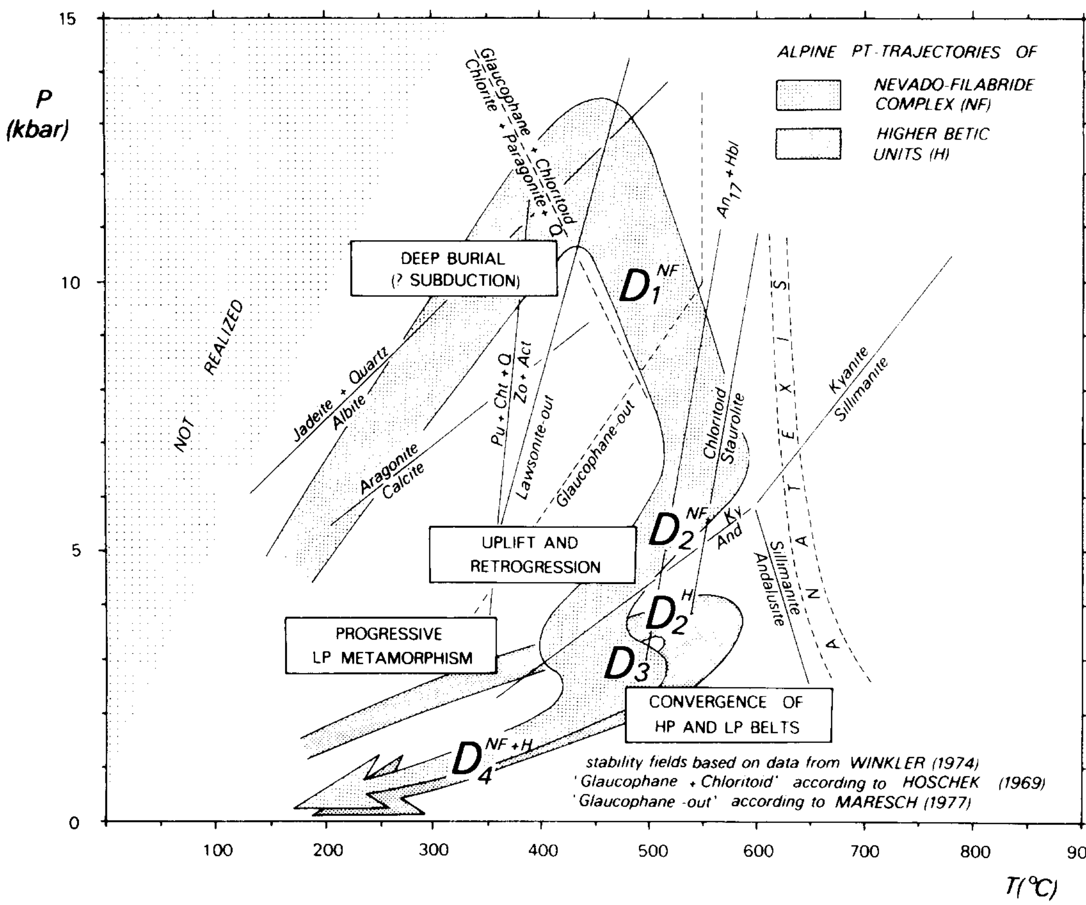


Fig. 10. Progressive metamorphic PT-trajectories for the rocks of the central Sierra de los Filabres. Light-shaded loop shows metamorphic evolution of the Nevado-Filabride Complex, dark-shaded loop pertains to the Higher Betic (Variegato and Tetica) units. The inferred geodynamics are indicated at the loops. A largely similar trend of metamorphic events has recently been compiled by Bakker et al. (1989). Critical minerals used are: *Act* – actinolite, An_{17} – oligoclase, *Cht* – chlorite, *Hbl* – hornblende, *Pu* – pumpellyite, *Q* – quartz, and *Zo* – zoisite (after Vissers, 1981, fig. 57).

Filabride nappes of the area around Macael (cf. Fig. 7a); and (2) Fe-deposits of Marquesado in two locations near Guadix, i.e. Alquife and Las Piletas. In 1987, the quarries of Macael together produced 600000 tons of marble involving a total turnover of 18 billion pesetas or 300 million DM (annual mining report). The Fe-reserve of the Marquesado mines is estimated at 512 million tons and the annual production is 2 million tons (Torres-Ruiz, 1983).

4.2. Higher Betic nappes

The Higher Betic nappes are thin thrust sheets, only up to several hundreds of meters thick, which were emplaced onto the Nevado–Filabrides moving roughly towards the N to NW (Aldaya, 1969; Platt, 1982; Behrmann and Platt, 1982; Platt et al., 1983; González-Lodeiro et al., 1984). The thrust masses can be grouped in at least two individual piles, i.e. the Alpujarride and Malaguide Complexes. The Alpujarrides were first termed so by Van Bemmelen (1927) and the Malaguides (or "Betic of Málaga") were named by Blumenthal (1927, 1930). The bulk of the exposed Higher Betic nappes are Alpujarride units, only overlain by Malaguides near Málaga and near the contact between the Betic and Subbetic Zone (Fig. 3). In the area of Almería province, Malaguide units have largely been removed by extensive erosion prior to the onset of Neogene sedimentation and only small outcrops are preserved in klippen (Fig. 11). Malaguide erosion products are abundant in the Older Neogene deposits (e.g. Aguilas Basin).

The Alpujarride nappes comprise low-grade metamorphic Triassic carbonate rocks (Van den Boogaard, 1966; Simon, 1966; Kozur et al., 1974), phyllites and quartzites (Egeler and Simon, 1969a,b) and locally they contain black greenschist facies schists (Vissers, 1981; Akkerman et al., 1980; Platt et al., 1983). Figure 12 shows a lithostratographic column

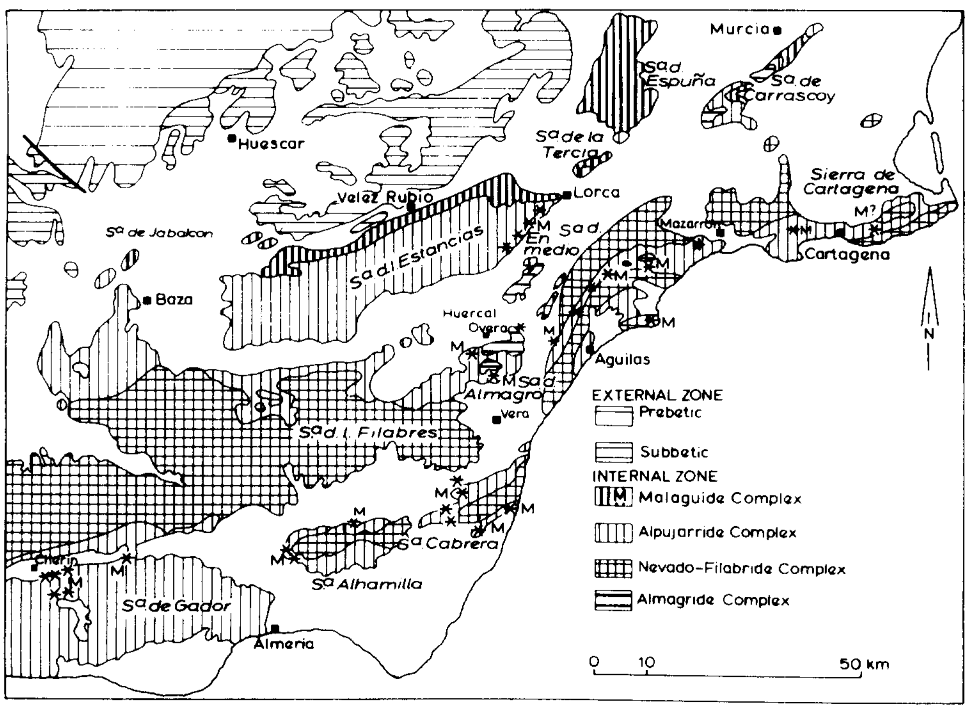


Fig. 11. Geotectonic map of the eastern Betic Zone showing the outcrops of Malaguide rocks. Most extensive exposures occur near Vélez Rubío and Lorca, directly south of the boundary between the Betic and Subbetic Zones. Small isolated klippen of Malaguide rocks occur throughout the Betic Zone as indicated by the starred locations (*M*) (after Mäkel, 1985, fig. 2-26).

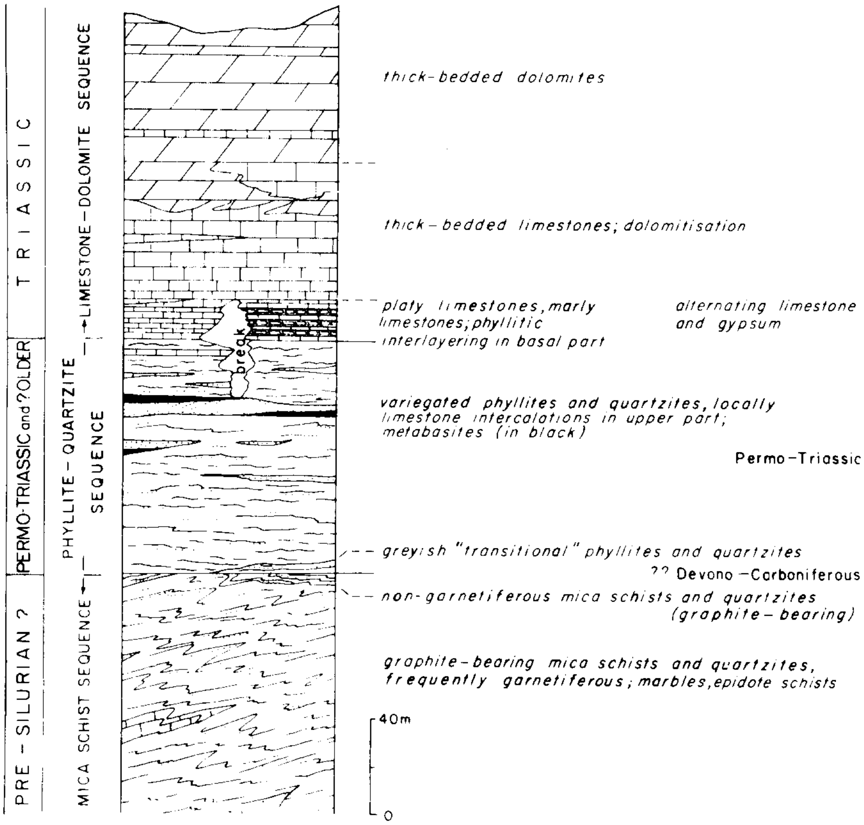


Fig. 12. Typical lithostratigraphic section of the tectonised Alpujarride Complex as exposed in the western Sierra Cabrera (from Rondeel, 1965, fig. 6).

for the Alpujarride rocks exposed in the Sierra Cabrera (Rondeel, 1965). The Malaguide nappes consist of non-metamorphic to very low-grade metamorphic carbonate rocks, sandstones, shales and conglomerates, ranging from Silurian to Oligocene in age (De Booy and Egeler, 1961; Soediono, 1971; Roep, 1972; Geel, 1973; Mäkel and Rondeel, 1979; Mäkel, 1985). Oligocene sedimentary rocks have been found in Malaguide klippen of the Vélez Rubio Region, near Huercal Overa and in the Sierra Cabrera (cf. Egeler and Simon, 1969a). Figure 13 illustrates the stratigraphy of the Malaguides.

The Almagride Complex, another Higher Betic nappe (Fig. 11), is thought to separate the Nevado–Filabrides and Alpujarrides in the Sierras Almagro, Enmedio, Carrascoy and Orihuela (Simon, 1963, 1964; Bodenhausen and Simon, 1965; Kampschuur, 1972). The lithostratigraphy of the Almagride Complex, which comprises only a single nappe (i.e. the Almagro unit), can be correlated with Triassic series of the Subbetic Zone (Besems and Simon, 1982). The inferred palaeontological ages range from Langobardian to Carnian, i.e. 250–225 Ma B.P. (Besems and Simon, 1982). A review of the distinction between Almagrides, Alpujarrides and Simon's (1963) discarded Ballabona-Cucharón Complex is given by Kampschuur (1972: pp. 22–24).

At present, the Alpujarrides in the Sierra de los Filabres and Las Estancias are interpreted as comprising only two tectonic units, i.e. the Almanzora-Tetica unit and Variegato-Coloraos-Partaloa unit (Simon, 1963; Bicker, 1966; De Vries and Zwaan, 1967; Akkerman et al., 1980; Vissers, 1981; Biermann, 1984). The Alpujarride has probably been studied in greatest detail in the Sierra Alhamilla by students of John Platt. The Sierra Alhamilla is a basement inlier with Alpujarrides enveloping a core of Nevado–Filabrides (Alhamilla unit) (Fig. 14). It had been studied previously (Hetzel, 1923; Westerveld, 1929; Weppe and Jacquin, 1966; Jacquin, 1970), but the presence of a north-closing isoclinal fold nappe was not recognized until the early 1980's

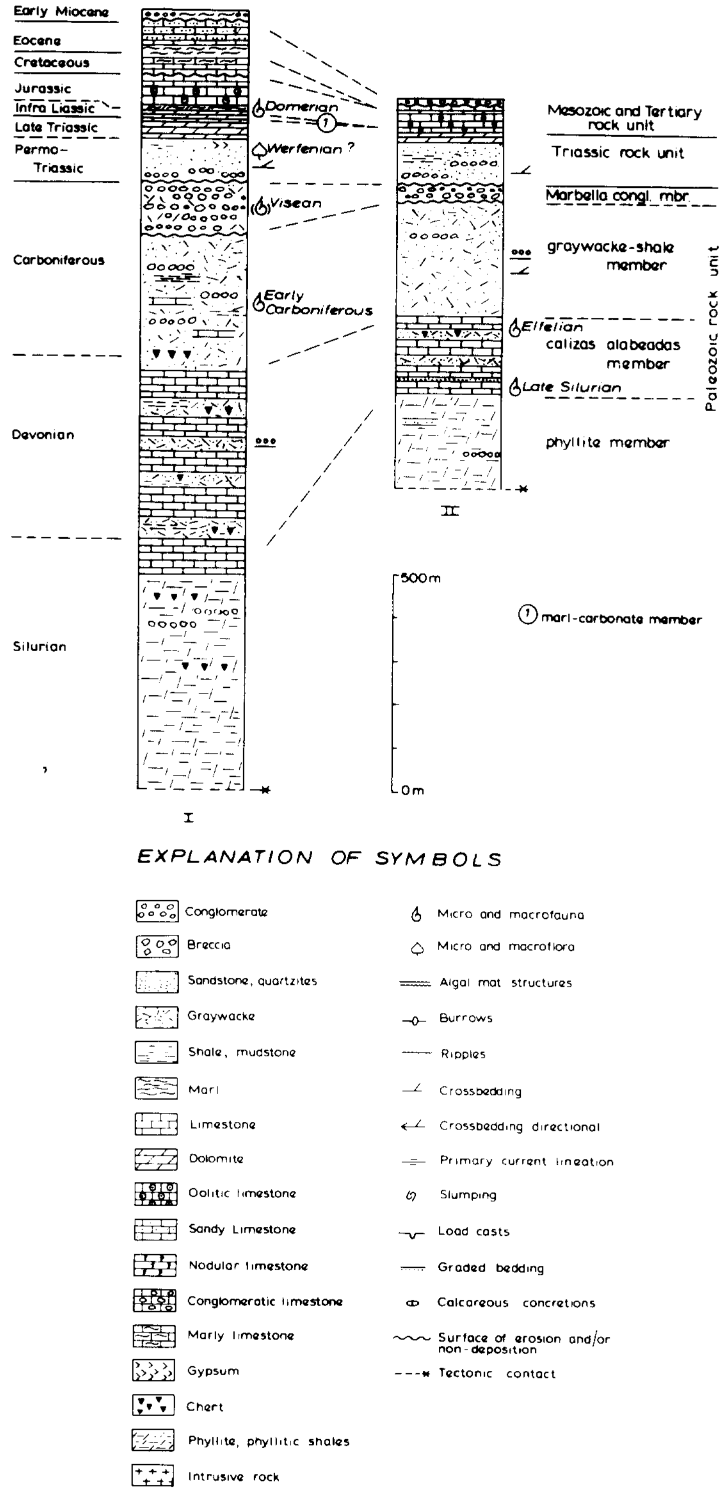


Fig. 13. Stratigraphic columns of the Malaguide Complex illustrating the sequence as exposed near Málaga (I) and Marbella (II), respectively (from Makel, 1985, figs. 2-21 and 2-1).

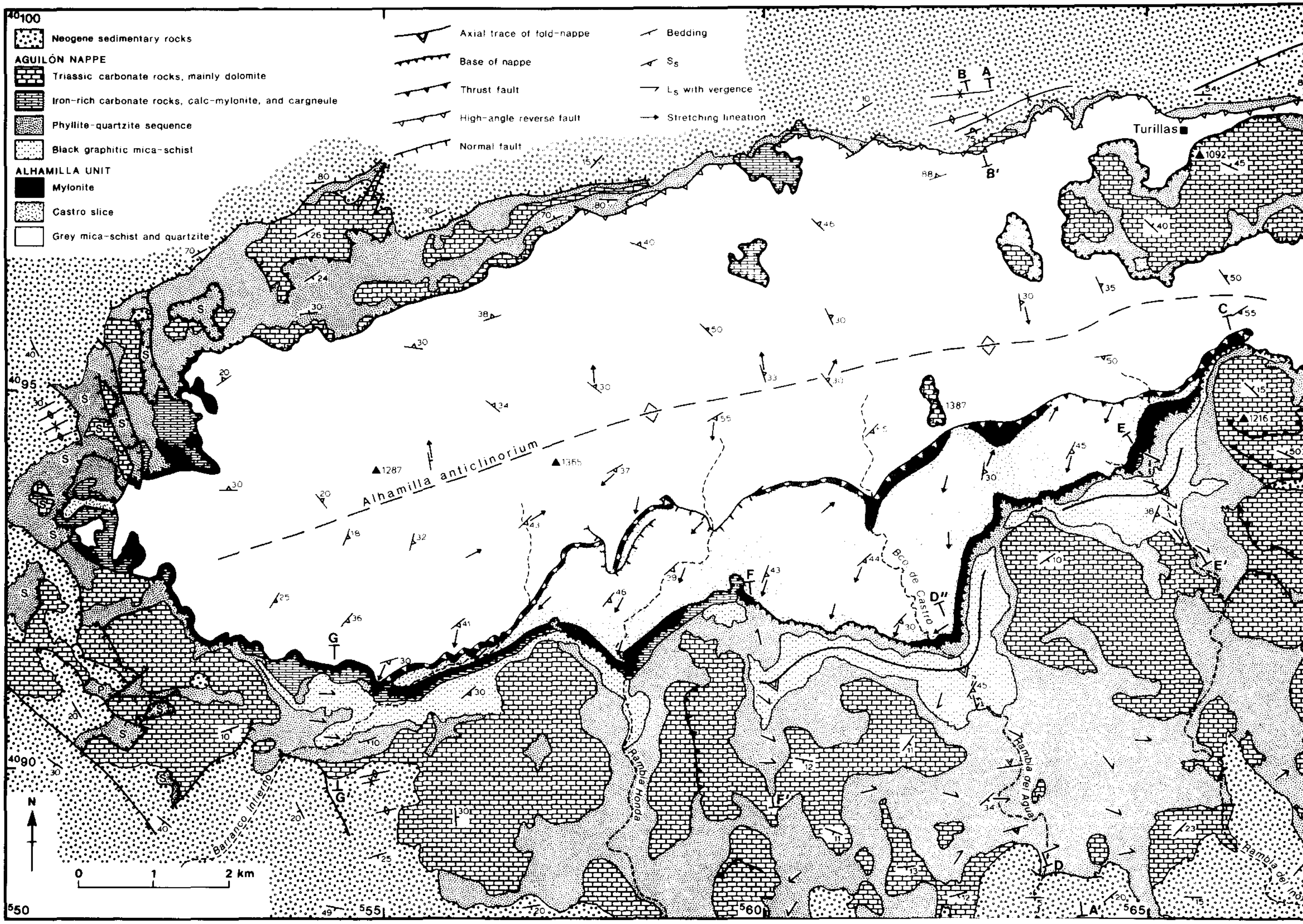


Fig. 14. Geological map of the western and central Sierra Alhamilla. Representative orientations of the main-phase foliation (Ss) and intersection lineation of minor fold axes (Ls) are shown in the Aguilón fold nappe. The fold-vergence or cleavage-vergence direction (cf. Weijermars, 1985c,d) is indicated by the half-barb so that a N-vergent minor fold has the half-barb on the north side of the arrow. Faults have teeth or ticks on the hanging wall. Minor units overlying the Aguilón nappe are the Savin (*S*) and Las Palmas (*P*) units. Sections given in Fig. 15 are indicated as A–A′, B–B′, etc. (from Platt et al., 1983, fig. 3).

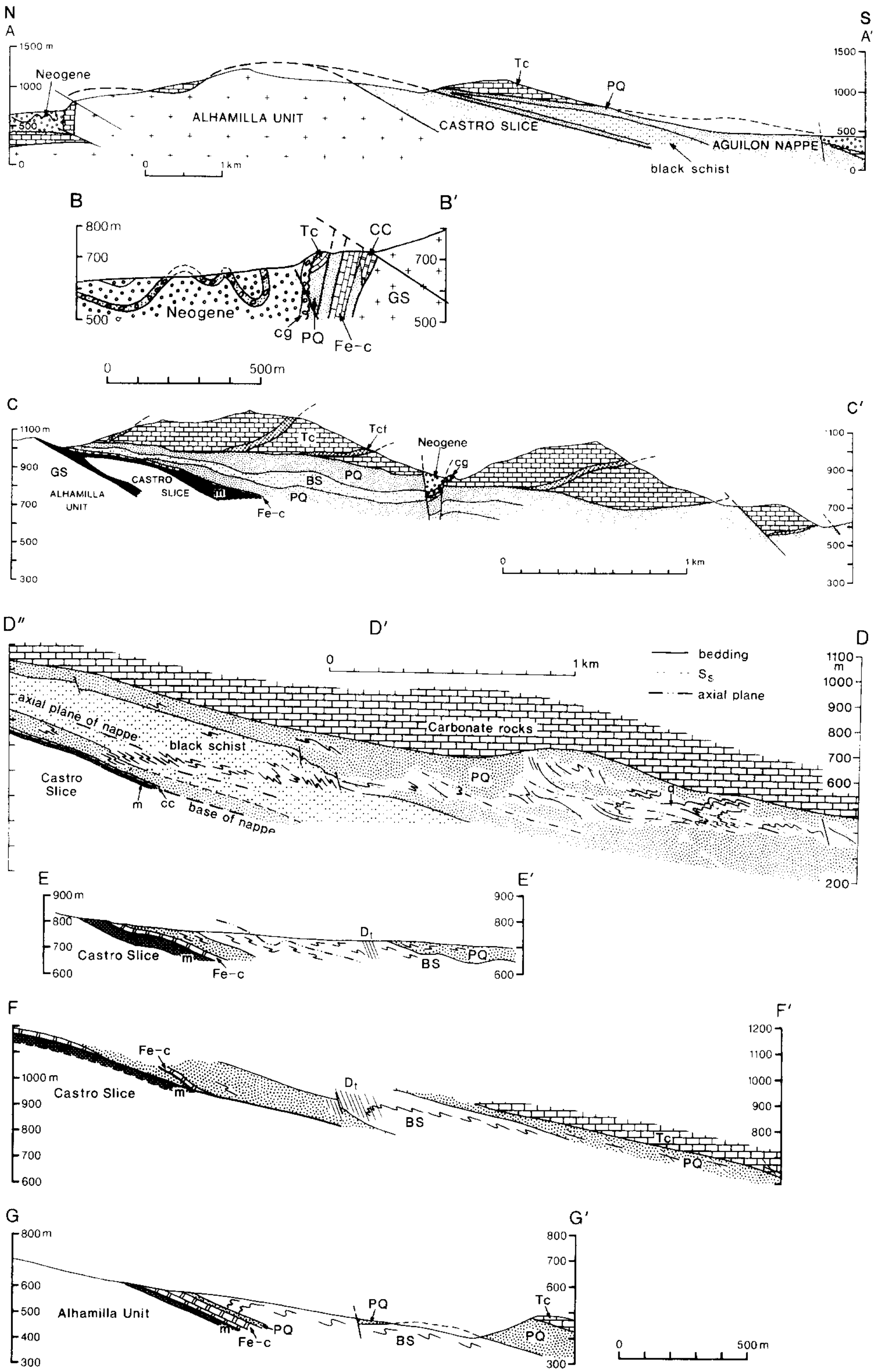


Fig. 15. Sections illustrating the geometry and dimensions of the Aguilón fold nappe. The positions of A–A′ to G–G′ and a legend to the ornament are given in Fig. 14. Annotations used are *Tcf* = basal fossiliferous marl and limestone layer, *Tc* = Triassic carbonate, *Fe–c* = iron-rich carbonate rock, *cc* = cargneule and calcmylonite, *m* = mylonite, *PQ* = phyllite and quartzite, *BS* = black mica schist with garnet and staurolite, *GS* = grey mica schist of Nevado-Filabride Alhamilla unit. Locations of the sections are: A–A′: section over Turrillas, central Sierra Alhamilla; B–B′: detail of Turrillas section; C–C′: secton La Mina east of Rambla del Inox; D–D′: section of Rambla del Agua; E–E′: section of Rambla del Inox; F–F′: section of Barranco Martinez; G–G′: section of Barranco del Infierno, near Baños (from Platt et al., 1983, figs. 4, 5, 7, 14 and 15).

(Platt, 1982; Platt et al., 1983; Weijermars, 1985c).

The Aguilón nappe has a core of black, garnet-bearing mica schist, locally boudinaged and enveloped by multicoloured phyllites, quartzites and carbonates on either flank of the recumbent fold (Platt et al., 1983). Cross-bedding in quartzites of the upper limb indicates that this limb is younging upwards so that the carbonates are the latest deposits affected by the nappe emplacement (Platt, 1982). The basal package of the carbonates contains abundant bivalves (*Lyriomyophoria betica*; Hirsch, 1966) and has been dated as Middle Triassic (Ladinian, i.e. 230–235 Ma B.P.) using ostracods (Kozur et al., 1974). The carbonates in the lower limb are extremely thinned and usually transformed into dolomite mylonite or breccia or have been heavily enriched in iron to form siderite marble. Figure 15 gives a series of typical sections through the fold nappe, illustrating its geometry and size. Field aspects of the Aguilón nappe are illustrated in Plate 5.

