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The Betic Cordilleras (SE Spain). Anatomy of a dualistic collision-type orogenic belt

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Abstract

The first of three main tectonic events in the orogenic evolution of the Betic Cordilleras of southern Spain involved crustal subduction during the Late Cretaceous. It included the stacking of nappes in a deep crustal environment, accompanied by HP-metamorphism and polyphase ductile deformation. It has only been recorded in the nappes of the Internal Zone of the Betic Cordilleras and took place after a Middle Jurassic initial phase of rifting and break-up of a Triassic and Early Jurassic carbonate platform. The second phase in the development of the orogenic belt starts with an important regional phase of extension in Late Oligocene-earliest Miocene time. Crustal thinning during this extensional phase and updoming of the subcrustal lithosphere in the Betic-Alboran domain resulted in heating of the extended crust. Heating has been recorded in the metamorphic nappes of the Internal Zone. Extension of the Betic-Alboran domain resulted in low-angle normal faulting in the nappe pile of the Internal Zone. In the Early Miocene an abrupt transition from regional tension to compression is responsible for the final thrusting of elements of the Betic nappe complex towards the passive continental margin of the Iberian plate. This 'final emplacement' of the nappes marks the beginning of Neogene thin-skinned deformation in the External Zone. In the Internal Zone, continuing convergence between the African and Iberian plates, resulted in strike-slip deformation from the latest Burdigalian onwards. Deformation in the eastern Internal Zone during this third tectonic phase is mainly characterised by basin subsidence and basement uplift in a strike-slip controlled regime under changing orientations of the main compressive stress. Theoretical lithospheric strength profiles predict differences in lithospheric strength between the eastern and western Betic Cordilleras, caused by differences in thermal structure and crustal thickness of the lithosphere. These differences are an inherited effect of the Late Oligocene-earliest Miocene extensional phase that influenced in particular the eastern part of the Betic Cordilleras. In the western Betic Cordilleras tectonic modelling predicts bending of the lithosphere and development of the Guadalquivir foreland basin under the load of the nappes, emplaced during the Early Miocene. In the eastern Betics lithospheric strength was restricted to the brittle upper crust, resulting in brittle strike-slip deformation and the development of pull-apart basins and basement uplifts.

Introduction

Processes involved in orogeny and the development of mountain belts have always been a main theme in earth science research. In the past, mountain building has been explained by diverse and contrasting theories. During the late 19th century, the concept that mountain chains result from horizontal crustal shortening was generally accepted, but the causes of horizontal compression in the earth's crust remained a matter of considerable debate. Static earth models, based on the contraction hypothesis and the geosynclinal theory, culminated in Stille's concept of the orogenic cycle, that remained the leading theory for mountain building until the early sixties of this century.

Subsequent important contributions to the understanding of the evolution of orogenic belts were achieved by detailed structural analysis and by the study of the timing of metamorphic mineral growth in relation to deformation phases (Zwart 1962). This approach culminated in the publication of 'The duality of orogenic belts' (Zwart 1967), which suggested that crustal deformational styles in orogenic zones were related to specific types of regional metamorphism. Zwart proposed a classification of orogenic belts on the basis of their deformational and metamorphic end products and not on the specific characteristics of the depositional environment. He stated that the depths at which the rocks have been metamorphosed and deformed can be calculated, since the pressures under which specific mineral parageneses form and which directly depend on the weight of the overlying rocks, are relatively well known.

In the same period however the dualistic character of orogenic belts was illustrated in a well-documented case of plurifacial metamorphism in the Betic Cordilleras (Nijhuis 1964, De Roever & Nijhuis 1963), indicating that a specific