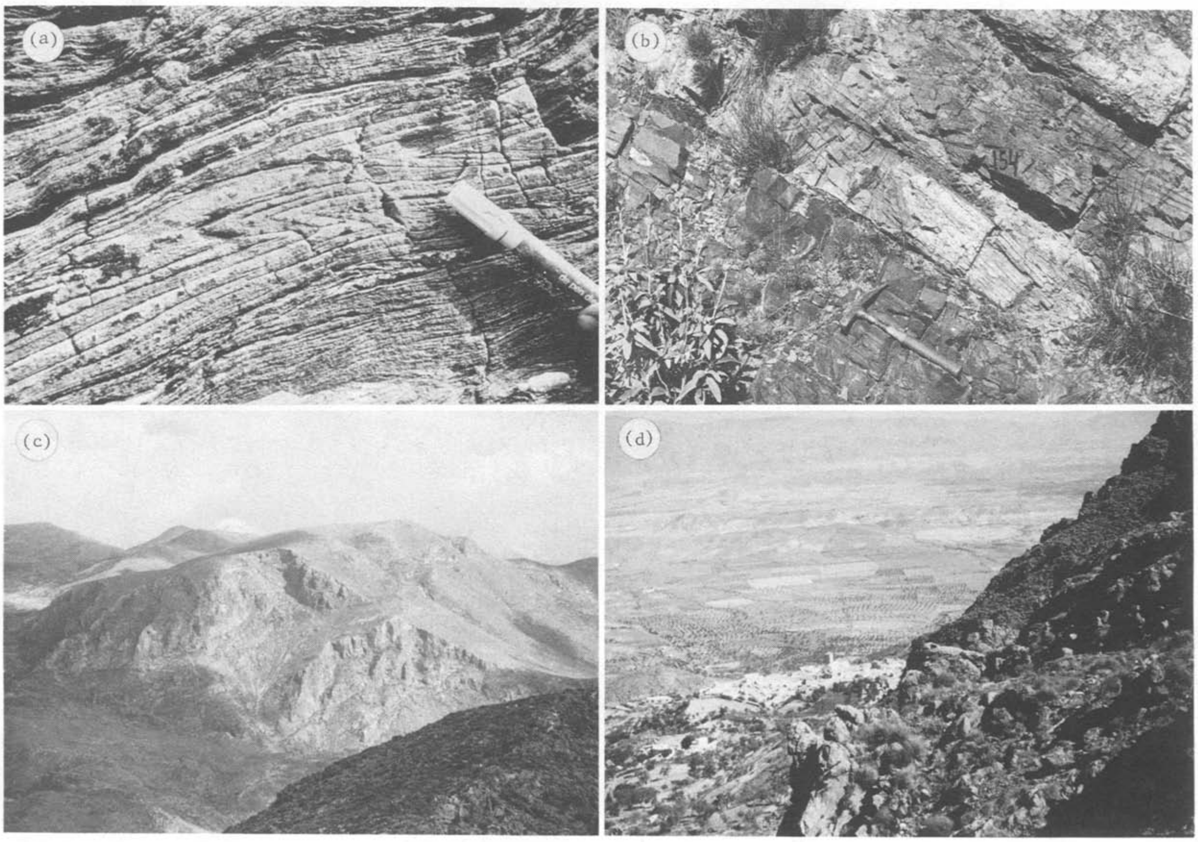


Plate 5. Higher Betics, Sierra Alhamilla. (a) Aguilón fold nappe. Quartzite intercalation in the black garnet-bearing mica-schist, showing S-verging minor fold in the lower limb of the Aguilón fold nappe formed during nappe emplacement. Rambla del Inox, Sierra Alhamilla. (b) Aguilón fold nappe. Bedding and cross-bedding preserved in quartzite horizon of the variegated phyllite-quartzite sequence in the upper limb of the Aguilón fold-nappe. Rambla del Inox, west of Loma Oscura (exposure 154 of Weijermars, 1981). (c) Aguilón fold nappe. Triassic carbonate sequence, comprising a N-dipping piggy-back thrust, in the upper limb of the Aguilón fold nappe. Cerro de los Caballos (el. 1051 m), east wall of the Rambla del Inox, Sierra Alhamilla. (d) Alpujarride carbonates in the upper limb of the Aguilón fold nappe near oversteepened northern edge of Cerro Minuto, Sierra Alhamilla. The edge has provided material for ancient landslides on which the village of Turrillas (el. 848 m) – visible at the foot of Cerro Minuto – was built. Incipient surface movement of Cerro Munito is indicated by open cracks, up to 20 m deep. The underlying impermeable phyllites with their subhorizontal foliation presumably act as a lubricant beneath the carbonates and allow creep down the topographical slope. Cerro Minuto (el. 1092 m), looking north.

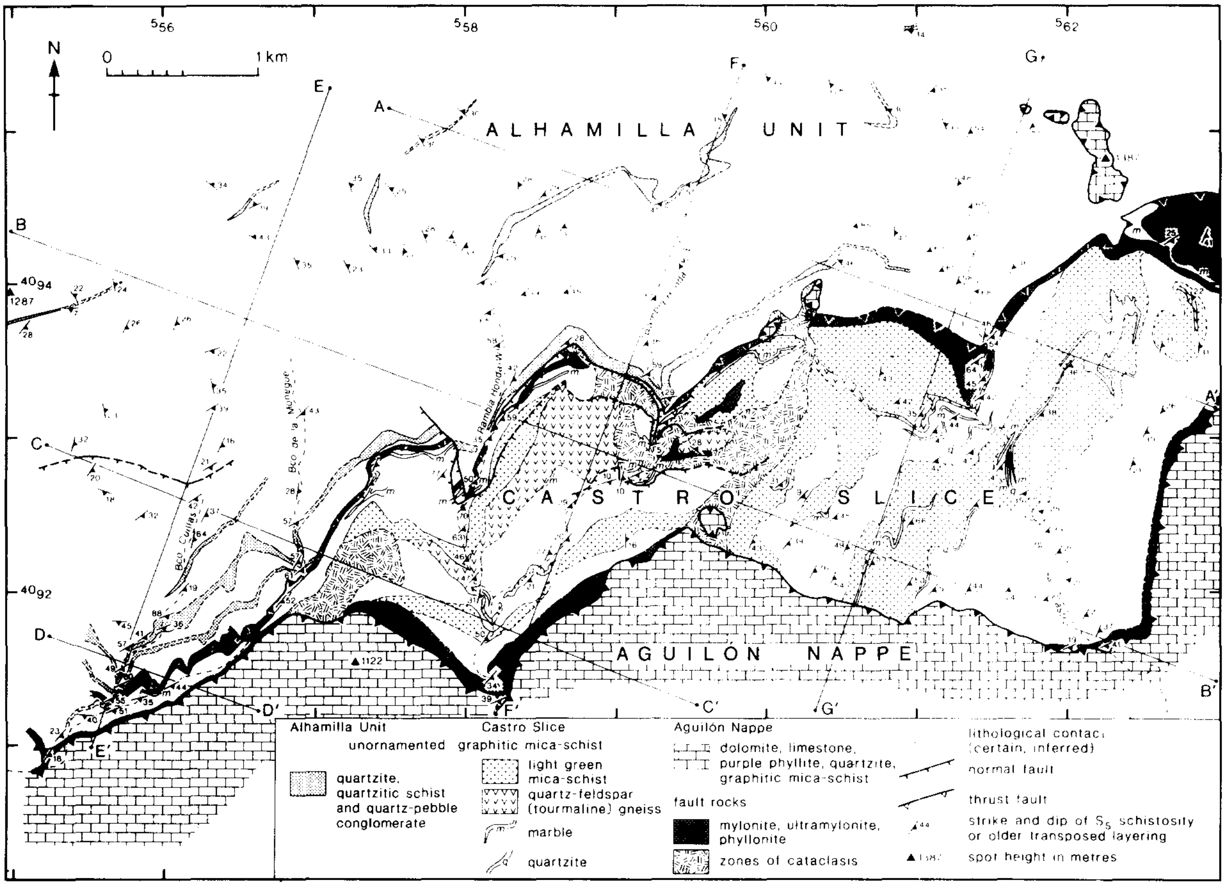


Fig. 16. Detailed geological map of the Castro-slice, a zone of intensive deformation overlying the Alhamilla unit and underlying the Aguilón fold nappe (from Platt and Behrmann, 1986, fig. 4). The outlined sections A–A' to G–G' are given in the original study of Platt and Behrmann (1986).

Mylonites, cataclasites and extensional crenulation cleavages are abundantly developed in the Nevado–Filabride Castro slice underneath the Aguilón fold nappe, coeval with the emplacement of the overriding nappe (Fig. 16). These structures have been documented in great detail elsewhere (Platt and Vissers, 1980; Konert and Van den Eeckhout, 1983; Van den Eeckhout and Konert, 1984; Weijermars and Rondeel, 1985; Behrmann, 1984a,b, 1985, 1987; Platt and Behrmann, 1986). The internal geometry of the nappe and oriented quartz *c*-axes in the mylonitic footwall rocks both indicate that the spreading which emplaced the Aguilón nappe involved northward transport (Behrmann and Platt, 1982; Platt et al., 1983). Geothermometry of a marble specimen from the mylonitic carpet indicates that ambient temperatures during flow were in the order of 300 °C (Behrmann, 1983). Greenschist-facies metamorphism, syngenetically developed with minor structures in the Aguilón fold-nappe, independently suggests that nappe emplacement occurred at a depth of 10–15 km by gravity spreading (Platt et al., 1983).

Syngenetic Pb-, Zn-, Ag-, and Fe-mineralisations have been mined in Alpujarride carbonates throughout the Betic Zone (e.g. near Mazarrón in the Sierra de Cartagena, Herrerías in the Sierra de los Filabres, Lucainena de las Torres and Mina Coto Laisquez in the Sierra Alhamilla). Lead–zinc ores of the La Uníon mining district in Alpujarride carbonates of the Sierra de Cartagena have been interpreted as subvolcanic hydrothermal ore deposits related to the Neogene volcanism (Oen et al., 1975; Kager, 1980). The magnetite deposits of Mina de la Concepcíon occur in skarns of the metamorphic contact aureole of the peridotite massif in the Sierra de Bermeja, western Betic Zone (Westerhof, 1975). Exploitation ceased in 1973 and the estimated Fe-reserve is 12 million tons (mining reports).

5. NEOGENE SEDIMENTARY COVER

Continuous research on the Tertiary sedimentary rocks in southeast Spain since De Verneuil and Colomb (1856) has provided a detailed picture of depositional age and environment. Monreal (1878) specified that the Tertiary deposits of De Verneuil and Colomb (1856) were of Miocene and Pliocene age. The discovery of foraminifers in marls near Garrucha and whale spines near Cuevas del Almanzora, indeed confirmed a Pliocene age (Schrodt, 1890). Early contributions were further made by Fircks (1906), Sánchez Lozano (1918), Hetzel (1923), Guardiola and De Sierra (1925), Gignoux and Fallot (1927), Durand Delga and Magné (1958), Trigueros and Navarro (1963) and Iaccarino et al. (1975).

However, major advances were made principally due to several excellent Ph.D.-studies carried out over the past twenty years (Rondeel, 1965; Völk, 1967; Montenat, 1973; Ott d'Estevou, 1980; Postma, 1983; Dronkert, 1985; Alvado, 1986; De la Chapelle, 1988). Continuity of research was due to efforts by Christian Montenat of the "Institute Géologique Albert de Lapparent" (IGAL), Paris (Montenat et al., 1975, 1976, 1980, 1987, 1988), and studies conducted and supervised by Tom Roep and Dirk Beets of Amsterdam University (Roep and Beets, 1977; Roep and Van Harten, 1979; Roep et al., 1979; Postma and Roep, 1985; Weijermars et al., 1985). The areal distribution of the Neogene sedimentary rocks and volcanic rocks is compiled in the geological map of Fig. 17. The sedimentary rocks range in age from Burdigalian to Quaternary (i.e. 22 Ma B.P. and younger). Most parts of the area have now been mapped in great detail and the map of Fig. 18 illustrates the extension of lithological units in the Neogene basins of Tabernas, Sorbas, Vera and Almería. These will be discussed in detail below.

It has been customary to subdivide the Neogene cover into Older Neogene (Burdigalian and Serravallian) and Younger Neogene (Tortonian–Pliocene), according to Völk and Rondeel (1964). The major argument for this subdivision was the sudden change from Alpujarride to Nevado–Filabride detritus. Older Neogene sedimentary rocks outcrop in steeply dipping cuestas along the northern margin of the Sierra Cabrera (Fig. 18) and are made up of conglomerates with pelitic intercalations with a total thickness of about 500 m. The Older Neogene (Umbria and Mofar Formations) is generally exposed over relatively small areas and for a more detailed discussion the reader is referred to Rondeel (1965) and Völk (1967). The stratigraphic subdivision adopted here was principally established by Völk and Rondeel (1964; Rondeel, 1965; Völk, 1967), Ruegg (1964) and Dronkert (1976). Attempts to corellate the various stratigraphic schemes of different authors have been published by Dronkert (1985: table 1.1, p. 182) and Kleverlaan (1989b: table 2, p. 64).

The Younger Neogene, discussed in more detail below, is characterised by three major transgressive cycles of the following ages: (1) Serravallian–Tortonian, (2) Messinian and (3) Pliocene – in decreasing order of importance. Relief existed during all depositional episodes as may be inferred from the spatial distribution of the exposed Neogene sedimentary rocks (Fig. 18). Plate 6 shows field aspects of the Serravallian–Tortonian Chozas Formation in the Tabernas Basin. Plate 7 illustrates the Azagador, Abad and Cantera Members of the Early Messinian Turre Formation. The overlying Caños Formation, partially recording the Messinian Salinity Crisis, comprises the Yesares Member (Plate 8), Sorbas and Zorreras Members (Plate 9).

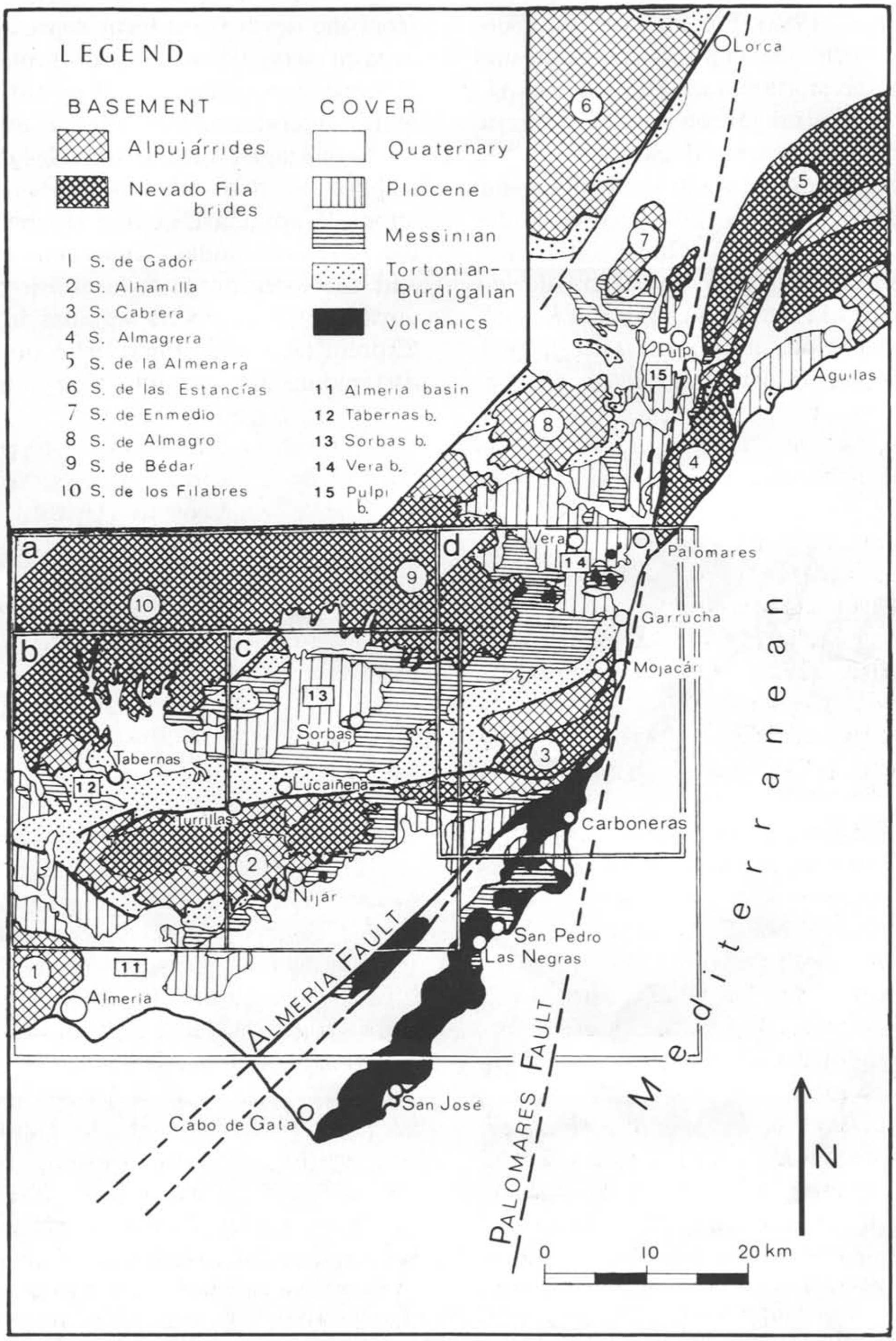


Fig. 17. Geological overview of the Neogene basins of southeastern Spain. The geographical names of the major basins and ranges (sierras) are indicated by numbers given in the legend. The outlines of detailed compilation maps are labelled a–d. Map a is given in Fig. 18 and is based on, a.o., the previous compilation maps b and c (Weijermars et al., 1985, fig. 2) and d (Weijermars, 1987a, fig. 6).

The Neogene sedimentation in southeast Spain is strongly controlled by coeval tectonism (e.g. Weijermars et al., 1985; Montenat et al., 1988). For example, the (Serravallian–) Tortonian deposits in the basins directly to the north and south of the Alhamilla–Cabrera range consists of a green marl sequence, up to 1 km thick, with intercalations of seismites, turbidite sandstones and conglomerates (Fig. 19). The turbiditic intercalations were supplied by a huge, submarine fan system extending from the Sierra de los Filabres (Fig. 18) towards the south and southwest, covering an area of at least 1000 km^2 (Kleverlaan, 1987, 1989a,b).

The Sierra Alhamilla was only a shallow regional basement high during the Tortonian transgression and the submarine fan influx from the north filled local depressions and is even preserved along the south margin of the modern Sierra Alhamilla (Fig. 18). Turbidites with a microfauna indicative of Serravallian–Tortonian age (Weijermars, 1981), have been trapped in two different graben structures about 25 km apart, i.e. the Huebro graben in the Sierra Alhamilla (Weijermars et al., 1985) and the intricate fault zone directly to the north of Minas de las Aguillas in the Sierra Cabrera (Rondeel, 1965). The slope instability feeding the turbidite fan complex may

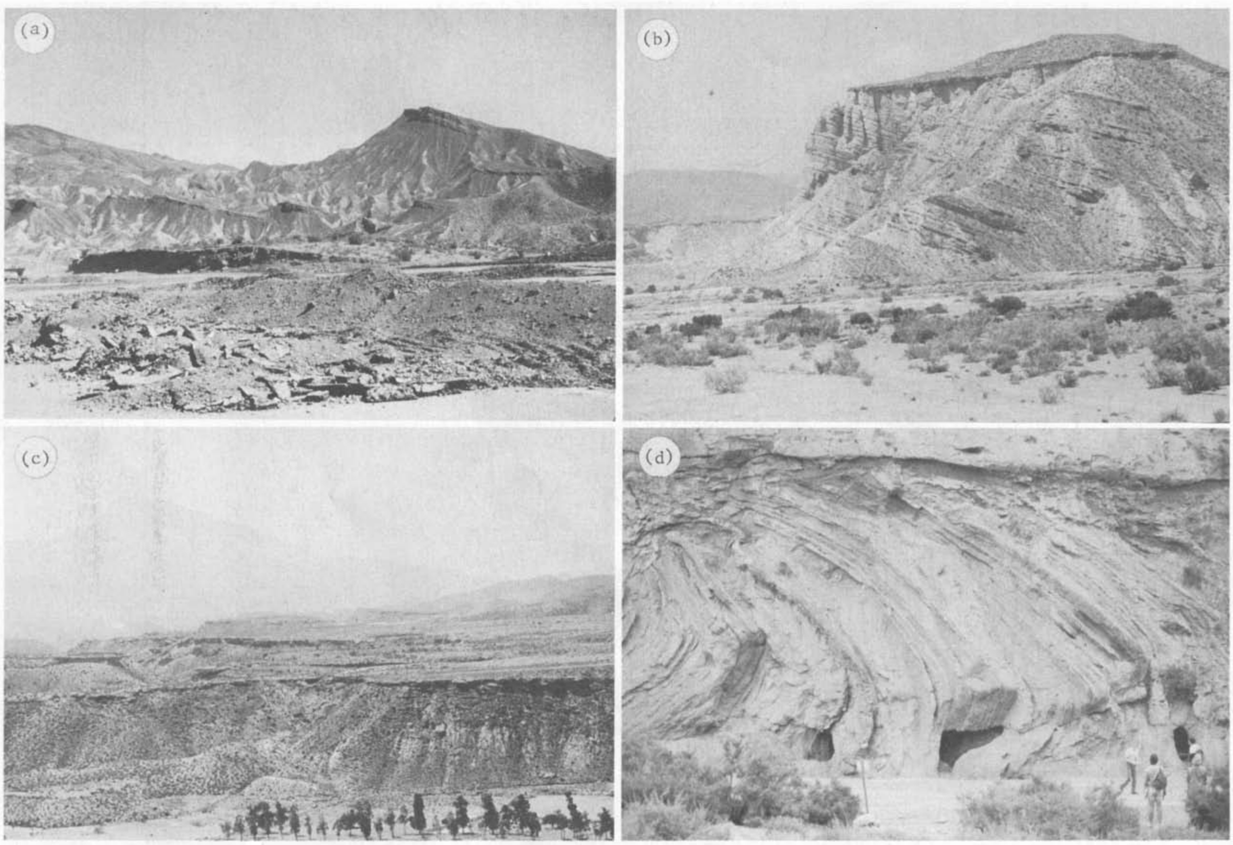


Plate 6. Serravallian–Tortonian sedimentary rocks; Chozas Formation. (a) Tabernas Basin, El Alfaro (el. 742 m). The Serravallian–Tortonian turbidite sequence of the Tabernas is tilted by upheaval of the Sierra Alhamilla, which is to the left, west-side of the picture. The bulk of EL Alfaro is made up of blue mudstones capped by a conglomeratic megabed. Looking south from Bar El Alfaro, road N-340. (b) Tabernas Basin, El Buho (el. 572 m). Outer fan deposits of Tabernas submarine fan complex. Looking north. Sandstone beds in marl dip to the southeast. El Buho is capped by Plio-Quarternary peneplain conglomerate. (c) Tabernas Basin. Looking SSW from El Buho towards El Alfaro (8 km). Note the carpet of peneplain conglomerates capping the Tortonian sedimentary rocks. Peneplain height at 400 m level. (d) Rambla de Tabernas, southwall. Large synsedimentary slump in outer fan turbidites. Unconformably covered by subhorizontal mudstone bed of fan complex.

have been enhanced by further uplift of the Sierra de los Filabres during the Serravallian–Tortonian period. The palaeobathymetry of the Tortonian sea in the Tabernas–Sorbas area has been estimated to lie in the range 150–350 m on the basis of sponges (Ott d'Estevou and Termier, 1978). The depositional depth of Late Tortonian (Early Messinian) turbidites in the Vera Basin has been estimated to be about 500 m, on the basis of ichnofossils (Montenat and Seilacher, 1978).

The Messinian transgression was preceeded by folding of the turbidites in the Tabernas Basin, around NE–SW trending fold axes (Weijermars et al., 1985). Late Tortonian fold structures are truncated in the NE by an erosional unconformity at the base of the Messinian succession in the Sorbas Basin (Fig.

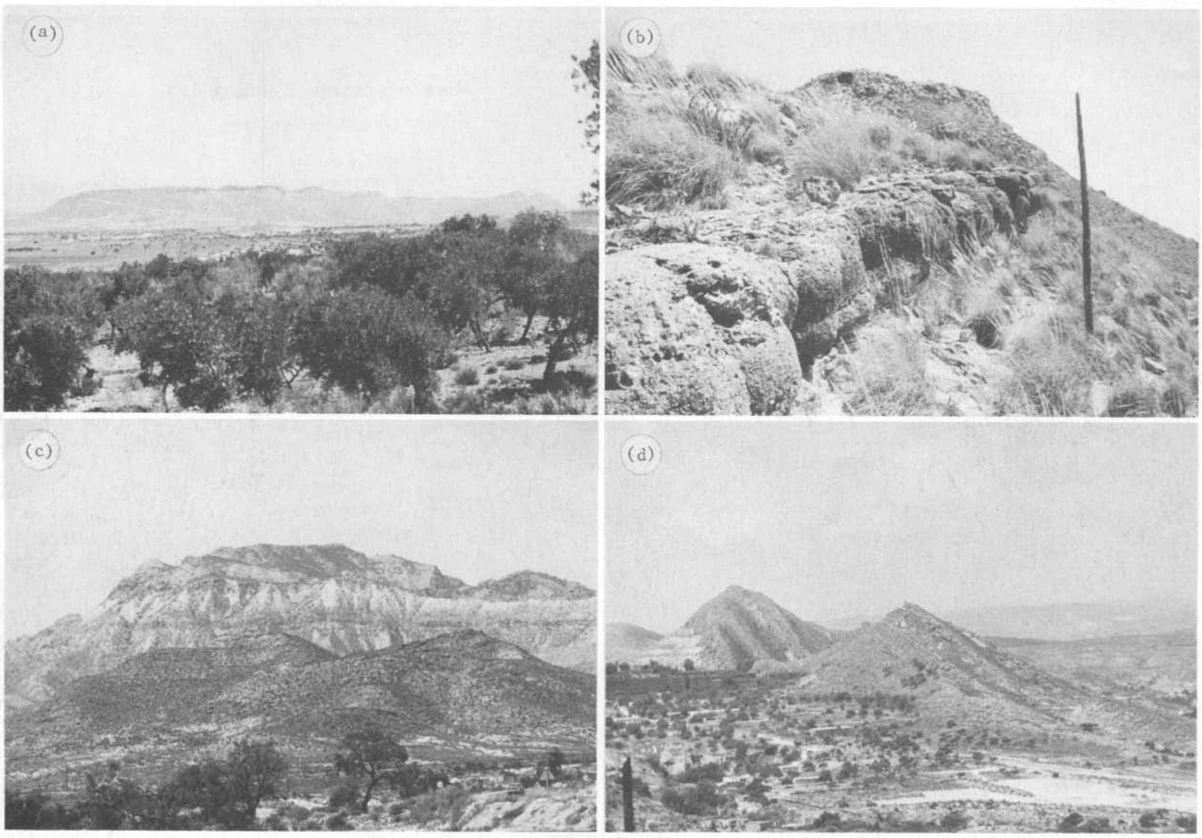


Plate 7. Early Messinian sedimentary rocks; Turre Formation. (a) Sorbas Basin. Cuesta of Azagador basal conglomerate discordantly capping Tortonian Chozas Formation on Loma de la Cumbro (el. 685 m) north of Lucaiñena de las Torres. Looking towards the northeast from road to Turrillas. (b) Sorbas Basin. Azagador basal conglomerate of the Turre Formation on eastward end of Loma de la Cumbro, Los Llanos (el. 682 m). Looking east. (c) Sorbas Basin. Panoramic view from Peñas Negras on the Sorbas to Carboneras road, looking NW, towards El Cerron de Hueli (el. 664 m). Hills in the foreground are Older Neogene of the Chozas Formation. Dark horizon halfway up the Hueli cliff is Azagador conglomerate. The Azagador is overlain by Abad marls. The Hueli is capped by gypsum of the Yesares Member of the Caños Formation. Remnants of Cantera reefs occur sandwiched between Abad marls and Yesares gypsum in the west part of the Hueli. (d) Vera Basin, looking towards the west from Mojácar village at the eastern end of the Sierra Cabrera. The two hills in the foreground (Cerros de Joancho and de los Caballones, el. ca. 100 m) comprise steeply tilted, NNW-dipping strata of the Azagador conglomerate, marking the boundary between the Tortonian and Messinian. The tilt is due to the northward transportation of the Sierra Cabrera–Alhamilla antiform by shear drag on the walls of the sinistral Palomares Fault which trace surfaces 3 km to the east. The moderate relief at the horizon is formed by the Sierra de Bédar, the eastern continuation of the Sierra de los Filabres, comprising exposures of the 270 Ma old Bédar granites.

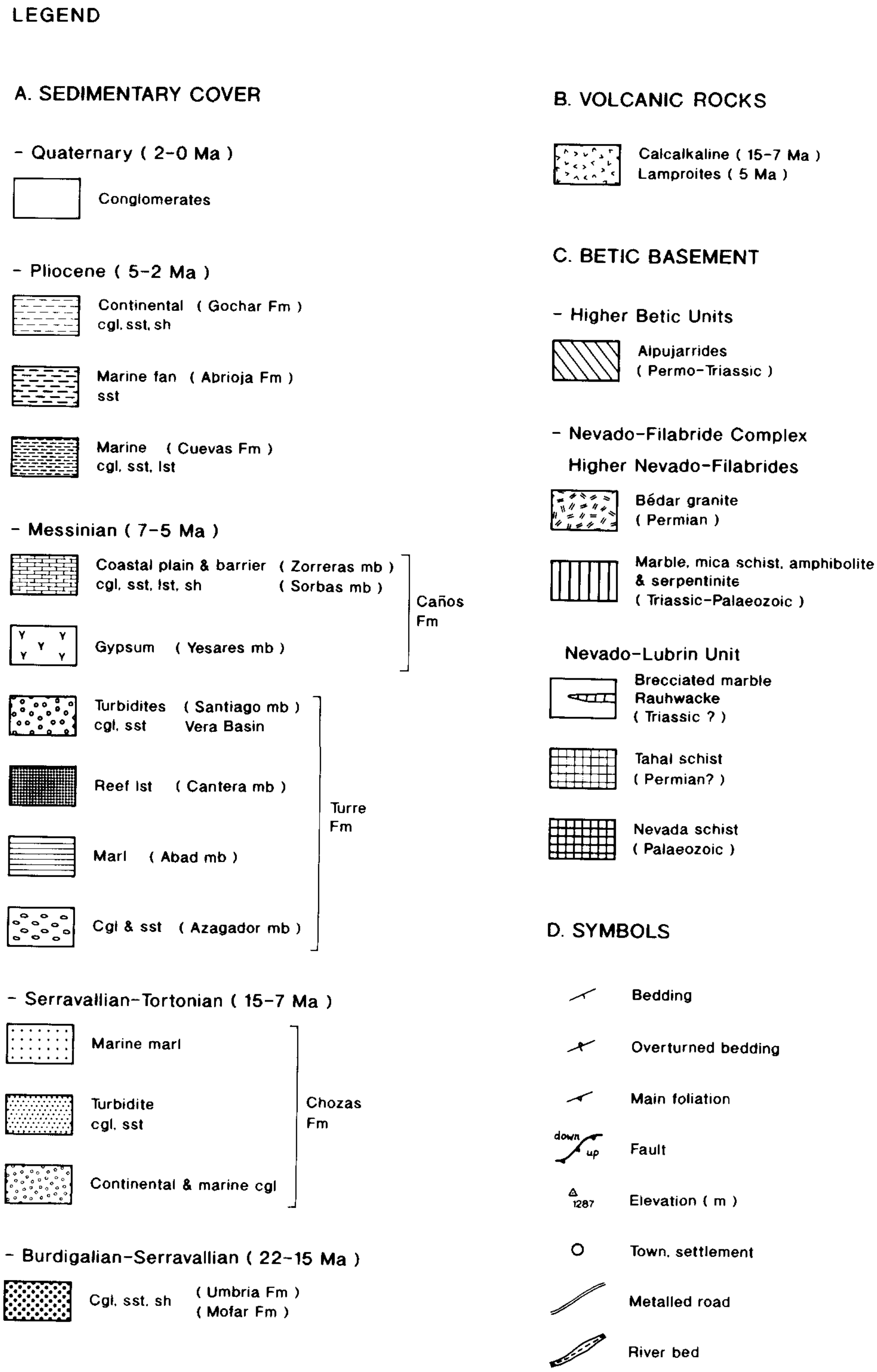


Fig. 18. Detailed geological map of the Almería–Tabernas, Sorbas and Vera Basins, including the metamorphic basement of the Sierra Alhamilla–Cabrera. The location of the map is outlined on Fig. 17. Topographic height of Messinian reef limestones and triangulation points are indicated in meters. The Sierra Cabrera's structural trend changes from E–W in the west to NNE–SSW near the Palomares Fault at its eastern termination. The strikes of Older Neogene (Burdigalian–Serravallian), Tortonian, and Messinian beds along the southern margin of the Vera Basin show a similar deflection. Gently dipping Lower Pliocene deposits of the Cuevas Formation unconformably overlying the steep to overturned older strata (e.g. 1 km SW of Garrucha) suggest that major deflection occurred early in the Early Pliocene. Compiled on the basis of Vissers (1981, Fig. 5 in this study), Bordet (1985, enclosure map), Weijermars et al. (1985, fig. 2), Weijermars (1987a, fig. 6) and De la Chapelle (1988, enclosure map).

20). The folding was accompanied by reverse faulting such as to accommodate shortening of the Tabernas Basin (Fig. 21). These Tortonian folds are also truncated by a reverse fault which separates the Tabernas Basin and the metamorphic basement of the Sierra Alhamilla (i.e. the Northern Boundary Fault – NBF). The NBF is a Late Tortonian fault genetically associated with the formation of the anticlinal in the Alhamilla–Cabrera range.

The shortening of the Tabernas Basin can best be explained by the northward displacement of the Sierra Alhamilla relative to the Sierra de Gádor. These mountain ranges are separated by a normal fault with up to 4 km sinistral, strike–slip displacement as can be inferred from the off-set of the anticlinal axis of the Sierra Alhamilla relative to that of the Sierra de Gádor (Fig. 20). The existence of this important fault structure has been noted previously (Hetzel, 1923; Jansen, 1936; Postma, 1983), but it remained nameless hitherto. I propose to term it the Baños Fault, after the thermal springs of Baños. The fault

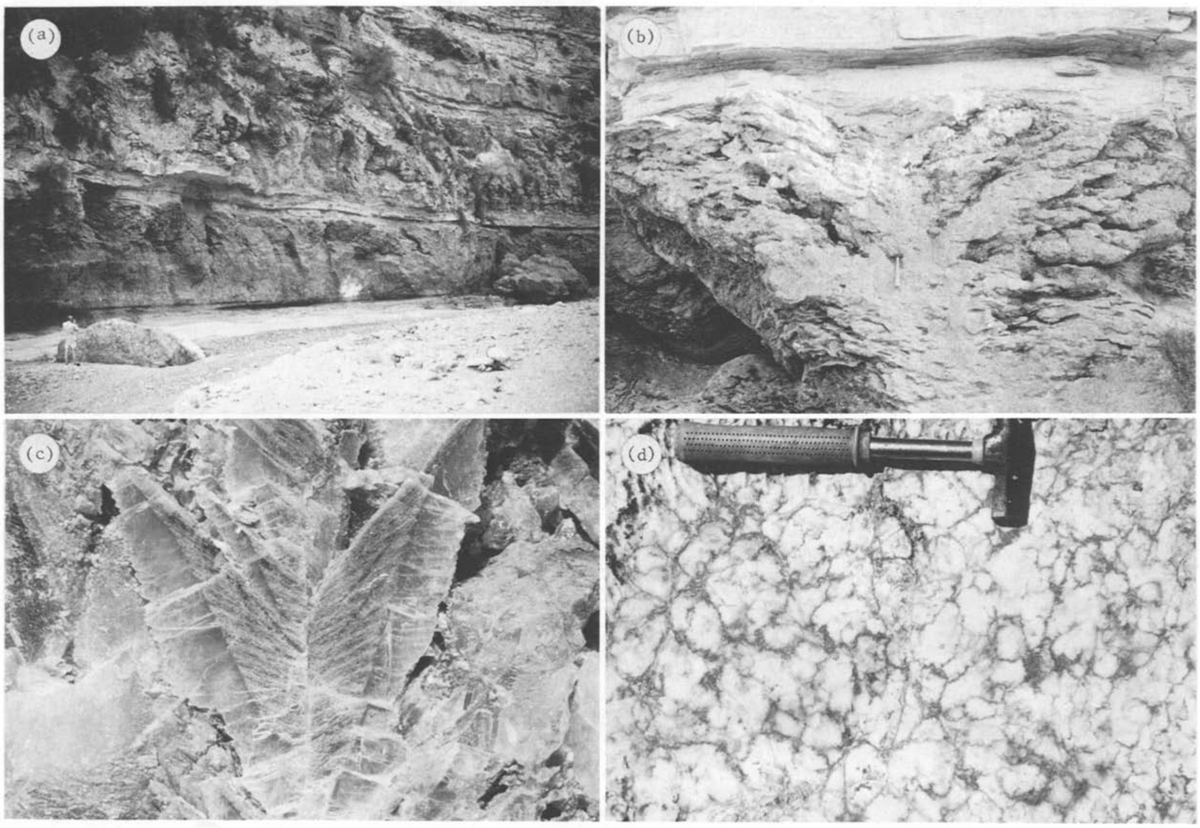


Plate 8. Messinian Salinity Crisis; Yesares Member, Caños Formation. (a) Sorbas Basin. Overview of Río de Aguas section through upper part of the Yesares Member. Base of the wall is incompetent, laminated carbonate overlain by massive band, 4 m thick, with large selenite twins. Rooted in the massive gypsum are so-called super-cones of selenite, upward widening (see Plate 8b for detail). The super-cones are capped by a second horizon of laminated carbonate and mudstone, which in turn support a younger bed of super-cones, manifesting a new cycle of evaporation. The gypsum-dominated sequence is capped by laminated clays and limestones of the Sorbas member. Exposure is located in Río de Aguas where crossed by road from Sorbas to Carboneras, 250 m west of bridge. Geologist is Chris Talbot, photograph was taken by Jeanette Bergman (see also Dronkert, 1985). (b) Sorbas Basin. Detail of super-cone gypsum in the lower, massive gypsum bed of the Río de Aguas section. Individual super-cones are laterally separated by laminated rock of carbonate, silt and clay. The super-cone horizon itself is capped by similar laminated carbonates and mudstones. (c) Sorbas Basin. Selenite twin from los Yesares pit. Height of crystal is 20 cm. (d) Sorbas Basin. Amorphous chicken-wire gypsum from Canada de Lucaiñena, near Turrillas.

still seems to be active, because it has been reported historically that the temperature of the Baños spring (with a flux of 650 1/min) was raised from 42° to 53°C immediately after an earthquake in 1865 (Tapia Garrido, 1980). The fault also played a major role in the Pliocene sedimentation of marine slump deposits (Postma, 1984a,b).

The contact between the non-folded Messinian basal sequence of the Sorbas Basin and the Tortonian marl-turbidite sequence of the Tabernas Basin, is an angular unconformity with angles of up to 60° between the respective beddings (Weijermars et al., 1985). The Messinian succession, best studied in the Sorbas Basin (Fig. 22a–c), has been divided into the Turre and Caños Formations (Völk, 1967; Dronkert, 1976). The Turre Formation, a circa 500 m thick transgressive sequence in the Sorbas Basin (Dronkert, 1976, 1985), is only about 100 m thick in the Agua Amarga Basin south of the Sierra Cabrera (Van de Poel et al., 1984). It comprises conglomerates and calcarenites of the Azagador Member

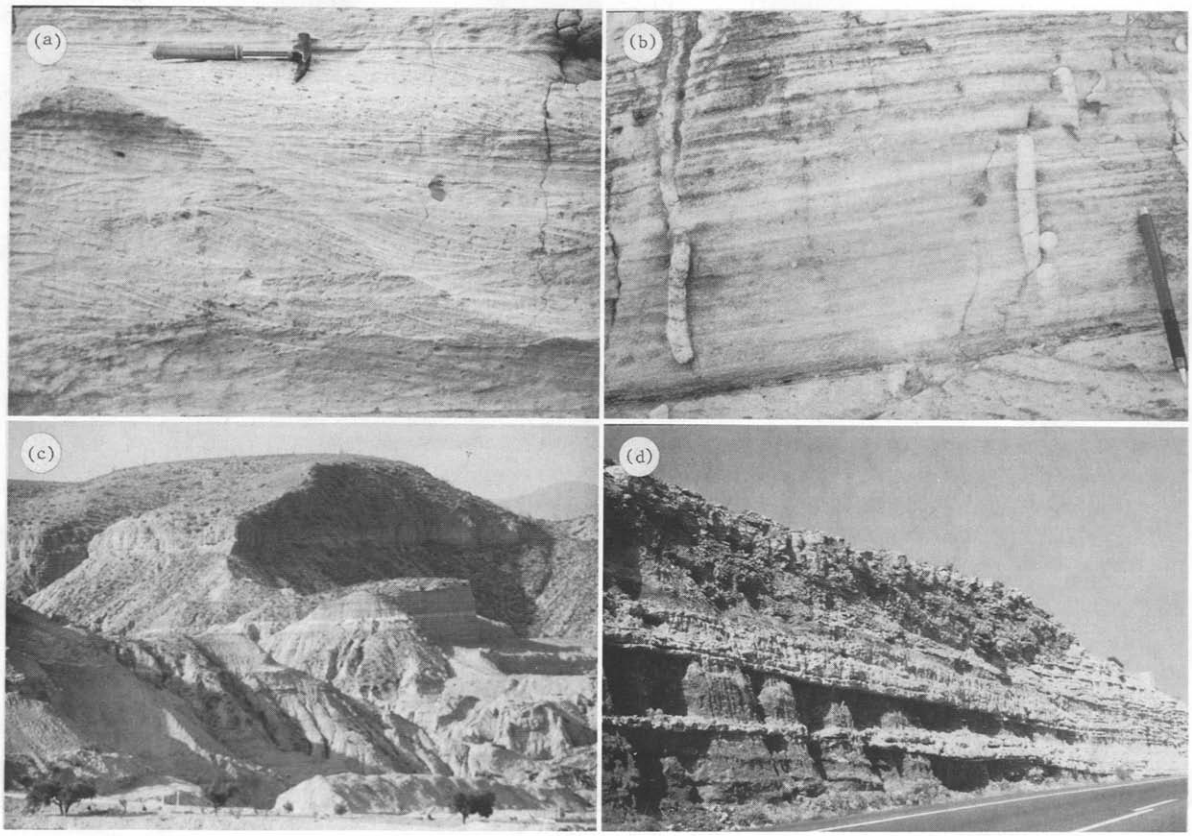


Plate 9. Late Messinian Sorbas Member and Pliocene Zorreras Member; Caños Formation. (a) Sorbas Basin. Herringbone lamination in white, coastal sandstone of the Sorbas Member. Río de Aguas, west of Sorbas (see also Roep et al., 1979). (b) Burrow channels in laminated white sandstone of the Sorbas Member. Río de Aguas, west of Sorbas. (c) Zorreras mountains, Sorbas Basin. The reddish coloured mud- and sandstones (dark albedo) of the Zorreras Member comprise three marine intercalations (light albedo). A first marine band occurs near the base of the section, which is the base of the Zorreras Member and comprises ostracods and molluscs of the terminal Messinian "Lago Mare"-facies. A second white marine band is visible half way up the cliff section. The thick competent bed under the cap of the mountain is a yellow sandstone, representing the last ingression of a shallow Pliocene sea in the Sorbas Basin. Exposure 2 km outside Sorbas along road to Lubrín (interpretation adopted from unpublished excursion guide by Tom Roep, now at the Free University, Amsterdam). (d) Red continental sandstones (dark albedo) interfingering with whitish, shallow marine conglomerate, Pliocene. Looking NE. Along road Níjar–Almería, Almería Basin.

(20–30 m), yellow diatomitic marls of the Abad member (30 m) and barrier reefs of the Cantera Member (10–30 m), containing coral limestone of in situ porites species (Van de Poel et al., 1984). Exposures of extremely fine-grained diatomitic marls of the Abad Member in the Rambla del Inox (Loma de las Patas, S-flank of the Sierra Alhamilla) have been exploited for cement production.

The largest extent of the Messinian trans-

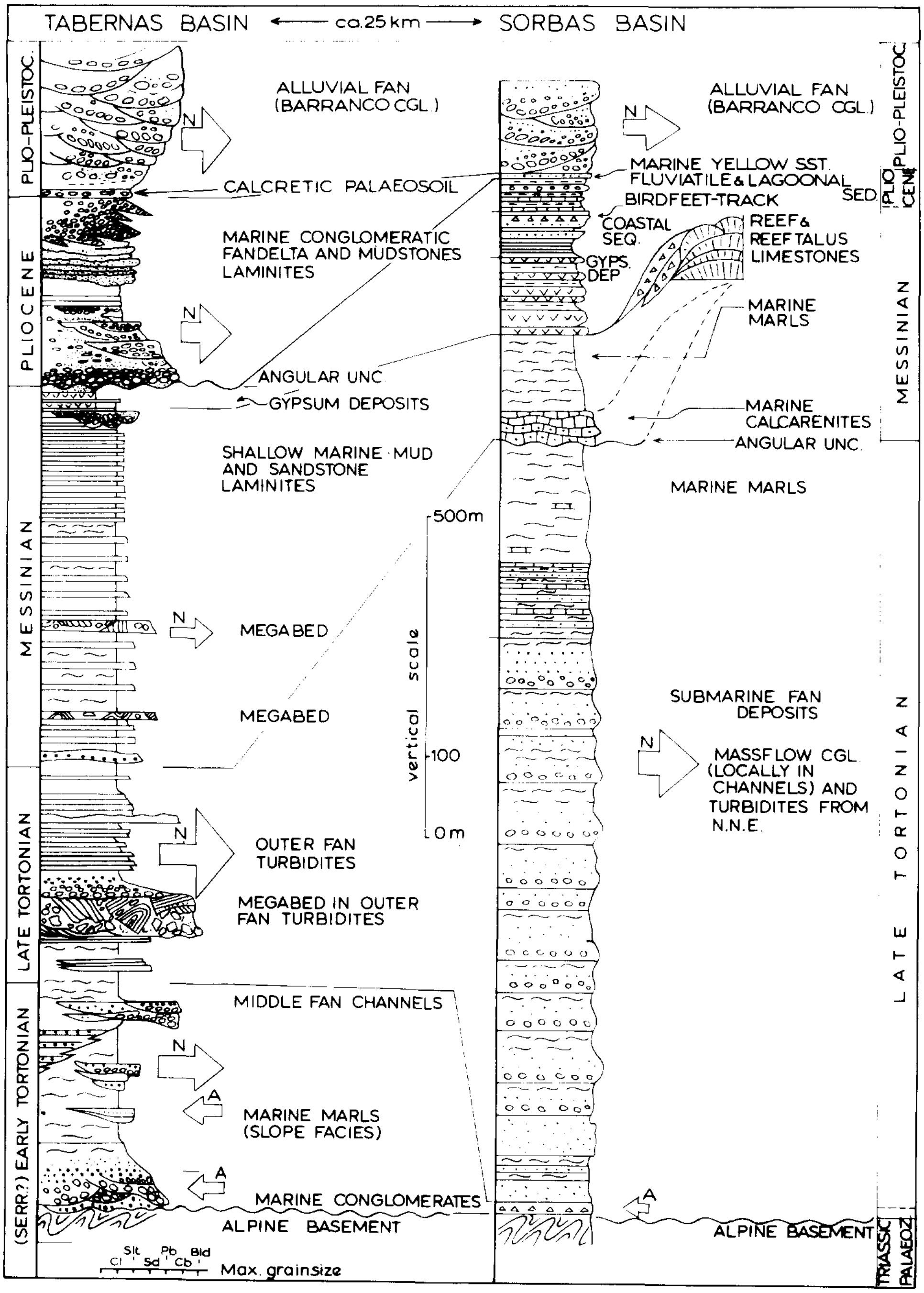


Fig. 19. Composite stratigraphic columns of (a) Almería and Tabernas Basins and (b) Sorbas Basin. Major influxes of early Alpujarride (A) without Nevado–Filabride detritus are shown by left-facing arrows. Right-facing arrows denote introduction of predominantly Nevado–Filabride detritus (N). The megabeds in the Tabernas Basin may be due to seismic enhancement of slope instability (from Weijermars et al., 1985).

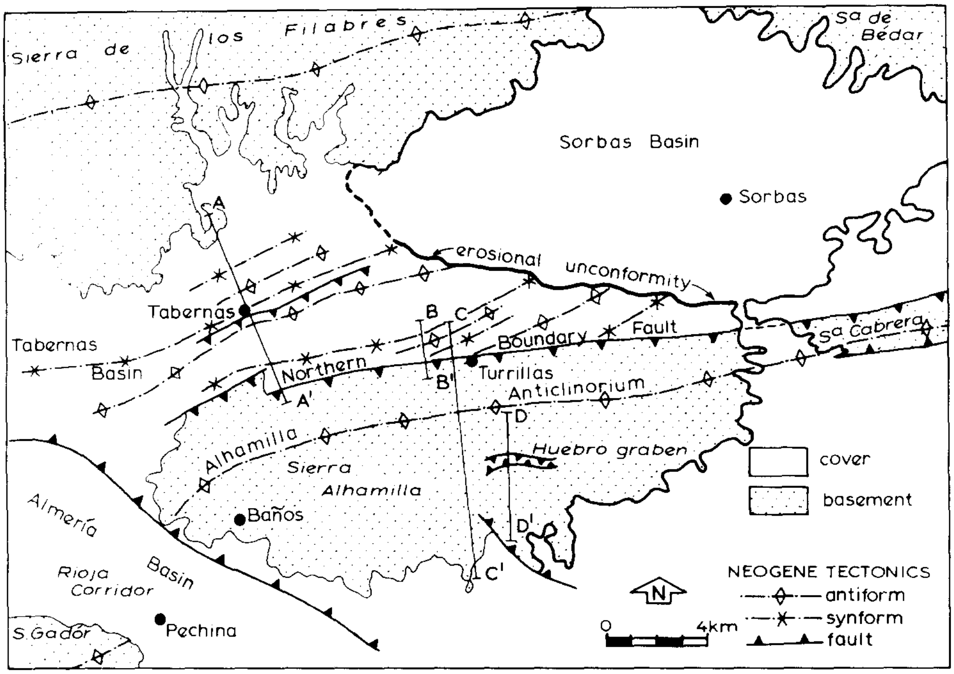


Fig. 20. Structural sketch map of the area covered by the western part of the detailed geological map of Fig. 18. Late Tortonian folds in the Tabernas Basin are truncated in the east by an erosional unconformity at the base of the Messinian sequence of the Sorbas Basin. The fold structures are truncated in the south by the Northern Boundary Fault of the Sierra Alhamilla and to the west by the Baños Fault. Sections *A–A′* and *B–B′* are outlined and shown in Fig. 21 (from Weijermars et al., 1985, fig. 5).

gression is marked by the barrier reefs of the Cantera Member (Fig. 18) (Völk, 1967; Dronkert, 1976; Dabrio and Martin, 1978; Van de Poel et al., 1984; Dabrio et al., 1985). The reefs of the Cantera Member have been described as the Plomo Reef Formation by

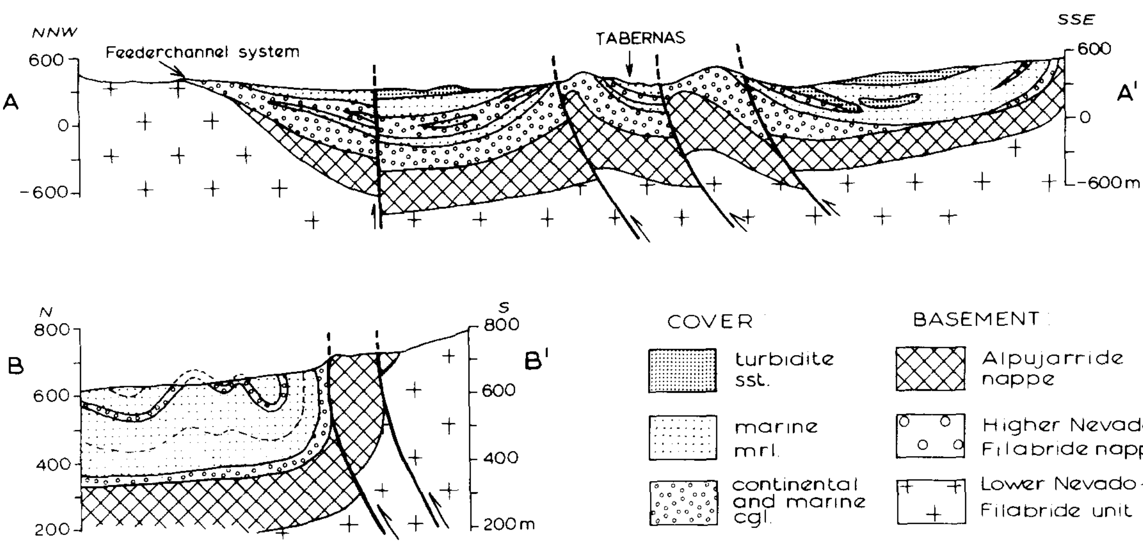


Fig. 21. Sections illustrating shortening of the Tabernas Basin. A–A′: Serravallian–Tortonian sedimentary rocks are folded and reversely faulted; B–B′: Imbricated Alpujarride rocks separate the Nevado–Filabride basement of the Sierra Alhamilla from the folded Neogene sedimentary rocks of the Tabernas Basin (from Weijermars et al., 1985, fig. 6a and b).

Armstrong et al. (1980). The latter authors suggested that the high porosity of these reefs provides excellent oil reservoir rock, especially off-shore, where the Messinian sequence has been buried to mesogenic depths. However, it is unlikely that the reefs – a marginal fringe of a transgression still exposed on land in southeastern Spain (Fig. 18) – were formed in parts of the Alboran Sea now subsided even further than 5 Ma B.P.

The transgressive Turre Formation is overlain by the regressive Caños Formation (Fig. 22a). The Caños Formation comprises gypsum (130 m), limestone (50 m) and mudstone (70 m) in the Sorbas Basin (i.e. the Yesares, Sorbas and Zorreras Members, respectively; Dronkert, 1976). These members can also be mapped along the southern flank of the Sierra Cabrera (i.e. near the Molata hill-side). However, the top of the Messinian in the Agua Amarga Basin, covering the Gata volcanic rocks south of Carboneras (Fig. 17), is developed in a facies slightly different from that of the classical members of the Caños Formation. This led to the differentiation of the following local members in the Caños Formation: a Manco Member comprising an alternation of marls and limestone (0–3 m), Agua Amarga Breccia Member composed of limestone breccia (0–40 m), and the Feos (= Zorreras) Member of laminated mudstone (0–3 m) (Van de Poel et al., 1984).

The Caños Formation comprises one of the best preserved continental remnants of the Messinian Salinity Crisis (MSC), a catastrophic event which led to the desiccation of the Mediterranean 7–5 Ma B.P. The Yesares Member in the Sorbas Basin, described in detail by Dronkert (1976, 1977, 1985), outcrops over an area of about 25 km^2 and reaches a maximum thickness of 130 m. The sequence comprises mainly gypsum but is intercalated with carbonatic mudstones (Fig. 22b). I estimate the total gypsum reserve of the Sorbas Basin at 1 km^3 and a similar volume surfaces along the south flank of the Sierra Cabrera and Carboneras (or Agua Amarga) Basin. This suggests a total gypsum reserve of some 5 billion tons. The gypsum had been quarried only on a small scale until the early 1970's, when large-scale quarrying commenced. The Messinian gypsum is presently exploited in eight, large open pits with a combined annual production of 10 million tons. The reserve of 5 billion tons estimated here implies that mining seems secure until at least the end of the next century. The calcium sulphate of the Yesares Member can only be mined because it has not been transformed into anhydrite. Gypsum ($CaSO_4 \cdot 2H_2O$) becomes unstable at temperatures above 45°–60°C which are reached after burial to depths of about 1.5 km where it would dehydrate into insoluble anhydrite ($CaSO_4$) (Marsal, 1952). The gypsum is now shipped from Garrucha, Carboneras and Almería. The Cantera reefs formerly exposed at the Mesa Roldán have been almost entirely removed and were used to construct the piers of the new gypsum harbour south of Carboneras in the early 1980's.

The base of the Caños Formation (i.e. Yesares, Sorbas and Manco Members) is interpreted as representing the first stage of the Messinian Salinity Crisis (Dronkert, 1976, 1977, 1985; Pagnier, 1976; Troelstra et al., 1980; see also section 8.4), regionally characterised by cyclic evaporite precipitation in a hypersaline, but still deep Mediterranean Sea 6.5–5.5 Ma B.P. (Weijermars, 1988b). Although Messinian halite has been drilled in the Lorca Basin, halite and other salts are almost completely absent in the Yesares sequence which can best be explained as follows. The precipitation of salts from evaporating sea water occurs in the order of reciprocal solubility; i.e. carbonate, gypsum, halite and bitter salts (Usiglio, 1849). Gypsum will not start to precipitate until either 1/5 of the original water depth is reached or the brine concentration is 5 times denser than that of standard Mediterranean sea water (Usiglio, 1849). Similarly, halite needs either 1/10 of the original water depth or 10 times higher concentrations and the remaining salts do not precipitate until 1/60 of the original

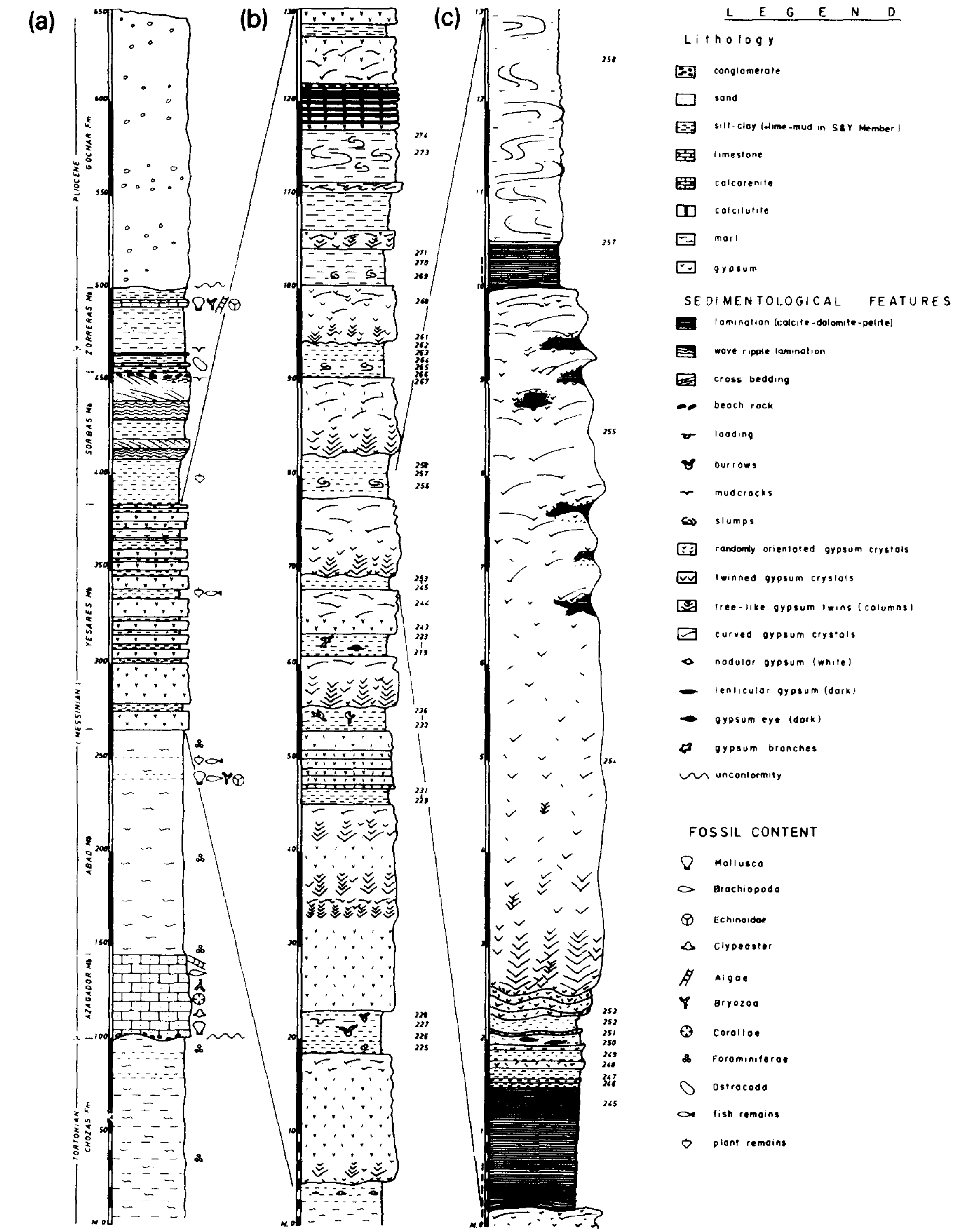


Fig. 22. (a) Composite section of the Messinian succession in the centre of the Sorbas Basin (Los Molinos del Río de Aguas section). The Messinian comprises the Turre and Caños Formations. The Turre Formation comprises the Azagador, Abad and, locally, Cantera Members. The Caños Formation corresponds to the Messinian Salinity Crisis and comprises the Yesares, Sorbas and Zorreras Members. (b) Enlargement of the cyclic-evaporite deposition of the Yesares Member, which represents the first stage of the Messinian Salinity Crisis. (c) Detail of a single gypsum-pelite couplet in the cyclic-evaporite sequence of the Yesares Member (from Dronkert, 1985, fig. 1.40).

water depth or 60 times higher concentrations are reached. The absence of halite in the Yesares Member of SE Spain can therefore be explained by a limited influx of Atlantic water over a narrow sill in the Arc of Gibraltar. This led episodically to brine concentrations over 5 times that of modal sea water, but never more than 10 times, otherwise vast amounts of halite would have been deposited. The intercalations of carbonate mudstones in the Yesares sequence (Fig. 22c) would then represent intervals of renewed influx of Atlantic water. This is in agreement with the progressive increase of the carbonate content upwards within these intercalations, suggesting progressive enrichment of the brine concentration. Also, open marine fauna has been described from marls sandwiched between gypsum (Montenat et al., 1980). Similar inferences have been made from the San Miguel da Salinas Basin near Alicante (Garcin, 1987).

Sedimentation rates estimated from the German Zechstein evaporites are: 0.06 mm a^{-1} for carbonate, 0.8 mm a^{-1} for anhydrite (gypsum dehydrated after burial) and 80 mm a^{-1} for halite (Richter Bernburg, 1960). Evaporation of 1 m of sea water would deposit a crust comprising 0.04 mm carbonate, 0.74 mm gypsum, 13.7 mm halite and 3.54 mm bitter salts (Usiglio, 1849). Together with the thicknesses of the Yesares sequence (Fig. 22b) this implies that the evaporites of SE Spain could easily have been deposited within as little as 100000 years (cf. Dronkert, 1977).

The complete desiccation of the Mediterranean Basin during the second stage of the MSC 5.5–5 Ma B.P. (deep-basin/shallow-water facies, Weijermars, 1988b) is represented by the top of the Caños Formation (i.e. Zorreras, Feos and possibly Agua Amarga Breccia Members). This is obvious, not only from dessication cracks in the Zorreras Member of the Sorbas Basin (Roep and Beets, 1976; Roep et al., 1979), but also from *Chara* sp. beneath the base of the Pliocene Cuevas Formation in the Vera Basin (Geerlings et al., 1980), a fluviatile origin and caliche soils of the Feos Member and karstic holes in the top of the Agua Amarga Breccia Member (Van de Poel et al., 1984). Additionally, ostracod fauna suggests that brackish ("Lago Mare") to fresh water conditions began to prevail at the termination of the Messinian deposition in the Sorbas and Agua Amarga (Carboneras) Basins (Roep and Van Harten, 1979).

Knowledge of the Messinian Salinity Crisis might be improved by detailed dating and the comparative study of sedimentary rocks showing imprint of continental facies related to the termination of the Messinian Salinity Crisis. Such studies could include the *Chara*-bearing sedimentary rocks just below the base of the Pliocene Cuevas Formation in the Vera Basin (Geerlings et al., 1980), caliche soils in the Feos Member of the Agua Amarga Basin (Van de Poel et al., 1984) and "Lago Mare"-facies near the base of the Zorreras Member of the Sorbas Basin (Roep and Van Harten, 1979).

The limestone breccia of the Agua Amarga Basin is a giant debris flow, up to 40 m thick. The cause of this gravity flow has been attributed to dissolution of underlying or intercalated (but hypothetical) salt layers (Van de Poel et al., 1984). However, the emplacement of this olistostrome by gravity collapse may also be explained by enhanced slope instability due to (1) a sudden retreat of the Mediterranean near the end of the MSC, and (2) seismic motion of the nearby Almería Fault Zone (cf. sections 7 and 8). Continental erosion associated with the end of the MSC is also obvious from the erosional unconformity between the Messinian and overlying Pliocene deposits. Pre-Pliocene erosion channels caused a palaeo-relief of up to 200 m, before the onset of the marine Pliocene transgression (Van de Poel et al., 1984).

The subsequent Pliocene transgression can be interpreted as the abrupt termination of the Mediterranean Salinity Crisis due to the entrance of saline waters into the Mediterranean, when local graben tectonics reopened the Straits of Gibraltar 5 Ma B.P. (Weijermars, 1988b). The deposition of gravel fan deltas (Gilbert-type) by slumping and mass

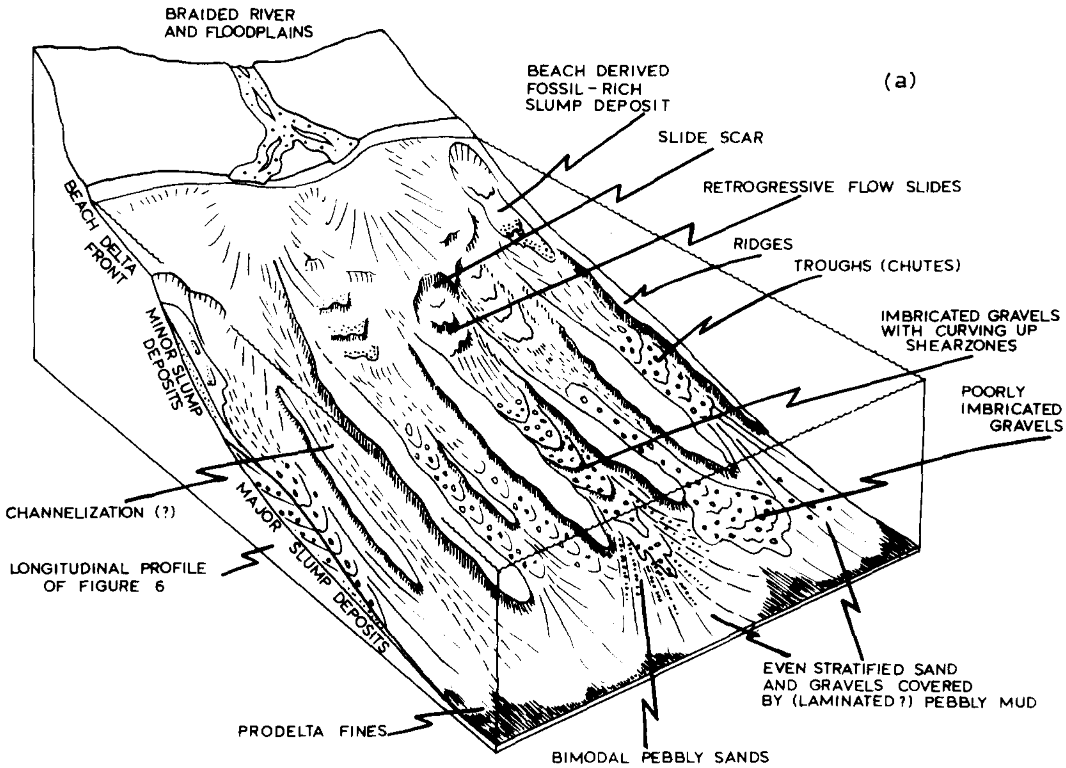


Fig. 23a. Sedimentation model for slumping and gravity collapse of meta-stable deposits near the Baños Fault scarp. Steep depositional slopes maintained the growth of the Abrioja conglomeratic fan complex (from Postma, 1983, fig. 2).

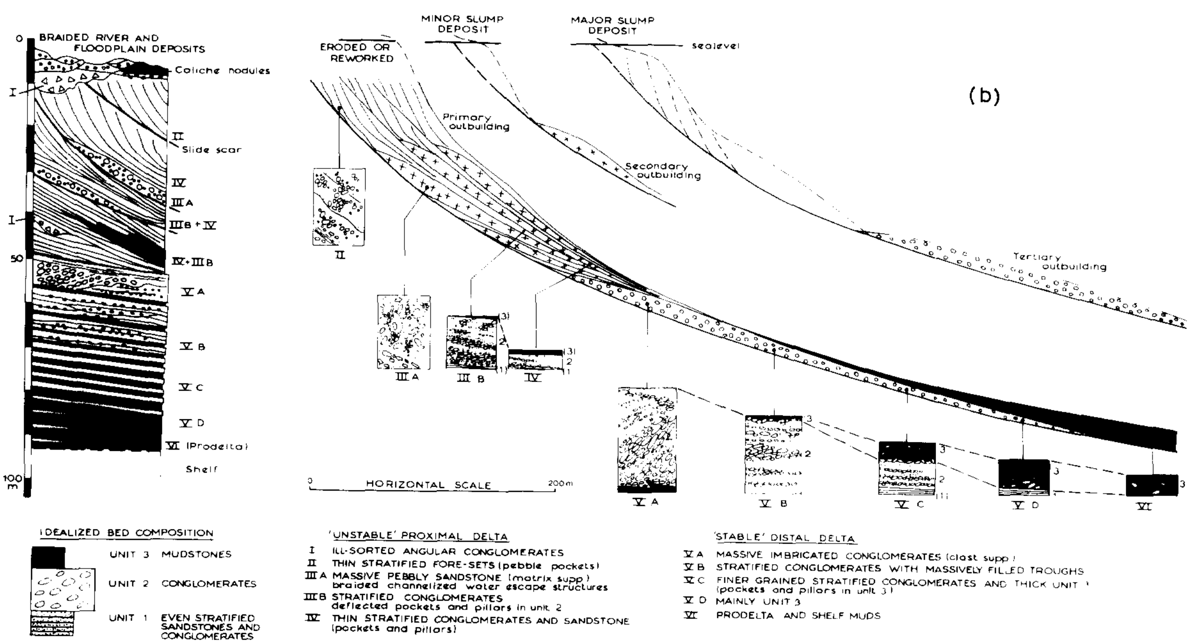


Fig. 23b. Schematic model of the Abrioja fan-delta complex. Proximal sedimentary rocks are unstable and may involve collapse of the beach, near-shore and upper delta-slope deposits. Distal sedimentary rocks have lower dipslopes and therefore are more stable than the proximal fan deposits (from Postma, 1984, fig. 9).

flow of unstable slope due to faulting is typical for the marine Pliocene sedimentation (Fig. 23a and b; Plate 12c and d). For example, such marine gravel deposits with giant foresets of up to 40 m high have been mapped in the following locations: (1) near the Baños Fault in the Río ja Corridor (Abrioja Formation; Postma, 1983, 1984a,b); (2) near the Almería Fault Zone in the Campo de Níjar (Morales Formation; Addicot et al., 1978); (3) near the Palomares Fault in the Vera and Pulpi Basins (Espiritu Santo Formation; Völk, 1966a,b; Veeken, 1983; Postma and Roep, 1985); and possibly (4) near the Alhama de Murcia Fault in the Lorca Basin (Ruzafas Formation; Geel, 1977). These important strike-slip faults are discussed in more detail in section 7.

Not all marine Pliocene deposition in the area mapped was controlled by the fast Neogene fault movements. For example, in the Agua Amarga Basin, the marine-Pliocene transgression is represented by a few meters of basal conglomerate overlain by up to 50 m thick yellow-white bioclastic calcarenites containing remnants of bryozoa, echinoid spines and plate fragments, molluscan shells, algae and abundant benthic (but no planktonic) foraminifers (Molata Formation; Van de Poel et al., 1984).

Finally, the entire area of modern SE Spain emerged above sea level in the Late Pliocene. This can be inferred from the regressive and continental Pliocene facies represented by alluvial fan conglomerates. These have been incorporated in the Campo de Níjar Formation of the Agua Amarga Basin (Ochoa et al., 1973; Van de Poel et al., 1984), Gádor Formation of the Almería Basin (Postma, 1984b), Gochar Formation of the Sorbas Basin (Roep et al., 1979) and Salmeron Formation of the Vera and Pulpi Basins (Völk, 1967; Veeken, 1983). These alluvial fan conglomerates may locally reach thicknesses up to 100 m and are difficult to distinguish from the Quaternary barranco deposits. Marine terrasses of Pleistocene and Holocene age locally occur near the modern coastal outlines of SE Spain (Baena et al., 1977; Goy et al., 1986; Hillaire-Marcel et al., 1986).

6. NEOGENE VOLCANIC ROCKS

The volcanic rocks of southeast Spain had already attracted the attention of earth scientists two hundred years ago. This was because of the discovery of a new dark-blue mineral possessing the peculiar property of displaying different colours in different directions of view. Due to this property it was initially termed "dichroit" by Cordier (1809) who had personally collected the samples from the Cerro el Hoyazo, a small volcanic centre exposed 4 km to the SW of Níjar. It was later termed "cordierite" by Lucas (1813: p. 219), because Cordier had been the first to collect his own samples from the type locality. Other mineralogists had described the same mineral before, but their cordierite samples had been purchased from a mineral supplier leaving doubt as to their origin (Von Schlotheim, 1801; Werner, see Karsten, 1808; Hintze, 1897: pp. 922–923).

Detailed study of the xenolithic Hoyazo rhyodacites led to the conclusion that two generations of cordierite coexist. Euhedral cordierite is of magmatic origin, but the anhedral cordierites, occurring in both the metamorphic inclusions and volcanic rocks itself, are restites (Zeck, 1968, 1970, 1972; Munksgaard, 1984, 1985). Garnets, also occurring as restite inclusions, have long been exploited for the production of abrasives (Wegner, 1933), but this activity ceased in 1966 (Zeck, 1968).

Regional research has shown that the morphology of the eastern coastline of Almería province is entirely controlled by an extensive Neogene ridge of calc-alkaline volcanic rocks with some intercalated and onlapping sedimentary rocks (Fig. 24). This ridge, the Cabo de Gata Volcanic Chain, is 10 km wide and 40 km long and stretches from Cabo de Gata to Carboneras. Similar acid volcanic rocks (and other, younger lamproites, see later) are also exposed within a 1 km wide ridge along the trace of the Almería Fault Zone (i.e. La Sierra Serrata) (Fig. 24). These have been dated radiometrically and yield K–Ar whole-rock ages between 7 and 15 Ma (Bordet, 1985). Early contributions to the regional study of all those volcanic rocks were made

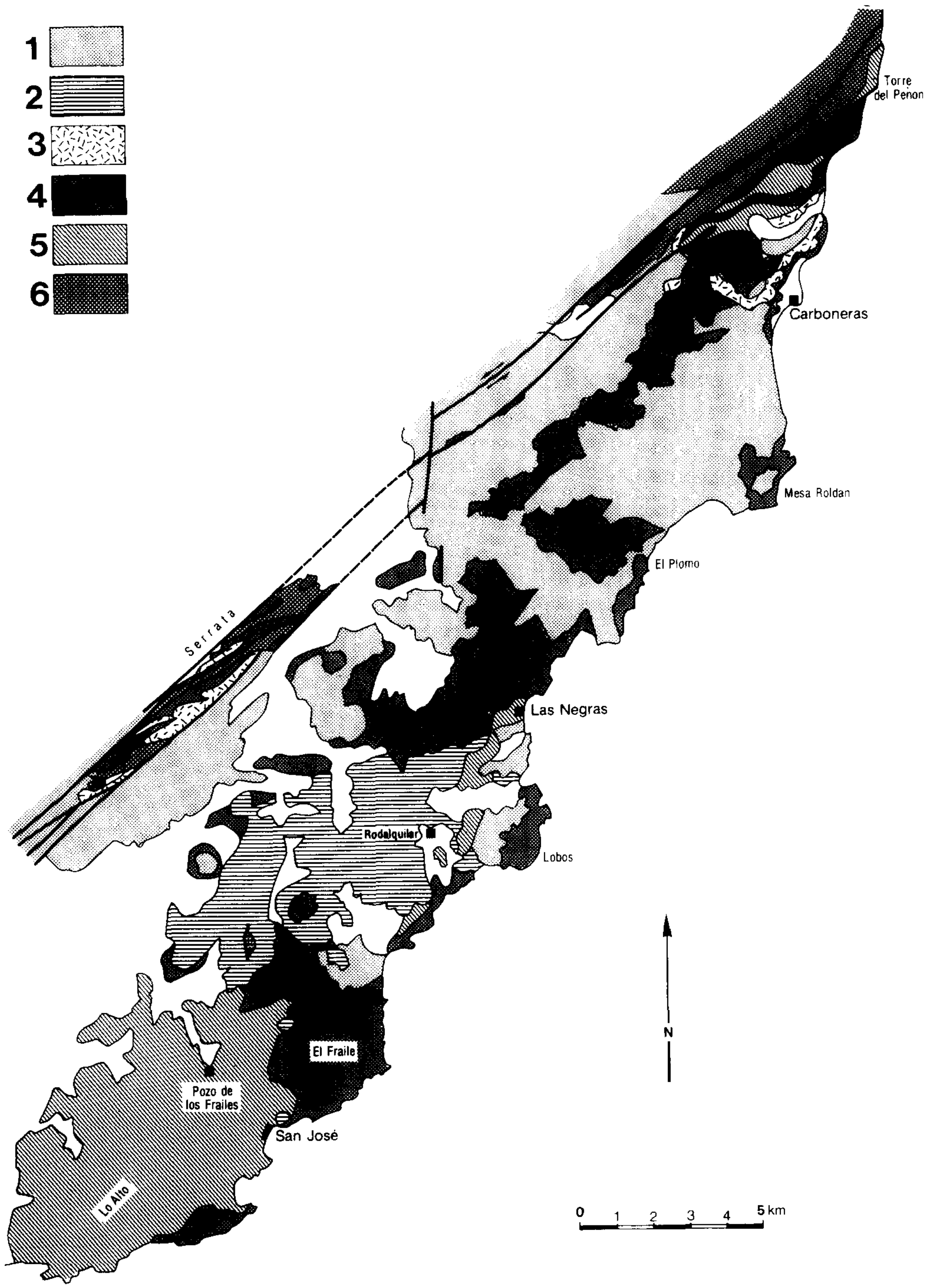


Fig. 24. Geological map of the Cabo de Gata Volcanic Chain. The NE–SW trending Almería Fault Zone is also indicated. Rock types distinguished are: (1) Neogene sedimentary rocks, (2) volcanic complex C of Rodalquilar, (3) Brèche rouge olistostrome deposit, (4) Tortonian volcanism B, including El Fraile, (5) volcanic complex A of Cabo de Gata s.s. (Serravallian) and (6) metamorphic basement of the Betic Zone (after Bordet, 1985, fig 2).

by Donayre (1877), Calderón (1882), Osann (1888, 1889, 1891a,b), Hetzel (1923), Wegner (1933), and Burri and Parga-Pondal (1936). The region comprises bentonite deposits (sodium montmorillonite), which were mined extensively in the 1950's and 1960's (Martin-Vivaldi et al., 1956; Martin-Vivaldi, 1963; Leone et al., 1983). Some characteristic features of the Serravallian–Tortonian volcanic rocks of the Sierra de Gata are illustrated in Plates 10 and 11.

The entire volcanic suite of Cabo de Gata was mapped in detail by students of professor Fúster of the University of Madrid in the 1960's (Paez and Sanchez, 1965; Coello and Castañon, 1965; Fúster et al., 1965, 1967a; León, 1967; Sánchez Cela, 1968). This work has recently been compiled and extended by professor Bordet (1985) of IGAL, Paris, who is the first to map clearly the volcanic centres of eruptions; up to 30 volcanoes are distinguished (map scale 1 : 50 000). About 15 of them have been dated radiometrically by K–Ar whole-rock methods and yielded ages ranging from 15 to 7 Ma B.P. (Bellon et al., 1983; Bordet, 1985; Di Battistini et al., 1987). The 15–7 Ma radiometric ages are consistent with biostratigraphic ages inferred from inter-

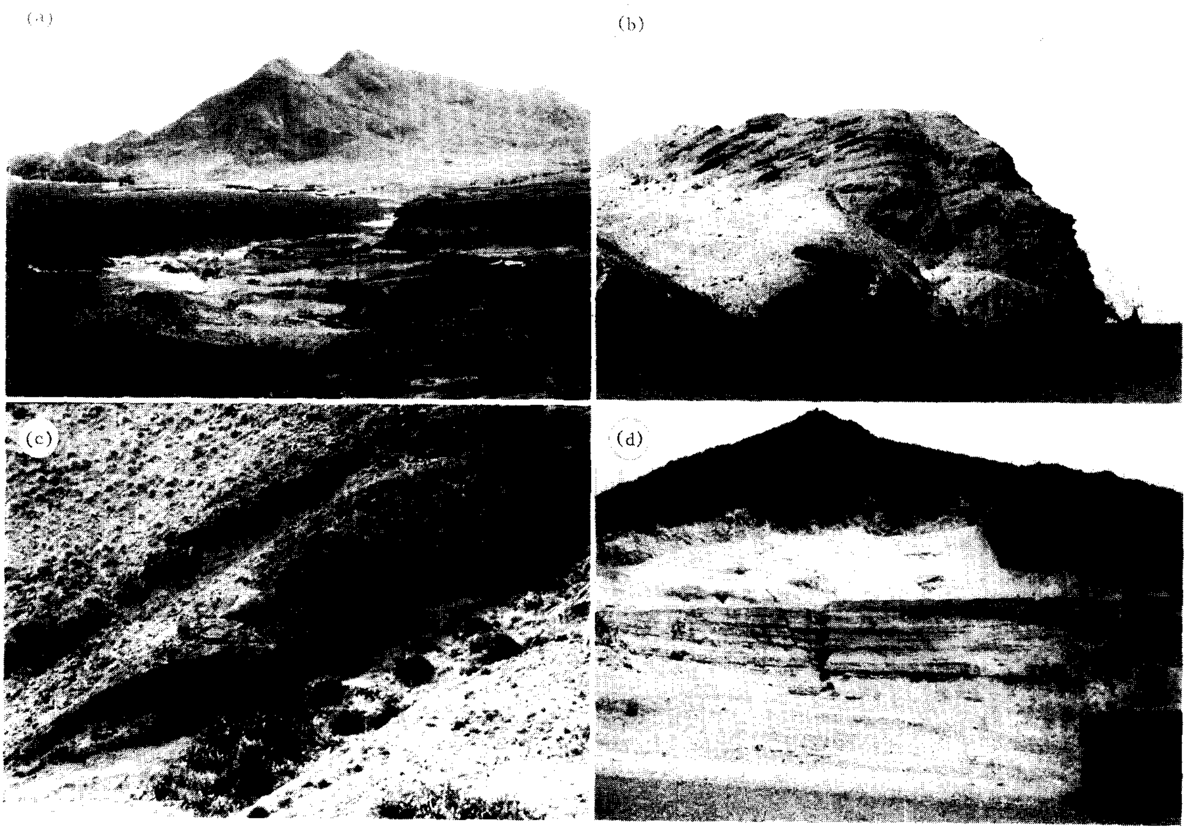


Plate 10. Serravallian–Tortonian volcanic rocks. (a) El Fraile Caldera complex with two eruption centers at the summit, El Fraile Grande (el. 493 m) and El Fraile Chico (el. 444 m). The caldera measures a diameter of about 5 km and good exposures are found along the contouring coastal road. The low area in the foreground is the bay of Los Escullos, which also was a bay in Tortonian times as may be inferred from onlapping Tortonian sedimentary rocks. Looking SSW from road Rodalquilar–San José north of Amatista, Sierra de Gata. (b) Cerro Negro (el. 171 m) with distinct layering in pyroclastics. Looking north from Las Negras beach. (c) Mega-foresets in pyroclastic surge deposit (lahars) SW of Galinaza volcanic center. Exposure below old road Las Negras–Fernan Perez, Sierra de Gata. Scale is indicated by the man in the cave in the upper right of the picture. (d) Light-coloured stratified bentonite originating from ignimbrite ash covered by dark lava flow. Bentonite was extensively mined in the area in the 1960's. Morron Mateo, Los Escullos.

calations of Serravallian–Tortonian sedimentary rocks (Bellon et al., 1983; Bordet, 1985). The youngest caldera determined hitherto (i.e. Garbanzal: 6.9 ± 0.2 Ma; Di Battistini et al., 1987) is covered by Messinian sedimentary rocks so that all volcanism of the Gabo de Gata type is pre-Messinian, i.e. > 7 Ma B.P. The SW part of the Cabo de Gata Volcanic Chain remained an island during the Early Messinian transgression as the highest level of the Messinian sea is marked by a fringe of barrier reefs of the Cantera Member (e.g. Dabrio and Martin, 1978; Esteban and Giner, 1980; Dabrio et al., 1981).

The oldest volcanic rocks, rhyolite-andesite ignimbrites in the SW part of the volcanic chain (i.e. Lo Alto or Cabo de Gata s.s.), yield Serravallian potassium–argon dates between 15 and 12 Ma B.P. (i.e. La Revancha and Alemanes Nuevos; Bellon et al., 1983;Di Battistini et al., 1987). East of Alemanes Nuevos occurs the 493 m high caldera of El Fraile which may have a Serravallian foundation according to a recent dating of 14.4 ± 0.8 Ma B.P. (Arribas et al., 1989). The base of El Fraile in Bordet's (1985: pp. 17–18) Caliguera section is formed by a sedimentary breccia, probably Early Tortonian. The bulk of the

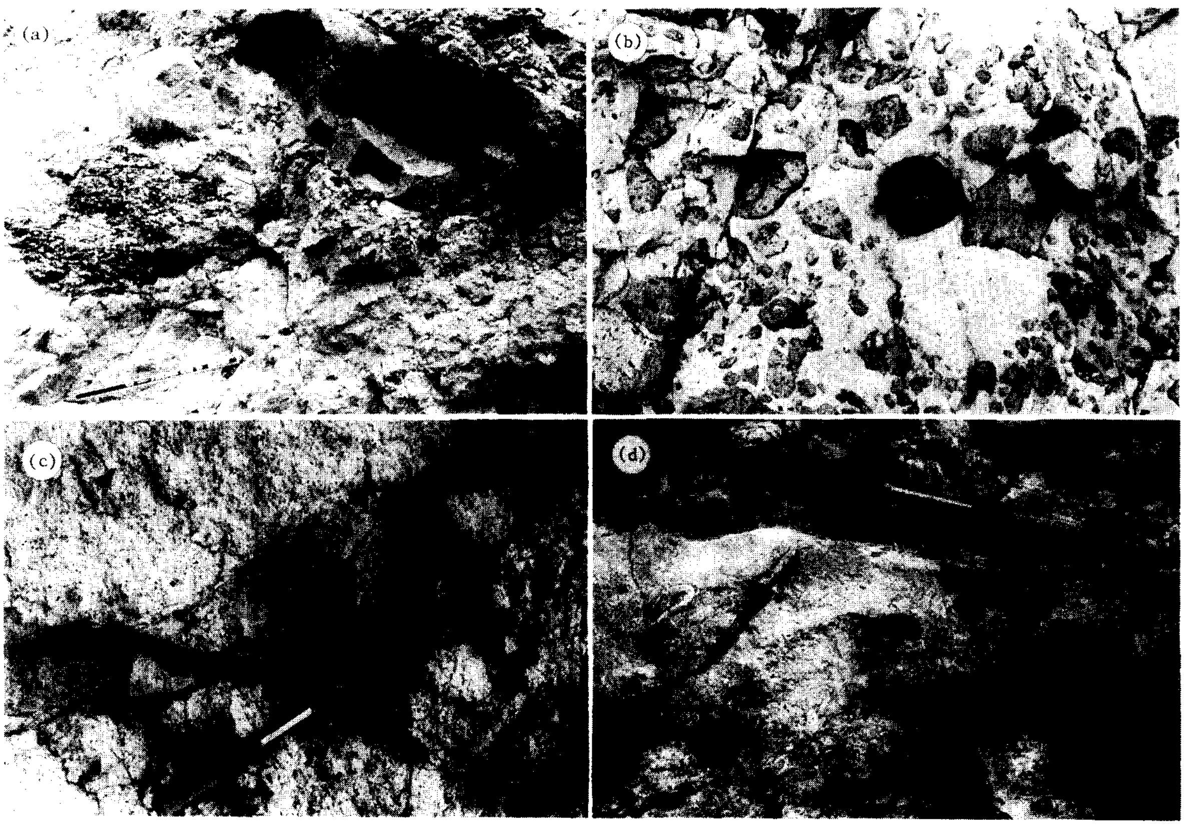


Plate 11. Serravallian–Tortonian volcanic rocks. (a) Hoyazo volcanic, Níjar. Two xenoliths crucial to the story of El Hoyazo. The inclusion to the left is a foliated granodiorite, interpreted by Zeck (1968) as a restite of crustal melting providing the rhyodacitic matrix. The inclusion in the upper right is a basaltic blob with a cooling rim, interpreted as entrained basic melt originating from greater depth and providing the heat required for subcrustal anatexis. (b) Pyroclastic of Garbanzal. Sierra de Gata. (c) Yellow pyroclastic intruded by pebbly dyke. El Fraile, Sierra de Gata. Picture is rotated clockwise over 90°. (d) Brèche Rouge, contains fragments of both (dacitic) volcanic rocks and fossils. Note the characteristic echinoderm cross-section in the upper-left part of the bed supporting the hammer. Interpreted as a terminal Tortonian olistostrome by Bordet (1985). Agua Amarga Basin, road cut near Majada, west of Carboneras.

volcanism is Tortonian (i.e. 12–7 Ma B.P.; Bellon et al., 1983; Di Battistini et al., 1987).

Tortonian sedimentary rocks in the Cabo de Gata region, represented by onlaps and intercalations of shallow water facies conglomerates, fossiliferous calcarenites and marls, nowhere exceed thicknesses of 10–20 m. Intercalations of rhyodacitic lapilli (beds of 10–20 cm) and influx of tephra are characteristic for the Tortonian of the Gata area. In contrast to the Tortonian sequence in the basins of Almería, Tabernas, Sorbas and Vera, the Sierra de Gata did not receive any significant turbidite influx. This suggests that the Serravallian volcanic centres of Alemanes Nuevos and El Fraile had already partly emerged above sealevel at the onset of or during the Tortonian, surrounded by shallow marine platforms. Tortonian calcarenite beds intercalated in the lower parts of the volcano-sedimentary sequence contain oysters and near the top occur horizons with lumachelles and calcarenites, also extremely rich in fossils (i.e. gastropods, bivalves, echinoderms, crustaceans, lammelibrachiats, *Clypeaster, Pecten, Heterostegina, Balanus*, Pycnodonts and foraminifers: *Nodosaria* sp., *Globigerina* sp., *Elphidium* sp.; cf. Bordet, 1985).

Early, nested calderas in the Rodalquilar area north of El Fraile may also have been formed in the Serravallian. The ignimbrite complexes of Rodalquilar and of Cabo de Gata s.s. are both covered by Tortonian sedimentary rocks, i.e. near San José, Cortijo del Fraile, Penoñes and Cerro de los Guardias. These ignimbrites have a similar petrochemistry, and both show a peculiar enrichment of potassium (Bordet, 1985). Pervasive hydrothermal alteration led to extensive metallogenesis in both areas. The Mn- and Pb-Zn(-Cu-Au-Ag) agate veins of Alemanes Nuevos were mined in the 19th century (Pineda Velasco, 1984; Bordet, 1985) and the Au-Te-chalcedony-alunite veins in the Cinto complex of Rodalquilar until 1966 (Lodder, 1966).

The hydrothermal alteration leading to metallogenesis in Rodalquilar's Cinto Complex, previously supposed syngenetic with the (Serravallian) emplacement of the ignimbrite complex (De Roever and Lodder, 1967), is now interpreted as epigenetic (Arribas et al., 1989). Potassium–argon dating of alunite and illite from alteration zones associated with gold deposits suggests that mineralisation occurred in the Early Tortonian 10 Ma B.P. (Arribas et al., 1989), before emplacement of unaltered pyroxene andesite flows dated at 8.4 to 7.5 Ma B.P. (Bellon et al., 1983; Di Battistini et al., 1987). This view is supported by the discovery of a gold grain in a conglomerate horizon at the base of the overlying Tortonian cover (near the Cerro de los Guardias; Lodder, 1966: p. 40). This gold grain is probably a remnant of a Late Tortonian placer deposit derived from the Cinto complex.

The gold in the hydrothermal veins of Rodalquilar is macroscopically visible only in exceptional cases. The gold is usually found in a gangue of quartz (agate) or ironoxide compounds, with particle dimensions of 3–25 micron and occasionally up to 0.1 mm (Lodder, 1966). The ambient temperature during the main period of metallogenesis was about 400 °C, as can be inferred from intergrowths of sphalerite and chalcopyrite and the occurrence of dickite and pyrophyllite in the marginal zones of alunite-quartz veins (Lodder, 1966: p. 85). The average gold concentration in the Cinto Complex is 4.5–5 g/ton with maximum grades up to 100 g/ton (Lodder, 1966). The production in 1964 was 700 kg pure gold, corresponding to an annual tailing accumulation of about 140 000 tons or 50 000 m^3. Exploitation of the silica-alunite veins was resumed in 1989, with an anticipated production of 200 000 tons/year over 4 years and an estimated gold yield of 2.0 g/ton (P.A. Hernandez, pers. commun., 23-07-1989). The Rodalquilar deposits are subjected to new investigations in a programme focussing on the origin of Spanish and US epithermal gold-alunite deposits, funded by the US–Spain Joint Committee for Scientific and Technical Cooperation (Arribas et al., 1989; Cunningham et al., 1990; Rytuba et al., 1990).

It should be noted that a distinction has been made between an Early and a Late Tortonian phase of green and black volcanism, respectively (Ba and Bb volcanic rocks of Bordet, 1985; cf. Venturelli et al., 1984). The green volcanic rocks typically comprise pyroclastic deposits with 1–2 cm idiomorphic black amphibole crystals and occasional smaller crystals of oxidized pyroxenes. The black volcanic rocks comprise both ortho- and clinopyroxenes but lack any amphibole. Both types can be recognized in the field on the basis of their typical greenish and blackish colours. This correlation between geochronology and mineral chemistry has recently been confirmed by radiometric work of Di Battistini et al. (1987) and detailed mapping of the El Fraile volcano by Fernández Soler (1987).

The abundance of block and ash flow deposits (lahars) in the Tortonian volcanic rocks of the Gata region is indicative of the explosive nature of the eruption mechanism, due to a high gas content (cf. Cas and Wright, 1987). Such an eruption may itself be accompanied by seismic activity due to displacements in the magma conduit (Ryan et al., 1981). Consequently, this particular type of seismic reaction need not be related to any strike–slip activity in the region. Deposition of Bordet's (1985) brèche rouge (Fig. 24) may also be due to explosive volcanism. This breccia contains fragments of both pyroclasts (amphibole and pyroxene dacite) and bioclasts (a.o. echinoderms) within a rose matrix of calc-arenite. The up to 100 m thick brèche rouge covers locally Tortonian marls with *Globorotalia humerosa*, and is itself covered by Messinian calcarenites rich in *Pentacrines* (Bordet, 1985). The brèche rouge has been interpreted as a Late Tortonian olistostrome by Bordet (1985).

The Volcanic Chain of Cabo de Gata is suggested to have been continuous with a similar volcanic arc extending from Vera to

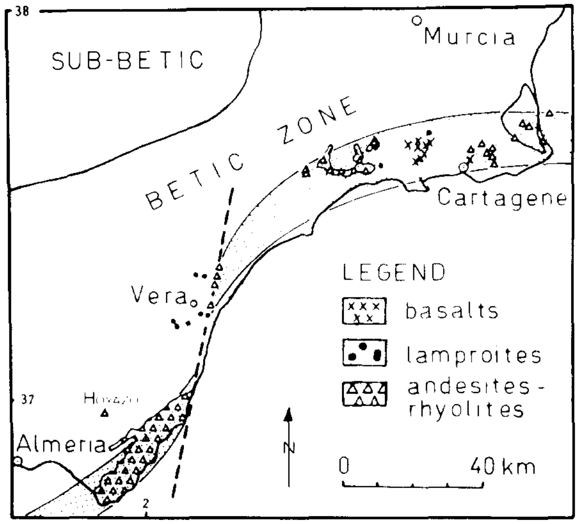


Fig. 25. The 15–8 Ma Neogene band (shaded) of calc-alkaline volcanic rocks (andesites-rhyolites) of SE Spain are disrupted by the Palomares Fault, suggesting a sinistral strike–slip displacement of about 30 km. Also indicated are a suite of lamproites (5 Ma B.P., see text) in the Vera Basin and basalts of Cartagena intruded about 2.5 Ma B.P. (Bellon et al., 1983) (from Weijermars, 1987a, fig. 5).

Cartagena (Fig. 25), before disruption by the Palomares Fault (Weijermars, 1987a,c). Calc-alkaline rocks exposed along the trace of the Palomares Fault in the Vera Basin yielded K–Ar ages of 7.0 ± 2 Ma (Nobel et al., 1981). This is within the range of 14–7 Ma ages determined for the potassium-rich calc-alkaline rocks (and shoshonites) exposed between Aguilas and Cartagena (Bellon et al., 1983). However, the enrichment in potassium of the latter volcanic rocks, has been used to argue that the volcanic rocks east of the Palomares Fault form a petrochemical suite entirely different from that of Cabo de Gata, but similar to that of El Hoyazo (De Larouziére et al., 1988; Montenat et al., 1988). Consequently, the volcanic rocks of the arc Hinojar, Mazarron, Cartagena, Mar Menor would not need to have been continuous with the Cabo de Gata Volcanic rocks before displacement by the Palomares Fault, according to the latter authors.

A second, petrochemically distinct, suite of volcanic rocks is exposed in the Vera Basin.

These can be classified as lamproites, characterised by high potassium and magnesium content and relative depletion in Al, Fe and Ca. The particular variety of lamproite encountered in the Vera Basin has been termed verite (Fúster and De Pedro, 1953; Fúster, 1956; Fúster et al., 1967b). Similarly, lamproites near Jumilla, Fortuna and Cancarix in the Prebetic Zone have been termed jumillites, fortunites and cancalites, respectively (Fúster et al., 1967b).

Phlogopite flakes from verite samples of a subvolcanic body in the Vera Basin (Cabezo Maria, 4 km SW of Antas, Fig. 19) have yielded radiometric K–Ar ages of 8.6 ± 0.3 Ma B.P. (Nobel et al., 1981). Veritic lavas have also been dated by K–Ar whole-rock measurements which gave 10.9 ± 1 Ma (Bellon, 1976; Bellon and Brouse, 1977). However, these particular radiometric ages should be regarded with scepticism in view of their inconsistency with biostratigraphic observations. For example, the veritic lavas rest upon the Late Messinian Turre Formation and are covered by the Early Pliocene Cuevas Formation (Völk, 1967; using biozones of Montenat et al., 1976). The Messinian–Pliocene time boundary is dated at about 5 Ma B.P. (Vass and Bagdasarjan, 1978), which seems therefore a better estimate for the age of the veritic lavas.

A third group of volcanic rocks is represented by young basalts exposed near Cartagena yielding K–Ar ages of 2.8–2.6 Ma B.P. (Bellon et al., 1983). These contain peridotitic and ultramafic nodules, resulting from fractionation of upper mantle rocks and segregation of alkaline melts (Sagredo, 1972, 1973; Dupuy et al., 1986; De Larouziére et al., 1987; Capedri et al., 1988).

The cause of volcanism in southeast Spain is a matter of considerable debate. Earlier models suggest involvement of an oceanic plate subducted northward under the Alboran Basin from an ancient suture in North Africa (Araña and Vegas, 1974). This model has been modified by Ruiz and Badiola (1980), who explain the occurrence of calc-alkaline, high-K calc-alkaline and shoshonitic volcanic rocks by subduction of the Alboran sea floor itself. Admittedly, it would fit with the worldwide observation that potassium content increases with the depth of the magma source on subducting oceanic plates. This also seems valid for the geochemical polarity of Quaternary volcanic rocks in the Aegean Arc (Keller, 1982).

However, two major objections (cf. Puga, 1980) can be raised against any model explaining the volcanic suite of southeast Spain by a subduction setting. These are: (1) the Alboran Basin comprises thinned continental crust and is not thickened as would be expected in a collision zone (Weijermars, 1985a,b, 1987a,b); and (2) the occurrence of similar volcanic rocks along the African coast of the Alboran Sea contradicts the alleged polarity in geochemical trend (Hernandez et al., 1987; cf. Hernandez, 1986). The latter authors suggest that the volcanism in the Arc of Gibraltar is genetically related to strike-slip faulting in the region (see also De Larouziére et al., 1987).

An alternative interpretation still seems possible, involving fractionation of mantle rocks during emplacement of a diapir under the Alboran Seafloor and the partial melting of the base of the continental crust downwarped in the return flow peripheral to the mantle diapir (cf. Torres-Roldán et al., 1986). Such a model needs elaboration but could possibly explain the gradual change in magmatic composition from acid to basic between 22 and 2 Ma ago. Previous study of the petrogenesis of the El Hoyazo rhyodacites suggests the involvement of a basic magma causing crustal anatexis at depth (Zeck, 1968). This particular rhyodacite contains about 1 vol.% xenolithic inclusions of sillimanite gneiss, quartz-cordierite rock and hornfelses (up to 60 cm diameter). These have been interpreted as Al-rich restites of a granitoid melt created by crustal anatexis. The El Hoyazo lava is therefore considered an erupted migmatite or migmatitic lava (Zeck, 1968: p. 144). The presence of additional basic igneous inclusions with chilled borders suggests that these are entrained volumes of a basic magma originating at greater depths. This basic magma may also have supplied the heat necessary for the migmatisation and anatexis of deep continental crust (Zeck, 1968: pp. 148–149).

7. SEISMIC FAULTS

Extensive geological surveying in southeast Spain started after the occurrence of the devastating earthquake of 25th December, 1884, near Málaga, which caused about 2000 casualities and destroyed 17000 houses (cf. Udías et al., 1976). Major cities in southern Spain had previously been destroyed by earthquakes, i.e. Almería (1487, 1522, 1659, 1804), Vera (1518, 1865), Murcia (1829) (Udías et al., 1976; Bousquet, 1979). After the 1884 damage, prominent academic societies of Paris and Rome sent a team of geologists to southern Spain to study the damage and cause of the earthquakes (Missión d'Andalousie). The earthquakes remained enigmatic at that time and the main result was the organisation of mapping projects (Missión d'Andalousie, 1889).

We now know that most of these earthquakes are shallow (< 60 km) and due to fault motions in the crust (Udías et al., 1976). However, the exact location of the faults responsible for the seismic movements is difficult to trace, because the permanent seismological network in the region cannot determine the theoretical epicentres better than within a 25 km radius of the actual epicentres (Mezcua et al., 1984). Independently, geological studies have suggested the existence of major wrench faults in the eastern Betic Cordilleras (Jansen, 1936; Fernex, 1964; Rondeel, 1965; Völk, 1967; Westra, 1969; Foucault, 1971; Bousquet and Montenat, 1974; Bousquet et al., 1975; Bousquet and Philip, 1976a,b; fig. 1 in Geel, 1976; Gauyau et al., 1977; Hermes, 1978; Bousquet, 1979; Van de Fliert et al., 1980; De Smet, 1984a,b; Bordet, 1985; Ott d'Estevou and Montenat, 1985; Sanz de Galdeano, 1983, 1987; Sanz de Galdeano et al., 1985; Montenat et al., 1987; De Larouziére et al., 1987, 1988; De La Chapelle, 1988; Coppier et al., 1989).

Many of the fault structures suggested are at variance with geological field observations, but several major fault traces emerge from mapping abrupt discontinuities in otherwise smooth geological boundaries (e.g. Westra, 1969; Bousquet, 1979; Bordet, 1985; Rutter et al., 1986; Weijermars, 1987a–c, 1988b). Figure 26 shows the geological locations of the major strike–slip faults traced hitherto in SE Spain. These are the Alhama de Murcia, Palomares, Almería, and Baños Fault systems. The minimum lengths of their traces are 100, 80, 35 and 20 km, respectively. Plate 12 shows some characteristic features of strike–slip faults in the area.

Studies with portable seismometers near the *Alhama de Murcia Fault* in 1977 registered continuous seismic noise, including about 20 shocks with magnitudes between 3 and 5 (Mezcua et al., 1984; cf. Bousquet, 1979). The main shock of magnitude 4.2 occurred on 6th June, 1977, and was felt over an area of 3000 km^2. The hypocentres of the 50 best located aftershocks showed a cluster at 6–8 km focal depth and suggests that the Alhama de Murcia Fault dips 45° NW. The focal mechanism solution indicates predominant dip–slip motion (Mezcua et al., 1984). The Alhama de Murcia Fault may also have generated the earthquake which destroyed Murcia in 1829 (Fig. 26). The 1829 earthquake caused at least 1214 casualities and damaged 5000 houses of which 3500 were in Murcia alone (Udías et al., 1976).

Detailed studies of the *Palomares Fault* suggest that this fault has been active since at least 8 Ma B.P. (see extensive treatment in Weijermars, 1987a). The historic earthquake of 1518 which destroyed the ancient city of Vera (Bousquet, 1979) and the tower of the coastal guard in Garrucha (Grima-Cervantes, 1987) may have been generated by movement of the Palomares Fault. The earthquake of 1928 which caused leakage of oil into water-

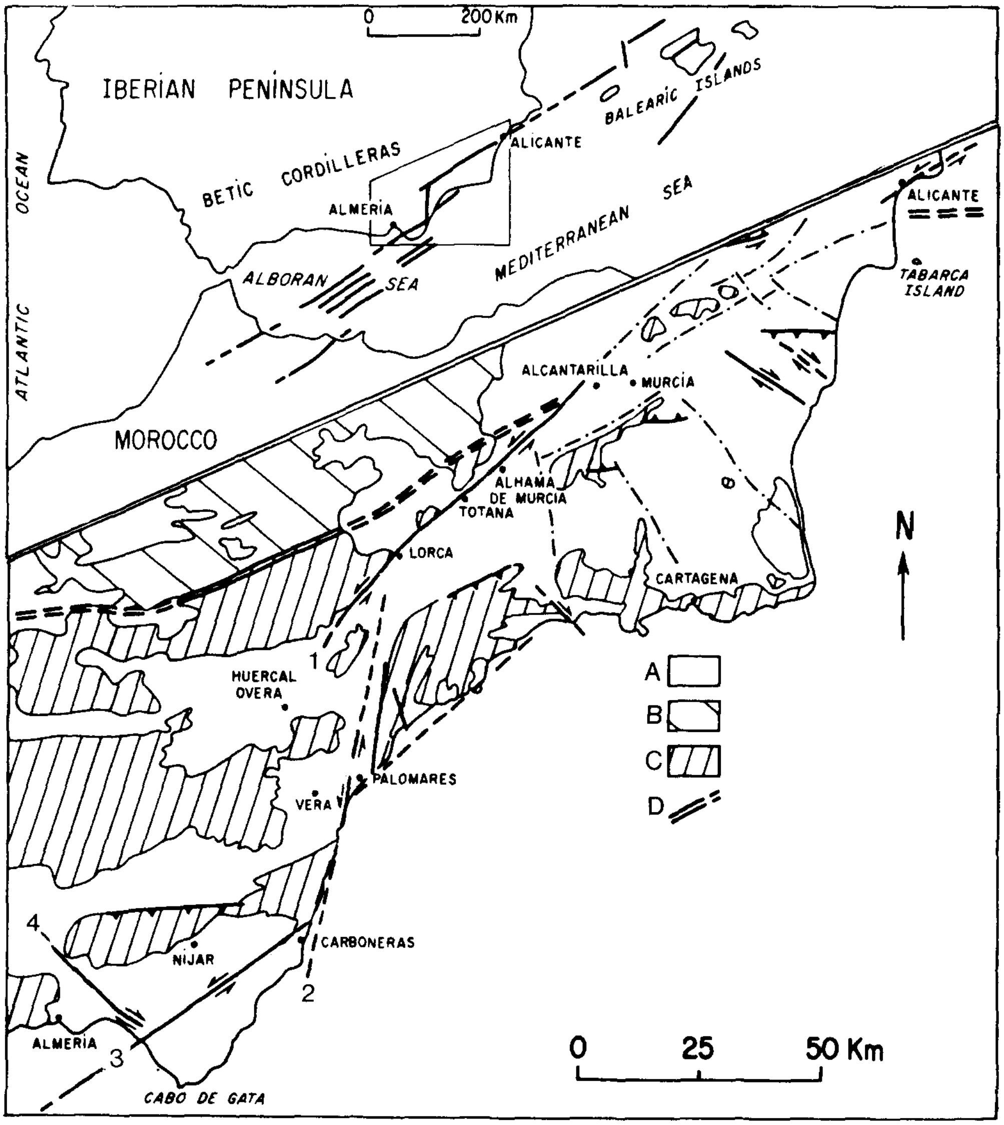


Fig. 26. Tectonic map showing the location of the major seismic faults in SE Spain (*1–4*). Neogene cover, Subbetic basement, and Betic basement are labelled A–C, respectively, in the legend. The Vélez Rubío Corridor (D), separates the Internal and External Zones, i.e. Betic and Subbetic basement, respectively. The faults labelled are: (*1*) Alhama de Murcia Fault, (*2*) Palomares Fault, (*3*) Almería Fault Zone, and (*4*) Baños Fault (after Bousquet, 1979, figs. 1 and 2).

wells of Garrucha (Dupuy de Löme, 1933) must also be connected to activity of the Palomares Fault. The time-averaged relative displacement rate between the walls of the Palomares Fault has been estimated at 2 mm a^{-1} and the time-averaged shear strain rate is 6×10^{-12} s^{-1} (Weijermars, 1987a,c). The sinistrial strike–slip component of the Palomares Fault is obvious from the regional map pattern in Figs. 17 and 26. Fold axes in the Tortonian Chozas Formation gradually rotate over 20° from E–W in the Sorbas to N70°E in the Vera Basin (Rondeel, 1965). Tortonian and Messinian strata gradually change in both strike and dip towards that of the subvertical Palomares Fault on approaching the fault plane (Fig. 19). The total horizontal shear displacement is in the order of 25–30 km (Weijermars, 1987a).

The *Almeria Fault Zone* (Fig. 24) comprises two parallel main faults which enclose a band of faulted volcanic rocks and Torto-

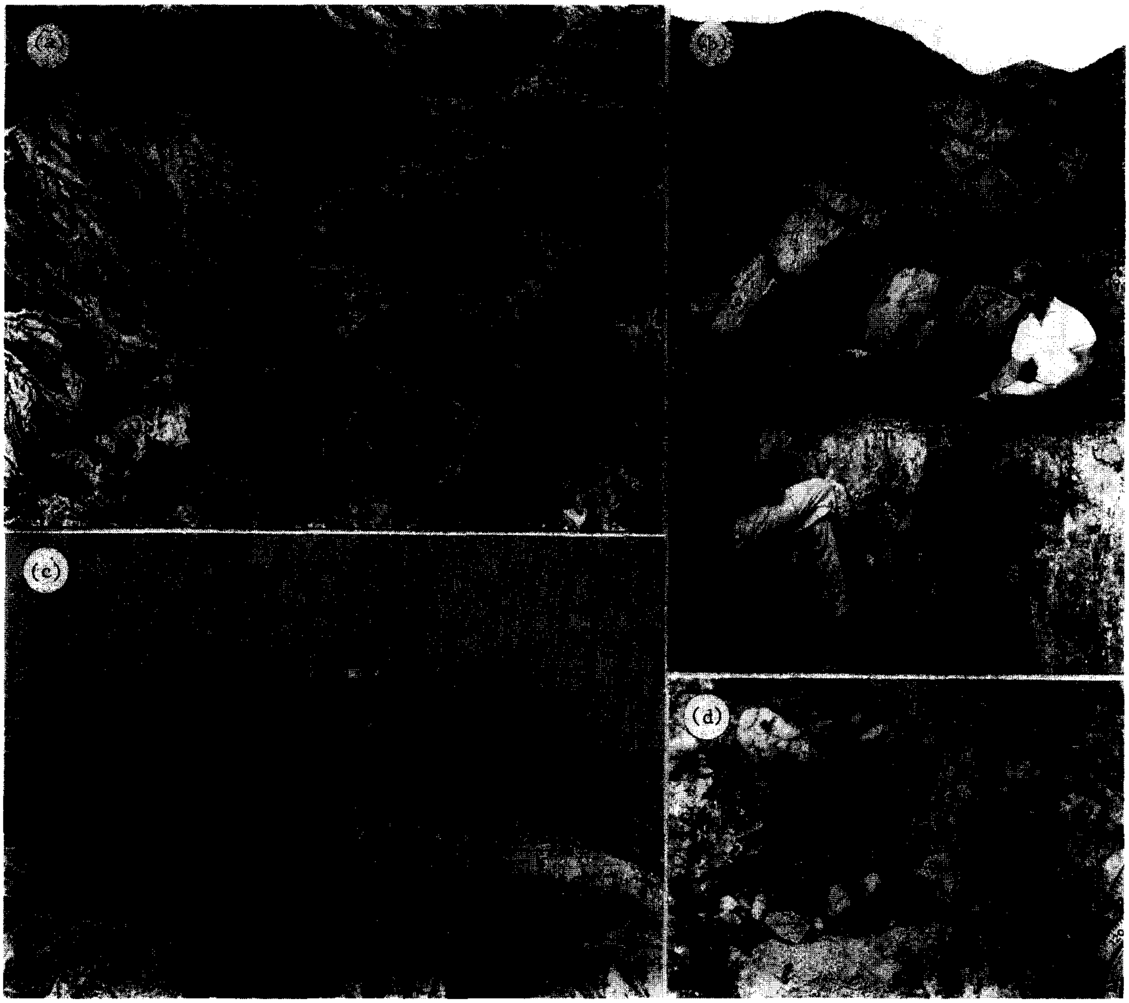


Plate 12. Strike–slip faults. (a) Aerial photograph of the Almería Fault Zone in the Sierra Serrata (el. 300 m), running NE–SW across the image. Geographical north is up. Flight altitude is about 8 km, original scale of resolution is 30 cm. Spacing between the two main fault strands is about 1200 m. Systematic left-lateral off-set of the drainage pattern is particularly obvious at the trace of the northern fault strand. (b) Looking north along the NE–SW strike of the Almería Fault Zone, Rambla de Granatilla, SE Sierra Cabrera. The fault zone is demarcated by a 50 m wide zone of cataclastic rocks separating Palaeozoic schists in the western wall from Neogene volcanic rocks in the east. The central part of the fault zone comprises virtually all the lithologies in the region, which have been sampled by the strike–slip motion. It contains cataclastic slices of red Malaguide, purple Alpujarride, black Nevado-Filabride, yellow Neogene sedimentary rocks and yellow-brown volcanic rocks. Episodic movements on this fault may have caused the historic earthquakes which destroyed the city of Almería in 1487, 1522, 1659 and 1804. Geologists are Stefan Bergman and Roberto Weinberg. (c) El Hacho (el. 192 m), Vera Basin. Thick bedded resedimented sandy gravels of upper delta slope in Gilbert-type fan of the Espiritu Santo Formation. Planar to tangential foresets indicate sediment is supplied by slope to the east, right-hand side of the picture. However, the east is presently occupied by Mediterranean Sea which led Völk (1966a) to suggest that the Vera Basin was a nearly closed marine bay during the Late Pliocene. This closure may be explained by left-lateral strike–slip displacement of the former eastward continuation of the Sierra Cabrera, now termed Sierra Almagrera, by the Palomares Fault. Looking north from Cabezo de la Fuente del Lobo (190 m), 800 m south of El Hacho and 1 km east of Vera. (d) Fractured and horizontally dislocated pebble in Espiritu Santo Formation of El Hacho, Vera Basin. This suggests fault displacement continued at least until after consolidation of the fanglomerate sedimentary rocks.

nian–Messinian sedimentary rocks (cf. Zeck and Soediono, 1970) exposed in the horst structure of the Sierra de Serrata (Bordet, 1985). The term Almería Fault zone adopted here is from Baena et al. (1977) and Greene et al. (1977). It has also been referred to as the Carboneras Fault Zone (Bousquet et al., 1975; Bousquet and Philip, 1976a,b; Rutter et al., 1986) and the Serrata Fault (intern. repts., Free University, Amsterdam). The term Carboneras Fault Zone is inconvenient as the city of Carboneras lies closer to the strike of the Palomares Fault than to the former fault (c.f. Fig. 19). The fault fabric comprises cataclastic and clay-rich fault gouges and syngenetic riedel shears indicate sinistral strike–slip motion (Rutter et al., 1986). The drainage pattern of the Sierra Serrata is locally displaced by the Almería Fault Zone in sinistral sense (Baena et al., 1977; Bousquet, 1979; Bordet, 1985). The river system cannot have existed before the Quaternary, since the rocks elevated in the Sierra Serrata stood in a shallow sea during the Pliocene. This suggests that a left-lateral strike–slip displacement of at least 200 m has occurred in the Quaternary, implying a minimum time-averaged shear rate of 0.1 mm a^{-1}. The Almería Fault Zone may have caused the earthquakes which destroyed the city of Almería in 1487, 1522, 1659, and 1804 (cf. Udías et al., 1976).

The northeast strand of the Almería Fault Zone has been studied in detail by Westra (1969) who left it nameless. Westra (1969: p. 73) distinguished two stages of fault activity. A first phase produced NE striking, subvertical faults, imbricating 10–20 m wide zones of sedimentary rocks containing pelagic foraminifers indicative of Older Neogene age (i.e. *Operculina* sp., *Rotalia* sp., Westra, 1969: p. 64). Calc-alkaline volcanic rocks have intruded the Older Neogene sedimentary rocks (Burdigalian), creating a 10 cm wide dark contact zone containing chlorite (Westra, 1969: p. 64). The first phase of faulting must be post-Burdigalian (i.e. < 16 Ma B.P.), but older than the Early Messinian Azagador conglomerates, discordantly covering some of the fault traces (i.e. > 7 Ma B.P.). A second faulting episode locally tilted and imbricated Messinian Azagador sedimentary rocks (Westra, 1969: p. 73) and seems predominantly characterized by strike–slip faulting (Westra, 1969: p. 74).

Westra (1969) further mapped metamorphic isograds separating sillimanite-, andalusite-, staurolite- and chloritoid zones. These particular metamorphic zones occur only in graphite-micaschists of the Sierra Cabrera basement trapped in and near the Almería Fault Zone. The isograds are closely spaced and subparallel to the main fault contact separating the micaschists from the bulk of the Neogene volcanic rocks south of the Almería Fault Zone (Fig. 27). There is also a genetic relationship between the formation of leucogranitic dykes and the metamorphic contact zones. This was concluded from the observation that the size of plagioclase porphyroblasts in the sillimanite-K-feldspar zone increases considerably in the vicinity of the leucogranitic dykes (Westra, 1969: p. 126). The bulk composition of all leucogranitic dykes is constant even if they are found 3.5 km apart, and this suggests that they are offshoots from the same mass of highly fractionated basic magma at depth (Westra, 1969: p. 123). These leucogranitic dykes have radiometric ages of 20–22 Ma (Bellon et al., 1983; thus being younger than the 50 Ma previously suggested by Westra, 1969). This suggests that Westra's (1969) first phase of faulting was preceeded by the intrusion of the leucogranitic dykes associated with the development of metamorphic contact zones above a magma front.

Another potential source of earthquakes is seismic motion on the *Baños Fault*. This fault may have played an important role in maintaining the slope instabilities necessary for the depositional environment of marine Pliocene conglomerate fans in the Ríoja corridor (Fig. 23a). It still seems active, as the temperature of the Baños thermal spring was raised from 42° to 53°C after an earthquake in 1865 (Tapia Garrido, 1980). Similarly, the occurrence of thermal springs in Alhama de Almería, NE of the Sierra de Gádor, has also been explained by frictional heating on fault planes (Sanz de Galdeano et al., 1985).

There are two potential sources of geological hazards threatening public safety and property in Almera province. These are (1) landsliding, and (2) earthquakes, both of which have been insufficiently documented. For example, Turrillas, a small village at the north slope of the Sierra Alhamilla, is situated below a potential landslide mass of 10^6 m^3. The potential surface of sliding is located under Cerro Minuto, near the 1093 m summit

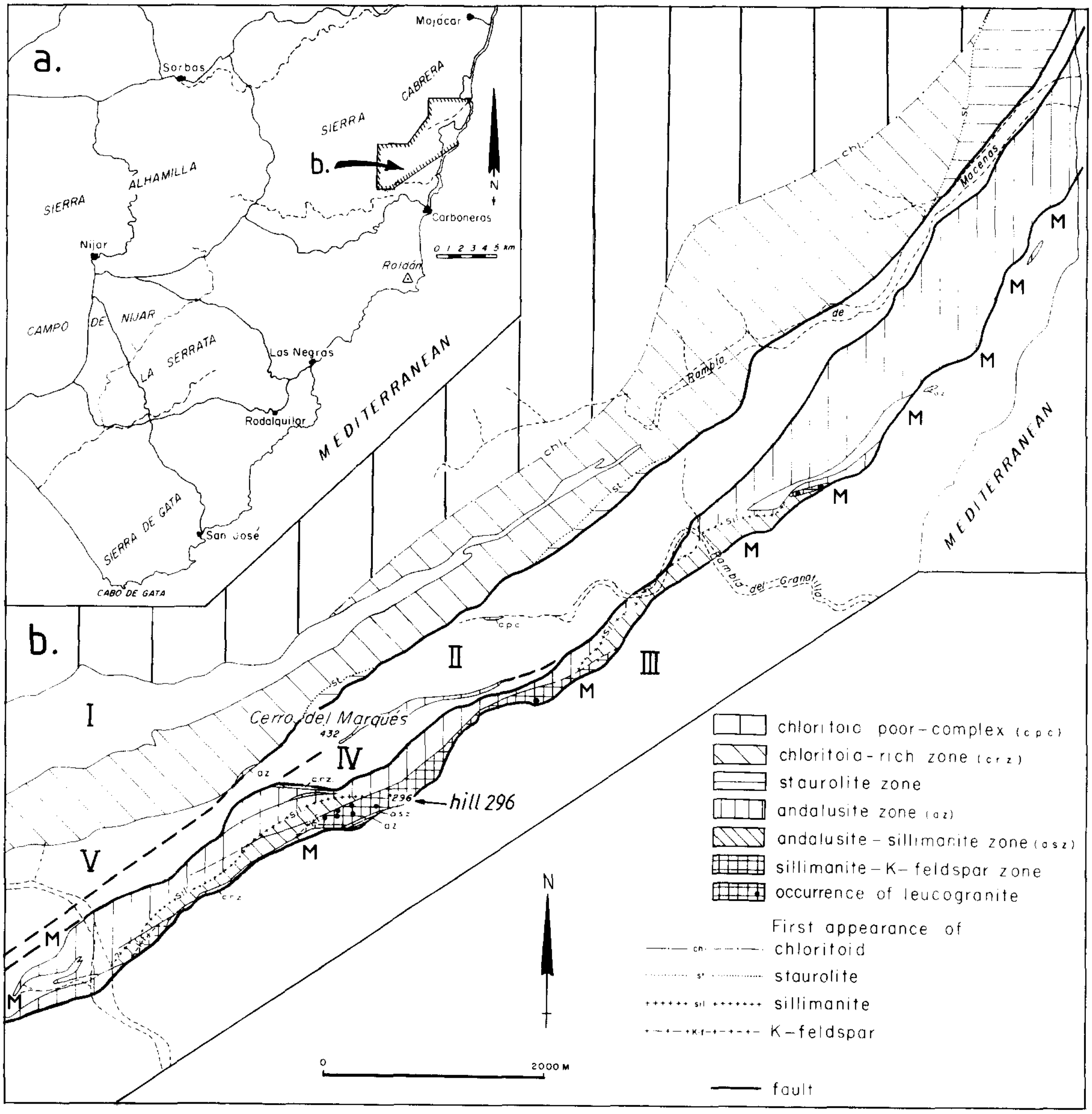


Fig. 27. (a) Location map of detailed area, shown in (b) Mineral zones along the trace of Almería Fault Zone in Nevado-Filabride micaschists of the Sierra Cabrera. Leucogranites and leucocratic migmatites occur in the sillimanite-K-feldspar zone along the trace of the main south fault. Roman capitals in the blank areas are portions of (*I*) Castro Zone and Alpujarride rocks, (*II*) Alpujarrides, (*III*) Neogene volcanic rocks, (*IV*) Older Neogene sedimentary rocks, (V) Younger Neogene sedimentary rocks (Azagador, Abad, and Yesares members). Slivers of Malaguide rocks (*M*) are almost entirely confined to the trace of the main south fault (modified from Westra, 1969, enclosure 2, fig. 1 and his geological map).

above Turrillas. Incipient surface movement is indicated by open cracks near the oversteepened edge of Cerro Minuto. Rapid landslide movements may be triggered by episodic seismicity of the area. Remnants of several previous landslides have been mapped at the foot of Cerro Minuto (Weijermars, 1981). The danger that new landsliding may erase Turrillas prompts detailed studies to estimate the safety factor of Cerro Minuto.

Historic examples of casualities and material damage caused by earthquakes in Almería province are abundant. It therefore seems advisable to determine and classify the seismic risk zones of the area. This was not possible until recently, when the exact location of the fault planes causing the earthquakes became apparent. A still better understanding of the seismic faults in the region would require preparation of detailed maps showing heat flow, aeromagnetics, and sidelooking radar data. Such studies should include a search for historic records on ancient earthquakes and documentation of the hydrothermal springs in the region. Detailed seismic reflection profiling and diamond core drilling also may help to improve knowledge of the subsurface structure and dip of fault planes.

However, only continuous monitoring of the seismicity of these fault planes will allow assessment of how dangerous the seismic faulting in SE Spain is at present. More reliable seismic data are becoming available since the installation of the Andalucian Seismic Network in 1983, which records microearthquakes with epicentral errors less than 3 km (e.g. De Miguel et al., 1983). It seems also advisable to install a permanent network of seismometers along the trace of the active faults in the area. In the meantime, the active strike–slip faults discovered so far might be put to man's use by studying possibilities for extracting geothermal energy from the hot movement zones (i.e. the Alhama de Murcia, Palomares, Almería, and Baños Faults).

8. GEODYNAMIC HISTORY

8.1. Western Mediterranean plate configuration

To most earth scientists the nappe structures, uplift of the Betic Zone s.s. and the polarity of the Betic Cordilleras as a whole, are due to the impingement of the African and Eurasian plates in the Straits of Gibraltar (e.g. Dewey et al., 1973; Tapponier, 1977). Established opinion further assumes that the boundary between the African and Eurasian Plates in the western Mediterranean is formed by the Azores Transform, which would extend through the Straits of Gibraltar (Fig. 28). This E–W trending plate boundary, initially proposed on the basis of a scattered distribution of earthquake epicentres, is now displayed in all textbooks and atlases of plate tectonics. But does this plate boundary really exist where geophysical studies suggest it to be? Geological field surveys can find no trace of an E–W trending transform fault transecting the Straits of Gibraltar. That the nature of the Mediterranean extension of the Azores Transform can be debated, was also evident from emotional discussions surfacing at a recent meeting of the European Geophysical Society (Barcelona, 13–17 March 1989). Being one of the opponents criticising the established view (Weijermars, 1987a,b, 1988b, 1989a,b), I would like to draw renewed attention to the problems connected with the interpretation of the plate configuration in the western Mediterranean and how these problems came into existence.

The picture of the Eurasian–African plate boundary in the western Mediterranean dominating in current literature is that inferred on the basis of seismic epicentre distribution in the late 1960's. Most earthquakes generated at shallow depths are due to episodic movement along fault planes (Sykes, 1967). A worldwide compilation of earthquake epicentres at the late 1960's indeed showed that their distribution for some 80%

of the data neatly coincided with the outlines of tectonic plates previously mapped by other means (Isacks et al., 1968; Morgan, 1968).

This finding was subsequently used to define the boundary between the African and Eurasian plates in the western Mediterranean, an area of extremely complex tectonics, on the basis of seismic epicentre data only. Widely spaced epicentres were connected by an imaginary line such as to represent the plate boundary (McKenzie, 1970, 1972). More specifically, the seismic Azores Transform Fault, marked by about 300 km of dextral offset of the spreading pattern in the Atlantic Ocean (Pitman and Talwani, 1972; Rona, 1980), was proposed to extend through the Straits of Gibraltar (approximately) and to separate the African and Eurasian plates in the western Mediterranean (Fig. 28). This procedure was in agreement with Morgan's (1968) discovery that the world's earthquakes are generally found along the boundaries of tectonic plates. Other compilations of the seismic epicenters in the western Mediterranean (Udías et al., 1976; Mezcua et al., 1980; Buforn et al., 1988) have essentially adopted the plate boundary first postulated by McKenzie (1970, 1972). However, a careful comparison of their maps shows that the exact location of this seismically inferred plate boundary is, in fact, poorly defined.

Unfortunately, the geological information available from detailed field studies in the area has apparently been neglected in geophysical studies attempting to determine the plate boundary in the western Mediterranean. Geological observations, in particular those that established the continuity of geological boundaries across the Straits of Gibraltar, contradict the existence of any major plate boundary in this area at present. For example, the outline of the flysch deposits shouldering the Betic–Rif orogen in the west can be traced across the Straits (Fig. 29), leaving no space for any significant offset. These deposits range in age from Late Jurassic to Tertiary (Bailey et al., 1951; Esteras Martin, 1984). Similarly, the outlines of the Messinian evaporite deposits on the floor of the W-Mediterranean, mapped by extensive seismic

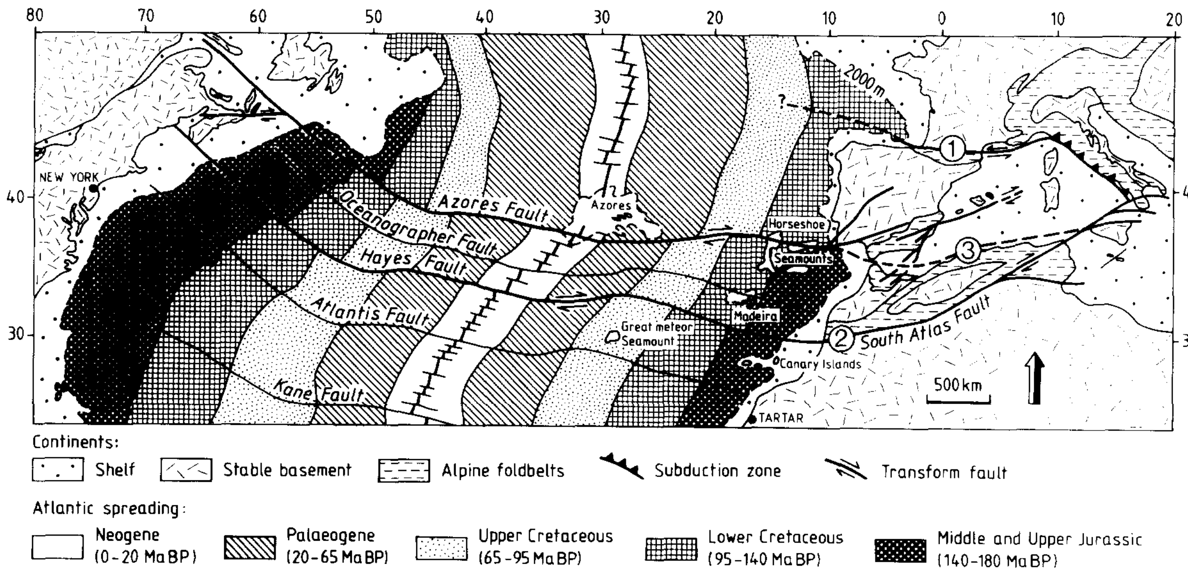


Fig. 28. Tectonic map of the North Atlantic spreading pattern and major active strike–slip fault traces. The Azores Transform Fault terminates near the Horseshoe Seamounts (Grimison and Chen, 1986), where it splays into a fault pattern exposed in southern Spain. The North Pyrenean Fault (Fault 1) and Hayes–South Atlas Fault (2) accommodate the motion of the Iberian microplate which indents Italy. The Hayes–Atlas Fault (2) is suggested as the modern northern boundary of the stable African plate assumed here. Fault 3 (dashed) was inferred on the basis of widely separated earthquake epicentres (McKenzie, 1972), but does not appear to exist (after Weijermars, 1987b, fig. 2).

reflection profiling (Auzende et al., 1975), show no apparent sign of any strike–slip displacement since their deposition 5 Ma ago (Fig. 30). A recent, higher resolution study of earthquake foci also suggests that the Azores Fault does not continue into the Straits, because no single narrow plate boundary can be identified there (Grimison and Chen, 1986).

Nonetheless, most of the earthquake foci in the Alboran region of the western Mediterranean are less than 60 km deep (Buforn et al., 1988) and therefore must be due to fault displacements in the crust (Sykes, 1967). These faults should therefore be visible at the surface as an abrupt displacement of otherwise smooth geological boundaries. The important task of the geologist now is to go into the field and show the geophysicist where these earthquakes actually come from and what they mean.

However, the geologist in his search for traces of these faults has little or no help from the seismic records because the epicentres of earthquakes in the western Mediterranean have not been located better than within a 25 km radius of the actual epicentres in the best cases (Mezcua et al., 1984). This is due to uneven station distribution, reading errors and

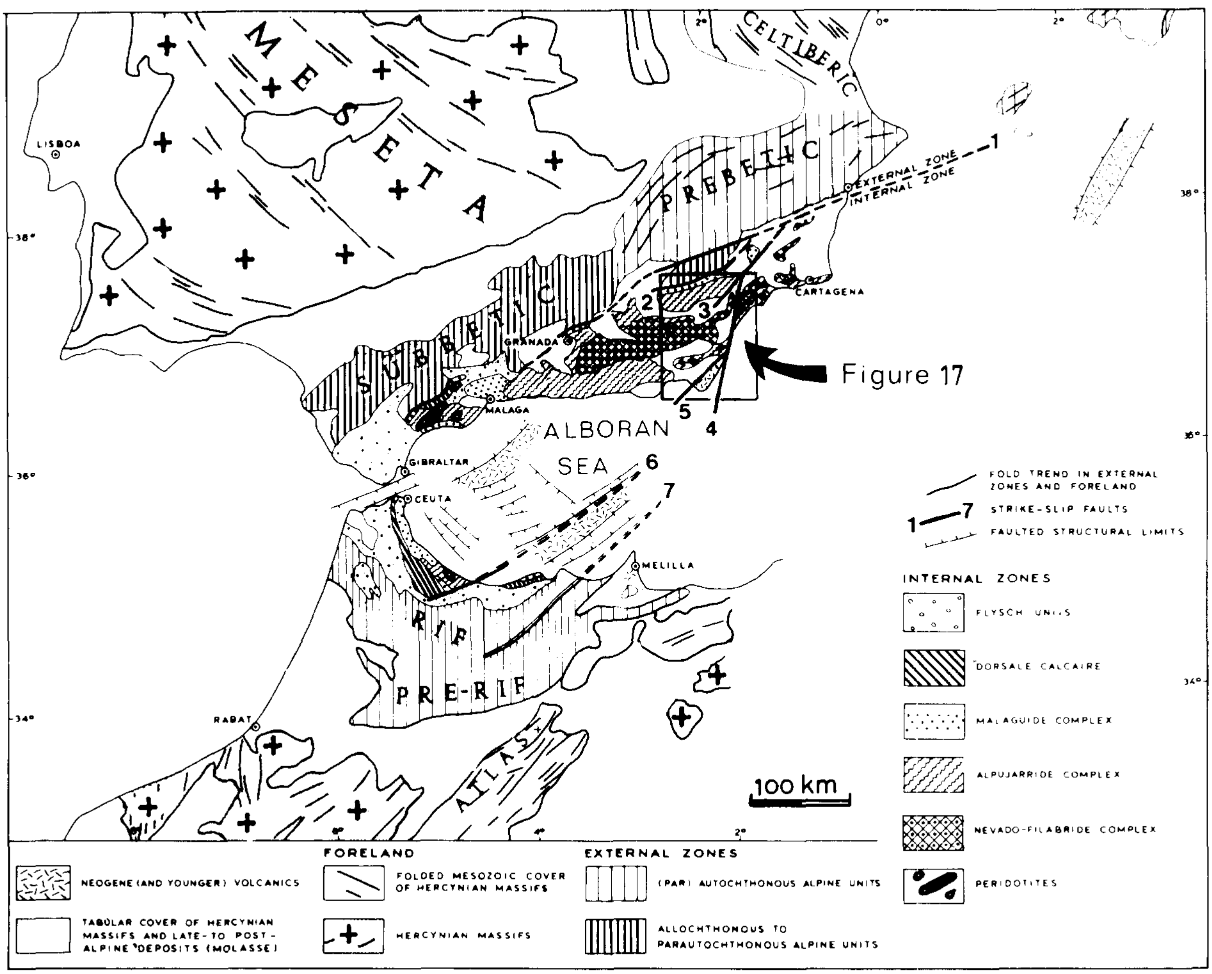


Fig. 29. Geotectonic map of the Arc of Gibraltar, showing the geology of Betic–Rif Mountains peripheral to the Alboran Basin. The Internal and External Zones of the Betic Cordilleras are separated by the Crevillente Fault (*1*). This fault splays into the Vélez Rubío Corridor (*2*), Alhama de Murcia Fault (*3*), Palomares Fault (*4*), Almería Fault Zone (*5*), and possibly the Jedha (*6*) and Nekor (*7*) Faults of Morocco. (after Weijermars, 1987a, fig. 1; modified from Kampschuur and Rondeel, 1975, fig. 1).

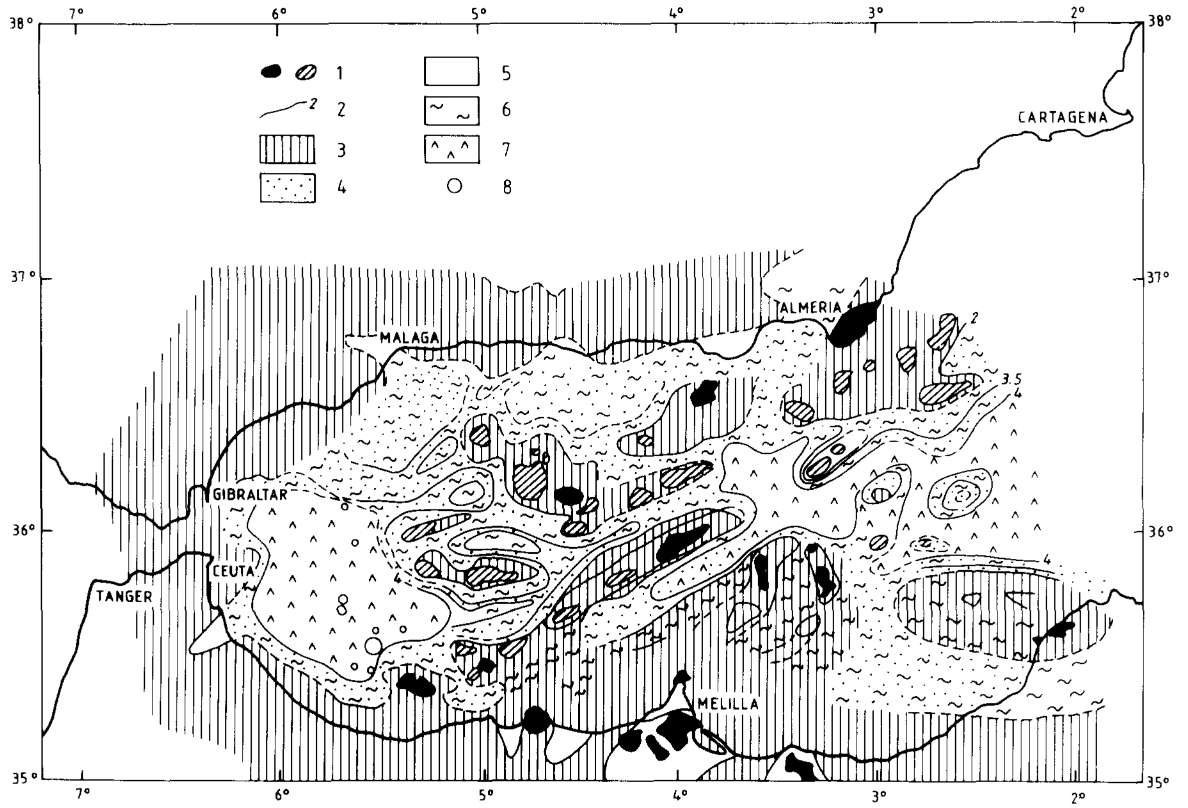


Fig. 30. Structural map of the Alboran Basin inferred from extensive seismic reflection profiling. Legend: (1) volcanic centres observed (black) and suggested (hatched); (2) isopach contours showing the thickness of the sedimentary cover above the crystalline basement in two-way travel time seconds; (3) zone where the basement is shallower than 1,500 m; (4) zones where the basement lies at 1,500 to 3,500 m; (5) zones where the basement is deeper than 3,500 m; (6) evaporite deposits; (7) halite deposits; and (8) salt domes (modified from Auzende et al., 1975, fig 3).

lateral heterogeneities in the Earth. An exception was provided by the El Asnam earthquake of 1980 in Algeria. A geological mapping team was flown to the area of destruction within 48 hours after the occurrence of the earthquake with magnitude 7.2. They located the fresh surface break of the earthquake, which was an NE–SW trending fault scarp (Frechet and Philip, 1981; King and Vita-Finzi, 1981; Ouyed et al., 1981, 1983; Nâbelêk, 1985). This yielded the first clear example of an earthquake epicentre located along the supposed trace of the plate boundary, but not corresponding to a E–W trending surface fault as would be expected according to the trend of the alleged plate boundary (Fig. 28).

Detailed geological observations in both North Morocco and Spain have revealed the presence of large fault structures, also with a NE–SW strike, similar to that of the El Asnam Fault. Visual illusion may be misleading in mapping of lineaments from remote sensing, because it is difficult not to confuse lineaments of a different nature and different ages without adequate ground control. Figure 31 shows the distribution of seismic epicentres together with the traces of several major Neogene strike–slip faults largely determined by ground-mapping abrupt discontinuities in otherwise smooth geological boundaries. Many other faults have been suggested but these were at variance with the geological data (see review by Weijermars, 1987b).

The above field observations suggest that the diffuse pattern of seismic epicentres in the W-Mediterranean is due to a network of smaller strike–slip faults. These faults seem

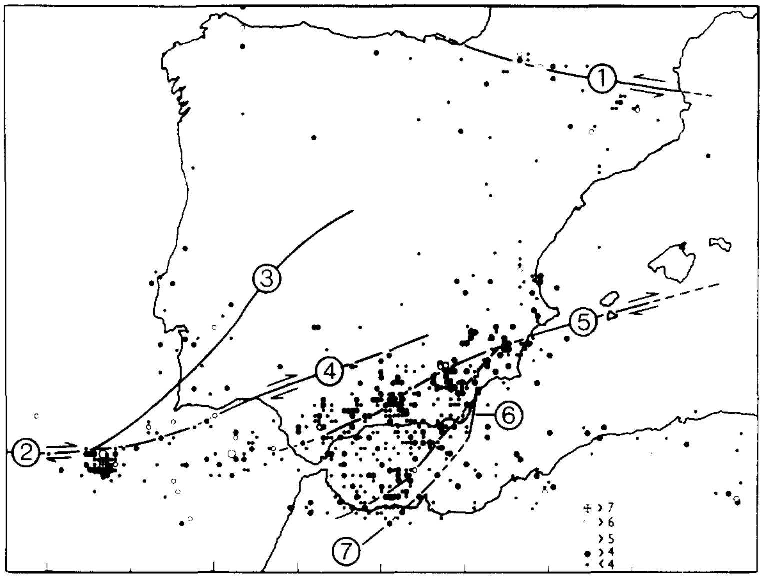


Fig. 31. Distribution and magnitude of earthquakes recorded by the Spanish Seismological Survey during the period 1961–1972. Major lineaments mapped so far are: (*1*) North Pyrenean Fault, (*2*) Azores Transform Fault, (*3*) Plasencia lineament, (*4*) Guadalquivir lineament, (*5*) Crevillente Fault, (*6*) Palomares Fault, and (*7*) Nekor Fault (from Weijermars, 1988b, fig. 1, using seismic data of Udías et al., 1976, and geological field observations by Westra, 1969; Bousquet, 1979; De Smet, 1984a,b; Harmand and Cantagrel, 1984; Bordet, 1985; Rutter et al., 1986; Pique et al., 1987; Weijermars, 1987a–c, 1988b; De Larouzière et al., 1988; Montenat et al., 1988).

to be connected with the Azores Transform Fault, here assumed to terminate in a pattern of splay faults mapped in southern Spain. Previously it was suggested that this transform fault continued through the Straits of Gibraltar. The question now arising is: *If the present-day boundary between the Eurasian and African plates in the western Mediterranean is not coinciding with the geographical divide between Africa and Europe, where is it*? Some alternative hypotheses are outlined below.

Carey (1976: p. 254) used the Arc of Gibraltar to illustrate his orocline concept and assumed that the suture between Africa and Europe lay to the south of the Arc. Others have suggested that the Arc is part of a triangular block, i.e. the Alboran Block (cf. Bourrouilh and Gorsline, 1979), which is bordered to the north, south and west by faults allowing its movement as a wedge between Africa and Europe (Andrieux et al., 1971; Andrieux and Mattauer, 1973; Bourouilh and Gorsline, 1979). The faults bounding the Alboran Block were originally entirely hypothetical in these models, but some evidence for the existence of two of the three necessary faults has appeared recently.

LeBlanc and Olivier (1984) have argued that the Alboran Block lies between the NE–SW Crevillente Fault in the north and the NNE–SSW Nekor Fault in the south (cf. Fig. 29). A weakness in the model proposed by LeBlanc and Olivier (1984) is that neither the Crevillente nor the Nekor Faults can be traced over the distances indicated on their maps on the ground. The Nekor Fault is only exposed for 100 km and seismic activity (Hatzfeld and Bensari, 1977) does not support LeBlanc and Olivier's (1984) suggestion that the Nekor Fault may be 700 km long.

The northern Crevillente Fault (Fig. 29) is indeed now recognized in the field as a major brittle shear with a dextral strike–slip displacement of at least 250 km – principal movement occurred in the Neogene between 20 and 5 Ma B.P. (Hermes, 1978, 1984; Van de Fliert et al., 1980; De Smet, 1984a,b). However, it seems to terminate abruptly in

TABLE 2

Geologic time scale outlining the geological history of SE Spain

Ma B.P.	Epoch/Period/Stage	Geological event
Present		Modern earthquakes and hydrothermal springs
Quaternary		Establishment of modern shorelines
Neogene (20–2 Ma B.P.)		
3–2	Late Pliocene	Alluvial fans after regression
2.8–2.6	Middle Pliocene	Cartagena basalts
5–3	Early Pliocene	Marine fan deltas; intensified strike–slip movements
5	Messinian/Pliocene	Opening of Straits of Gibraltar; veritic lavas; angular unconformity
5.5–5	Late Messinian	Salinity Crisis (2nd stage): desiccation of Mediterranean
6.5–5.5	Messinian	Salinity Crisis (1st stage): cyclic evaporization
6.5	Early Messinian	Reef complexes mark transgressional shoreline
7	Tortonian/Messinian	Angular unconformity Tabernas Basin
8–7	Late Tortonian	Crustal shortening within Iberian microplate moving west along converging strike–slip faults, causing: (A) Thrusting of Subbetic onto Prebetic (B) Narrowing of the Betic Straits (C) Formation of Basin and Range structure in Betic Mountains
8	Tortonian	Palomares Fault active
12–7	Tortonian	Ignimbrites and hydrothermal gold mineralisation of Rodalquilar; turbidite fan complexes
15–12	Serravallian	Calc-alkaline calderas formed, due to passage of isotherms in lower crust
15	Burdigalian/Serravalian	Sierras Nevada and Filabres rise above sea level: first occurrence of Nevado–Filabride detritus; angular unconformity Older/Younger Neogene
20–15	Burdigalian	Onset subsidence Alboran Basin, uplift Betic–Rif orogen
20	Aquitanian/Burdigalian	Onset Crevillente strike–slip by movement of Iberian microplate
22 ± 2	Middle Aquitanian	Intrusion of leucogranitic/aplitic dykes
22 ± 4	Middle Aquitanian	Emplacement of Ronda peridotite in hot thrust slice
25–20	Aquitanian	Alboran Diapir creates crustal high, northward emplacement of Higher Betic nappes under (LP-LT)-conditions
Palaeogene (65–20 Ma B.P.)		
35–25	Oligocene	Deposition of youngest Malaguide sedimentary rocks
36	Late Eocene	(A) Final closure of Tethys in western Mediterranean (B) Formation of Atlas Mountains
Cretaceous (140–65 Ma B.P.)		
85–80	Late Cretaceous	Emplacement of Nevado–Filabride nappes by subduction and obduction under (HP-LT)-conditions
140–85	Early Cretaceous	Spreading in Tethys with passive margins
Jurassic (200–140 Ma B.P.)		
146 ± 3	Late Jurassic	Basic dyke swarms in Nevado–Filabrides: Tethys opened
200–140	Aalenian–Tithonian	Rifting in central Subbetic – inferred from basic intrusives
Triassic (250–200 Ma B.P.)		
230–225	Carnian	Deposition of Almagride sedimentary rocks
235–230	Ladinian	Deposition of Alpujarride carbonates
Palaeozoic (600–250 Ma B.P.)		
269 ± 6	Early Permian	Intrusion of Bedar granite in connection with Hercynian orogeny
600–270	Palaeozoic	Deposition of Nevado–Filabride sedimentary rocks
Precambrian (?–600 Ma B.P.)		
800–600	Late Proterozoic	Gneissic basement formed

southwestern Spain where Tertiary flysch deposits are undisturbed by the Neogene Crevillents Fault (De Smet, 1984a,b). The Guadalquivir lineament en-echelon to Crevillente Fault (Fig. 31) may be geodynamically connected by means of the Guadalquivir Basin lying in between (Weijermars, 1987b). The Guadalquivir lineament in turn may join the Azores Transform Fault to the west.

I have also argued that it is not the Azores Fault, but the Hayes–Atlas Fault which might be the principal zone of movement between the Eurasian and African plates (Weijermars, 1987b). Geological observations show that the Alpine Atlas Mountains are bordered to the south by a continuous 2500 km long lineament, termed the South Atlas Fault (e.g. Weijermars, 1987b). The Atlas Fault is a land fracture which may continue into the oceanic crust and join up with the Hayes Transform Fault near Agadir (Fig. 28). A major earthquake destroyed parts of Agadir in the early 1960's. An epicentre map of the Mediterranean region, published by Buforn et al. (1988), shows that there are earthquakes along the entire length of the South Atlas Fault.

The North Pyrenean Fault is also seismically active which could be used to suggest this fault as the southern boundary of the stable foreland of Europe. More specifically, the Hayes–Atlas Fault may be the modern northern boundary of the stable part of the African plate, and the North Pyrenean Fault could border the stable foreland of Europe (Fig. 28). The area between these faults might be interpreted as the Iberian microplate which indents Italy in a fashion similar to the impingement of the Indian plate with Tibet prior to continental collision. This proposition would reduce the size of the African plate by 2 million km^2 (Weijermars, 1988b).

Note that continuous spreading at the Atlantic ridge and movement along these two strike–slip zones (bordering the Iberian microplate to the north and south) would cause subduction of the Mediterranean oceanic crust under the continental crust of Italy. Possibly, this could also explain the modern volcanism along the coast of western Italy. Perhaps even the enigmatic curvature of the W-Alps, might be due to this collision with the Iberian plate superposed after the formation of the Alps.

8.2. Tectonic history of the Arc of Gibraltar

The formation of the fold belts in the tectonic Arc of Gibraltar seems no longer to be connected straight forwardly to the modern boundary between the Eurasian and African plates. Table 2 gives a geological time scale outlining the geological history of southeastern Spain. A possible tectonic scenario will be discussed below. The major features to be accounted for are (the references given are to studies which suggested first the absolute timing indicated):

(1) HP-LT metamorphism in the Nevado-Filabrides 80–85 Ma B.P. (De Jong, 1987);

(2) metasedimentary nappes and slices of peridotite emplaced in the Betic-Rif orogen 25–20 Ma B.P. (Weijermars, 1985a,b, using Priem et al., 1979);

(3) subsidence of the Alboran Basin, now enclosed by the Arc of Gibraltar since at least 15 Ma B.P. (Weijermars, 1985b);

(4) coeval uplift of the nappes in the margins of the Alboran Basin in such a fashion as to form the Arc of Gibraltar (Weijermars, 1985b);

(5) crustal emplacement of calc-alkaline volcanic rocks 15–7 Ma B.P. (Bellon et al., 1983);

(6) active strike–slip faulting since at least 8 Ma B.P. (Weijermars, 1987a);

(7) thrusting of the Subbetic on to Prebetic about 7 Ma B.P. (Hoedemaeker, 1973);

(8) the Messinian Salinity Crisis 6.5–5 Ma B.P. (Van Couvering et al., 1976).

Pangea remained a single continent throughout the Permo-Triassic so that any tectonic rearrangement in the relative position of rocks in the future Gibraltar Arc could not take place until the opening of Tethys in the Middle Jurassic (Weijermars, 1989b). Unaltered olivine diabase in the core of metabasite in deeper units of the Betic

Mountains in SE Spain yields a Rb–Sr whole-rock age of 143 Ma (Hebeda et al., 1977). This diabase could be the remnant of a basic dyke swarm weakening the continental crust coeval with rifting of Tethys, analogous to the geodynamic implications of basic dyke swarms reported from elsewhere. Jurassic rifting in SE Spain is also suggested by the occurrence of 200–140 Ma old basic intrusions in the central Subbetic (Puga et al., 1988b). The opening of Tethys initially only involved passive continental margins, but finally subduction led to juxtaposition of HP-LT rocks – now exposed in the Higher Nevado–Filabride nappes of SE Spain – about 80–85 Ma B.P. (De Jong, 1987). Tethys may not yet have closed entirely, but it is certain that the Atlantic continued opening at the expense of Tethys. The formation of the Atlas Mountains in the Early Oligocene 36 Ma B.P. (Dewey et al., 1973) could manifest final closure of Tethys. Any Tethyan suture was apparently unimportant in the Neogene tectonic evolution that followed.

Although lithospheric stretching preceeds basin subsidence by cooling in many areas

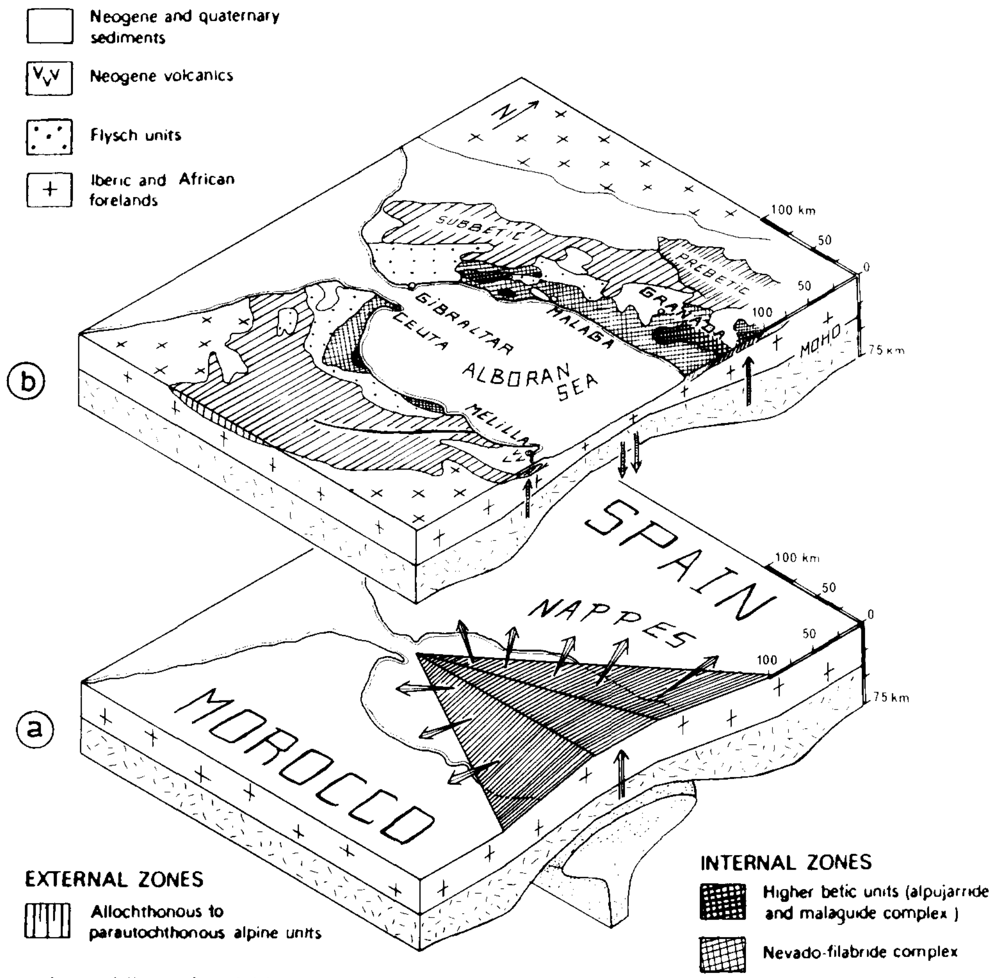


Fig. 32. Schematic equidimensional block diagrams illustrating the Neogene geodynamics of the Alboran Basin and the peripheral Betic–Rif orogen. (a) Nappe-shedding was caused by lithospheric bulging above a mantle diapir emplaced about 20–25 Ma B.P.; (b) subsidence of the Alboran Basin followed upon the tectonic unroofing by nappe-shedding after cessation of the diapiric mantle rise and onset of lithospheric cooling about 15–20 Ma B.P. Vertical arrows indicate the directions of Neogene isostatic motions. Uplift of the Betic–Rif Mountains occurred by isostatic recovery of the downwarps peripheral to the Alboran Diapir according to a mechanism discussed in detail elsewhere (Rowley and Sahagian, 1986; Parmentier, 1987; Keen, 1987a,b). From Weijermars (1987a, fig. 2).

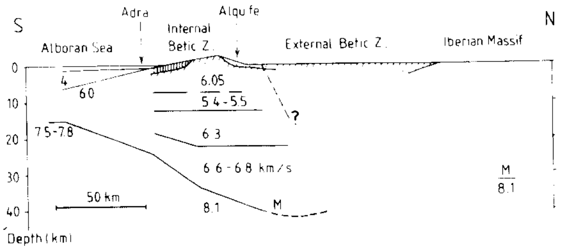


Fig. 33. Interpretative section of the crust under the Betic Cordilleras, according to seismic data. The Moho (*M*) is only 13 km deep under the floor of the Alboran Basin and thickens rapidly to over 40 km under the Betic Mountains. Note that the vertical scale has been exaggerated (after Vegas and Banda, 1982, fig. 13).

(e.g. North Sea Basin), the Neogene subsidence in the Alboran region is more likely to be due to lithospheric doming without any lateral rifting. Both the morphology of the subsiding Alboran Basin and its Bouguer gravity anomaly pattern (Bonini et al., 1973) suggest an anomalous upwarp in the top of the mantle. Seismic sounding experiments indeed show a bulge in the Moho of 17 km amplitude and 300 km N–S half wavelength beneath the Alboran Sea (Banda et al., 1983). The geology and crustal structure of the Arc of Gibraltar is summarized in the isometric block diagram of Fig. 32b. Crustal thicknesses to the north and south of the Alboran Basin increase rapidly to about 30 and 40 km under Morocco and Spain, respectively (Fig. 33).

There seems to be general agreement that at least the Higher Betic nappes were emplaced 20–25 Ma B.P. by nappe shedding from an ancient topographic high on the present site of the Alboran Basin (Fig. 32a). This view is supported by polarity in the transport direction of the nappes exposed in the Betic–Rif orogen (Torres-Roldán, 1979; Platt et al., 1983). The ancient Alboran High is assumed to be due to thermal bulging of the crust above a diapiric intrusion of buoyant mantle rock (Van Bemmelen, 1954, 1973; Loomis, 1975). The reversal of the topographic relief into that of the modern Alboran Basin can be understood as follows. The tectonic unroofing caused rapid denudation and crustal thinning above a mantle diapir. The mechanism of subsidence suggested is crustal cooling after thinning of the lithosphere, by thermal upwarp from below and coeval mechanical erosion at the surface (Weijermars, 1985a,b).

A group of workers led by John Platt suggests that this ancient high would not be due to crustal doming and thinning above a mantle diapir, but rather to crustal thickening by continental collision between Africa and Europe (Platt et al., 1983; Platt, 1986, 1987, 1988; Platt and Behrmann, 1986; Platt and Vissers, 1989). The occurrence of HP-LT metamorphism in the Higher Nevado–Filabrides is in support of a crustal thickening model, and thus can well be explained by the subduction-obduction model of Platt (1986). However, Platt (1986) did not realize that emplacement of the Higher Nevado-Filabride units during the HP-LT metamorphic event 80–85 Ma B.P. (De Jong, 1987) seems to predate the formation of Higher Betic nappes by some 60 Ma. Continental collision should therefore be separated from the later doming and crustal thinning by thermally driven diapirism in the upper mantle. This sequence of events would also provide time for the Nevado–Filabride complex to establish glaucophane schist facies metamorphism before undergoing greenschist facies metamorphism together with and during, the emplacement of the Higher Betic units (c.f. Fig. 10).

Crustal thickening of the former Alboran high by a plate collision as late as 20–25 Ma B.P. seems further unlikely because it cannot explain the topographic reversal from a high to a depression. The thinning of the thickened crust by subsequent spreading, invoked by Platt (1986) to explain the subsidence of the Alboran Basin, would need to be too fast to be realistic. The argument applies particularly if the orogenic root under the alleged Alboran collisional ridge were to be removed by convective peeling of the mantle part of the lithosphere, a model suggested by Platt and Vissers (1989). The authors do not seem to have realized that their model would initially cause a second phase of thermal uplift of the surface (after the nappe emplacement) due to the convective thinning. Subsequent subsidence by cooling could not take place until after erosion of the newly bulged surface. Such surface denudation requires a long period of extensive sedimentary erosion if not due to tectonic unroofing. The latter would be exactly the sequence of events implied by Weijermars' (1985a,b) nappe-shedding model. Nappe emplacement from the collisional high 20–25 Ma B.P. was directly followed by onset of subsidence of the Alboran seafloor between 22 and 15 Ma B.P. (see later), so that the model of Platt and Vissers (1989) is physically unfeasible. Emplacement of the Higher Betic units onto the Nevado–Filabrides 20–25 Ma B.P. can be accounted for by the nappe-shedding model of Weijermars (1985a,b). Nonetheless, *confirmation of the absolute timing of the HP-LT event and emplacement of Higher Nevado–Filabrides seems extremely important since it would help to settle the current controversy concerning the mechanism causing the modern upheaval of the Betic Cordilleras.*

The age of the crustal doming and associated shedding of Higher Betic nappes is well constrained. The youngest rocks comprised in the nappe sheets of the Betic orogen are 25 Ma old Oligocene sedimentary rocks (i.e. in the Malaguide Complex of the Betic Zone s.s.; see Egeler and Simon, 1969a,b). The Alboran surface bulge must obviously be younger than the youngest rocks translated by the nappe shedding it caused. It therefore cannot be older than 25 Ma. Further constraints follow from the 22 ± 2 Ma Sm–Nd cooling age of a garnet clinopyroxenite in the Ronda massif (Zindler et al., 1983). The enveloping rocks show aureole metamorphism which preceeded and accompanied its emplacement in a thrust sheet (Westerhof, 1977; Lundeen, 1978). Dykes of anatectic granites originating from Ronda's emplacement have a 22 ± 4 Ma Rb–Sr whole-rock age (Priem et al., 1970, 1979). Consequently, emplacement of the nappes took place about 22 Ma B.P.

The timing of the subsequent collapse of the Alboran dome (Fig. 32a) is also well constrained. Reconnaissance surveys have shown that the present-day Alboran Sea is about 1 km deep (Stanley et al., 1970) and is underlain by a basin filled with Neogene sedimentary rocks between 5 and 6 km thick (Mulder, 1973; Mulder and Parry, 1977). Drill core 121 taken by the Glomar Challenger for the Deep Sea Drilling Project near the base of the Neogene succession of the Alboran Basin contains Serravallian sedimentary rocks at least 15 Ma old (Ryan et al., 1970; Olivet et al., 1973a,b; Nesteroff and Ryan, 1973). Such sedimentary rocks were deposited on the subsiding crustal bulge already tectonically unroofed by nappe shedding. These age constrains imply that the Alboran Basin started to subside between 22 and 15 Ma B.P. Subsidence of the Ligurian and Balearic Basins occurred coevally after cessation of spreading connected with the rotation of Corsica and Sardinia 19 Ma B.P. (Rehault et al., 1984; Montigny et al., 1981).

Contemporaneous with the subsidence of the Alboran Basin, the Betic–Rif orogen or Arc of Gibraltar peripheral to the Alboran Sea was probably formed by isostatic recovery of the downwarps of the lithosphere peripheral to the Alboran Diapir (Fig. 32b). The geodynamics of such an uplift of a diapir shoulder has recently been discussed in detail by several workers (Rowley and Sahagian, 1986; Parmentier, 1987; Keen, 1987a,b). The generation of calc-alkaline volcanic rocks in southeastern Spain 15–7 Ma B.P. (Bellon et al., 1983; Di Battistini et al., 1987) can possibly be explained by partial melting of the base of the continental crust thickened in the return flow peripheral to the Alboran Diapir, but this idea needs to be tested by advanced modelling.

The location of the Cabo de Gata volcanic rocks may have been determined by preexisting zones of weaknesses created by previous intrusion of anatectic leucogranites. Migmatites and leucogranitic dykes have been mapped on hill 296 in Westra's (1969) field area along the Almería Fault Zone (Fig. 27) and the dykes have been dated radiometrically between 22 and 20 Ma (Bellon et al., 1983). The ages of these anatectic events suggest a correlation with Priem et al.'s (1979) dykes of similar age and composition, associated with the emplacement of the Ronda peridotite. Leucogranitic dykes associated with the Ronda peridotite would occur throughout the Betic Zone according to Torres-Roldán et al. (1986). The extrusion of large volumes of calc-alkaline volcanic rocks of the Cabo de Gata Volcanic Chain ceased at the end of the Tortonian. Shortening and the uplift of the Tabernas Basin, inferred from folded Serravallian–Tortonian sedimentary rocks discordantly covered by Messinian Azagador conglomerate, also ceased at the end of the Tortonian.

The detailed uplift history of the land masses in the Arc of Gibraltar has been clarified by investigations of one of the nappes in southeastern Spain – the Aguilón fold nappe. This particular nappe, emplaced at 10–15 km depth as inferred from marbles in its mylonitic footwall formed at 300 °C (Behrmann, 1983), surfaced in southeastern Spain (Sierra Alhamilla) by the end of the Tortonian 7 Ma

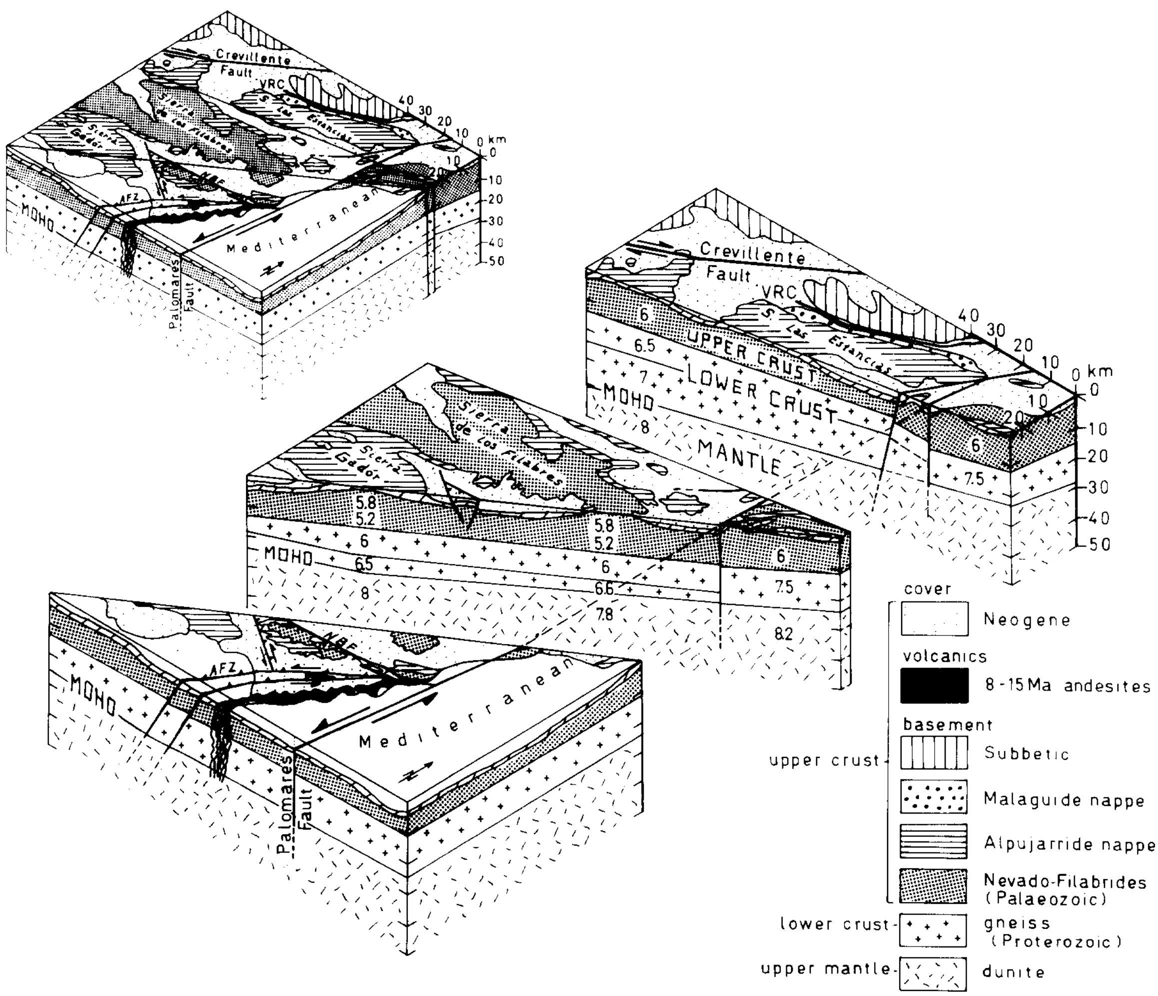


Fig. 34. Isometric block diagram of the crust and upper mantle beneath the eastern Betic Zone. The surface geology is observed in field surveys and the subsurface structure is inferred using seismic velocity profiles of Banda and Ansorge (1980, indicated in km s^{-1}) along the lines separating the blocks. The NW dip of the Almería Fault Zone (*AFZ*) is assumed on the basis of surface observations of Westra (1969). The Vélez Rubio Corridor (*VRC*) is a major dextral strike–slip fault mapped by Hermes (1978, 1984) (from Weijermars, 1987a, fig. 8).

B.P. due to continuous uplift since its Aquitanian emplacement 22 Ma B.P. (Weijermars, 1985b). This corresponds to a time-averaged uplift rate of 0.5–0.7 mm a^{-1} (Weijermars et al., 1985).

The trend of the basin and range which defines the modern grain of the Betic Cordilleras is due to the Neogene uplift and associated refolding of the nappes (Weijermars et al., 1985). Late Tortonian folds in the basement of the Sierra Alhamilla have an amplitude of 1 km and wave-length of about 16 km. The isostatic uplift estimates and the Neogene sedimentary record suggests that the Aguilón nappe was near the surface when it was refolded by brittle buckling (less than 1 km deep) some 7 Ma B.P. Consequently, it seems that lateral crustal shortening was superposed upon the isostatic uplift during the Late Tortonian. This crustal shortening might well be explained by a wedging effect, caused by transport of the Iberian microplate towards Italy along two converging strike–slip faults (Fig. 28). The seismic faulting of SE Spain since at least 8 Ma B.P. obviously also has its origin in such plate reorganisation forces.

8.3. Crustal structure

The geology and crustal structure of SE Spain is summarized in the isometric block diagram of Fig. 34. The crustal subdivision is based on seismic velocities inferred from Banda and Ansorge (1980). These authors contend that their data indicate the upper crust to comprise mainly Nevado–Filabride rocks. This seems tenable and particularly that rocks of the Nevado-Lubrín unit may be autochthonous and equivalent to Palaeozoic rocks exposed in the Hercynian belts of Portugal and Morocco. The lower crust may be composed of gneiss, as xenoliths of migmatitic gneiss are entrained in andesitic volcanic rocks at El Hoyazo (Zeck, 1970), Mazarron and Cartagena. Tentative Rb–Sr whole-rock ages of 800 Ma B.P. for the Hoyazo xenoliths (Zeck, 1970) imply that Proterozoic basement probably occurs beneath the Palaeozoic rocks of the Nevado–Filabride complex (Fig. 34). The seismic sections of Banda and Ansorge (1980) have been used to suggest that the crust east of the Alhama de Murcía-Palomares-Almería Faults is entirely different from that to the west (De Larouzière et al., 1988). This was taken even further by LeBlanc (1990) who stated that these faults form the boundary between the African and Eurasian plates (LeBlanc, 1990). These interpretations seem to conflict with the presence of all the Betic units at either side of the Palomares Fault (Fig. 17).

8.4. Messinian Salinity Crisis

The Messinian Salinity Crisis (MSC) was a catastrophic event which led to the deposition of 10^6 km^3 of evaporite in the Mediterranean between 7 and 5 Ma B.P. (Ryan et al., 1973; Hsü et al., 1977; Adams et al., 1977). The Neogene deposits of southeastern Spain comprise in part a late Messinian evaporite sequence which is one of the best preserved continental remnants of the MSC. Neogene tectonics suggests a scenario explaining how the Mediterranean could desiccate 7–5 Ma B.P. (Weijermars, 1988a).

Subsidence in the Alboran area maintained a major seaway between the Mediterranean and Atlantic 15–7 Ma B.P. (Fig. 35a). The major Miocene seaway probably lay through the Guadalquivir trough, because the Late Miocene mammalian assemblage of southern Spain seems of African rather than European affinity (Van Couvering et al., 1977). However, this argument must be regarded with scepticism in view of correlations by De Bruijn et al. (1975). Whatever its exact location, there was a connection between the Mediterranean and Atlantic, the so-called "Betic Straits", which became progressively narrower near the end of the Tortonian 7 Ma B.P. (Weijermars, 1988a). At least the Guadalquivir seaway was narrowed and finally closed by folding and thrusting of the Subbetic over the Prebetic about 7 Ma B.P.

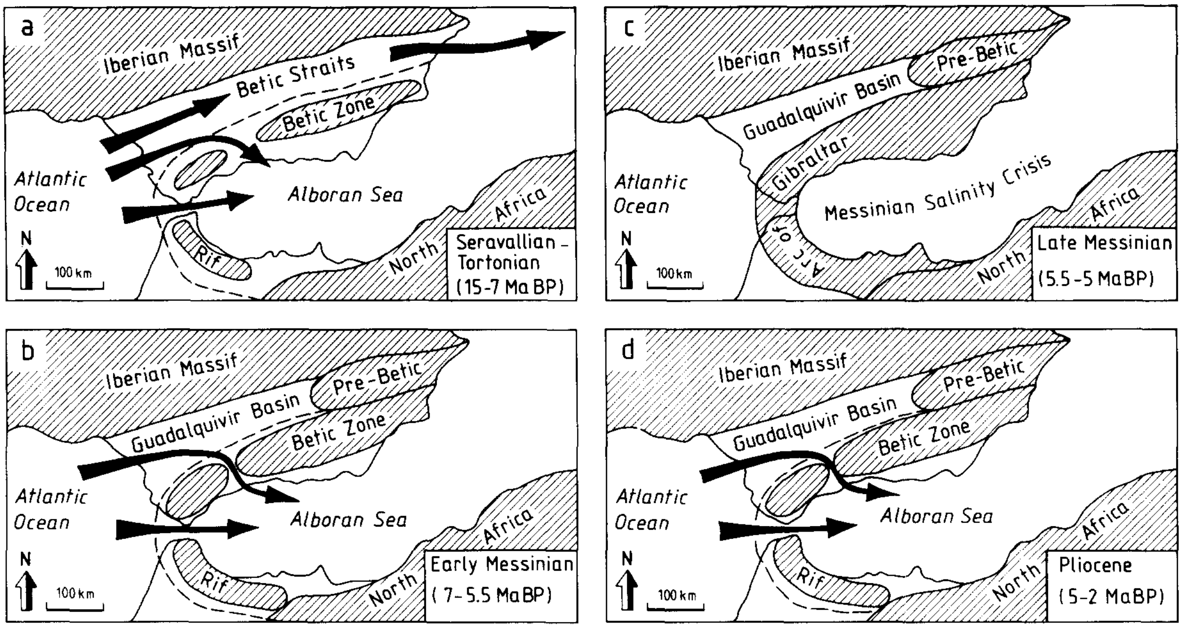


Fig. 35. Palaeogeographic maps illustrating four major stages in the evolution of the Messinian Salinity Crisis: (a) the main seaway, the Betic Straits, lay through the Guadalquivir Basin and some deep sills transecting landmasses in the immature Arc of Gibraltar between 15 and 7 Ma B.P.; (b) progressive uplift of the Arc of Gibraltar and the associated folding in the Prebetic closed the Betic Straits and only a few shallow sills in the Arc of Gibraltar maintained the connection between the Atlantic and the Mediterranean between 7 and 5.5 Ma B.P. This caused the onset of the Messinian Salinity Crisis; (c) continued uplift in the Arc of Gibraltar, possibly in combination with glacio-eustatic sea level reduction, disconnected the Mediterranean from the Atlantic and caused the peak of the Messinian Salinity Crisis between 5.5 and 5.0 Ma B.P.; (d) local graben tectonics and erosion, possibly in combination with a world-wide sea level rise due to the end of the Kaptian–Optian glaciation, (re-)opened the Strait of Gibraltar about 5 Ma B.P. The distribution of Pliocene deposits in southern Spain suggests that the modern coastline was not established until the onset of the Quaternary 2 Ma B.P. (from Weijermars, 1988b, fig. 3).

The Atlantic and Mediterranean were thus only connected by shallow sills in the Arc of Gibraltar during the Early Messinian 7–5.5 Ma B.P. (Fig. 35b). This encouraged the formation of dense brines in the deeper parts of the Mediterranean and may have caused deposition of the main body of the Messinian evaporite (cf. Weijermars, 1988a). A combination of the eustatic lowering (up to 200 m) of sea level due to glaciation and progressive uplift of the tectonic Arc of Gibraltar may finally have disconnected the Mediterranean from the Atlantic during the Late Messinian 5.5–5 Ma B.P. (Fig. 35c). This led to desiccation of the entire Mediterranean and periodic influx of marine water leading to cyclic evaporite sequences of laminated gypsum, anhydrite, dolomite and halite. Examples of such rocks are exposed in the Sorbas Basin, discussed in section 5. The so-called Lago Mare deposits, fresh to brackish water sedimentary rocks, followed upon the dessication period and suggest considerable influx of continental water. This would explain the continental facies of the top of the Caños Formation in the Basins of Sorbas, Vera, and Agua Amarga (see section 5). Formation of the Pliocene Straits of Gibraltar by graben formation in connection with strike–slip faulting, flooded the Mediterranean and terminated the Messinian Salinity Crisis 5 Ma B.P., introducing marine Pliocene deposits in southeast Spain (Fig. 35d).

8.5. Hydrocarbons

The accumulation of Neogene sedimentary rocks in the subsiding Alboran Basin may have been favourable to the formation of hydrocarbons. Seismic reflection profiles reveal that diapiric structures are abundant in the deeper parts of the Alboran Basin (Fig. 30). However, there is some uncertainty concerning the nature and age of the diapiric source layers. Auzende et al. (1975) interpreted the diapirs as salt stocks derived from Messinian source layers, whereas Mulder and Parry (1977) could not decide whether the diapirs consisted of either salt or clay, possibly of pre-Serravallian age.

Seismic reflection profiles also suggest that the crystalline basement under the floor of the Alboran Sea is a palaeorelief covered by a 4–6 km thick Neogene succession (Pastouret et al., 1975). The basement investigated at drill-site 121 of DSDP Leg 13 occurs at only 864 depth, capped by 194 m consolidated marls, 380 m Pliocene marls and turbidites and 290 m Quaternary pelagic marls (Nesteroff and Ryan, 1973; Olivet et al., 1973a,b; Ryan et al., 1973). The 194 m thick basal marl sequence, initially assigned to the Tortonian (Olivet et al., 1973a,b) was later reinterpreted as Messinian (Montenat et al., 1975; Auzende et al., 1975) and contains 30–40% nannoplankton and some foraminifers, e.g. Globigerinidae (Nesteroff and Ryan, 1973) potentially suitable for hydrocarbon genesis. Decomposition of organic material into hydrocarbons starts at temperatures in excess of 60 °C, and temperatures between 80° and 150 °C necessary for petroleum maturation are likely in the deeper parts of the Alboran Basin. Traces of subsurface oil allegedly contaminating water wells of Garrucha immediately after a minor earthquake in 1929 (Dupuy de Lôme, 1933), suggests that conditions favourable for hydrocarbon formation were also met in the Vera Basin, SE Spain.

Armstrong et al. (1980) have suggested that Messinian reefs similar to those exposed in southeastern Spain would form good reservoir rock if present in off-shore locations. Preliminary surveys in the Alboran Basin thus seem to suggest that diapiric caprock, source, reservoir, and temperature history may have been favourable to hydrocarbon accumulation.

9. DISCUSSION

This review attempts to integrate most of the data collected in the Betic Zone of southeastern Spain. The data discussed range from metamorphic basement inliers, Neogene sedimentary rocks, calc-alkaline volcanic rocks to Neogene fault patterns. Admittedly, it is a challenging and time-consuming task to attempt covering such a broad range in one review. One justification for discussing all these data together is to emphasize their interrelationship. For example, the distinction between Older and Younger Neogene sedimentary rocks is defined by the sudden change in detritus from Alpujarride to that derived from Nevado–Filabride basement (Völk and Rondeel, 1964). Consequently, there is the important palaeogeographic and geodynamic implication that the Nevado–Filabride core of the regional anticline in the Sierra de los Filabres became exposed and was already elevated at the onset of the Younger Neogene (Serravallian times). Conversely, the presence of the Huebro graben, filled with Serravallian–Tortonian sands and marls, at 900 m height on the anticlinal axis of the Sierra Alhamilla basement suggests that the area south of the Sierra de los Filabres did not elevate into topographic relief until the Late Tortonian.

Detailed knowledge of the Neogene sedimentary rocks in turn is important for assessing the reliability of radiometric datings of the various Neogene volcanic complexes in the area. Tortonian intercalations and onlappings provide marker horizons for the El Fraile and Rodalquilar calderas, respectively (see section 6). K–Ar ages of 10.9 ± 1 Ma suggested for lamproites in the Vera Basin (Bellon, 1976; Bellon and Brouse, 1977), can be invalidated by their biostratigraphic intercalation between the Late Messinian Turre Formation and Early Pliocene Cuevas Formation. The age for these lamproites implied by the biozonation is about 5 Ma (see section 6).

Understanding the large strike–slip faults in southeastern Spain requires comprehensive knowledge of all the tectonic elements affected and displaced by the faults. For example, the minimum displacement on the Palomares Fault s.s. can be estimated at 25–30 km by assuming that the Volcanic Chain of Cabo de Gata was continuous with the volcanic arc extending from Vera to Cartagena before fault movement (Fig. 25). A minimum age for the fault of about 7–8 Ma is inferred from the radiometric dating of the displaced volcanic rocks. Cataclastic slivers of andalusite- and sillimanite-bearing micaschist of the Nevado–Filabride basement are also trapped by the Almería Fault Zone. This indicates that the Nevado–Filabride basement was already near the surface at the onset of the strike–slip movement on the Almería Fault (i.e. 7 Ma B.P., Westra's second stage, see section 7). Pliocene sedimentation patterns and displaced Quaternary river beds suggest that the fault is still active.

The geology and structure of the Betic Cordilleras have been studied in considerable detail so far, but fundamental uncertainties still arise in tectonic reconstructions and speculations on the geodynamic mechanisms involved. Table 2 is a first attempt to compile both the chronology and geodynamic history of the geology in SE Spain. The complexity of events implies that this scheme may need adjustment if new and better data become available. It would be challenging to design a similar scheme using data from the Rif orogen, and test the correlation of events with those inferred for the Betic Cordilleras. Montenat et al. (1988) recently attempted a geodynamic interpretation of the Neogene Basins in the eastern Betic Zone including palaeostress estimates. It would be useful to continue studies of the regional deviatoric stress patterns, both for the present day and remote past.

A new dimension has recently been added to the importance of current doubts about the exact location of the present-day boundary between the European and African plates in the western Mediterranean. This is because the Straits of Gibraltar are now under study as a possible site for a permanent traffic link between Africa and Europe. Spain and Morocco have been elaborating pilote studies for a permanent link after an agreement signed in 1980. The preliminary results have recently been presented in Paris (Serrano, 1989), at a symposium otherwise mainly dedicated to evaluation of the progress of the Channel tunnel. That the construction of expensive tunnels under the sea floor is not necessarily jeopardized in areas of high seismic activity is evidenced by the 54 km long Seikan tunnel connecting the Japanese Islands of Honshu and Hokkaido since 1988. However, detailed understanding of plate sutures and fault patterns in the western Mediterranean is essential for assessing the feasibility of a permanent link between Africa and Europe. (Hannibal would have loved the idea when he crossed the Straits of Gibraltar with his elephants some 2000 years ago.)

It would also be useful to improve the dating of rocks suggestive of an ancient Betic Subduction suture. This could involve advanced study and indexing of the various ophiolitic mafic and ultramafic rocks metamorphosed in glaucophane and eclogite facies. Existing sections of Nijhuis' (1969) Muñoz amphibolites in the Umbría nappe of the Sierra de los Filabres ought to be reexamined in connection with further sampling of similar amphibolites in the Arto and Castro units of the Sierras Cabrera (Westra, 1969) and Alhamilla (Platt and Behrmann, 1986). Additionally, eclogite facies rocks, a.o. eclogitised skarns exposed near the Bédar granite, need to be studied in more detail.

The origin of the Neogene magmatism in the area is still poorly understood, although an important compilation of field data and geochemical analyses has recently been published by Bordet (1985). This suggests advanced studies involving various techniques aimed to clarify the geodynamics and mechanism of magma generation leading to emplacement of leucogranites, rhyodacites and lamprophyres. Another challenging field aspect is that volcanic ash layers – likely to have emanated abundantly during the main phase of volcanic activity 15–7 Ma B.P. – so far have not been recognized in the Tabernas Basin, where the bulk of Serravallian–Tortonian sedimentary rocks is exposed. Ignimbrites reportedly have very widespread air-fall ashes which may well preserve in subaqeous sedimentary rocks and provide excellent time markers (Cas and Wright, 1987).

Finally, the ancient Greek referred to the white cliffs of Ceuta and Gibraltar at either side of the Straits as the pillars of Heracles. These marked the connection of their familiar Mediterranean realm with "Okeanos", the world's feared and infinite water envelope. Heracles had to navigate through the Straits leaving his domains to defeat Geryon and Cerberus at the mouth of the Guadalquivir in southwestern Spain – and he won the golden apple of the Hesperidae. Whether true or not, comparison of this ancient myth with modern accounts suggest that we have come a great deal closer to unravelling the secrets of Heracles' pillars. Nonetheless, it remains important to continue coordinated efforts in unravelling the various geological problems remaining in this area.

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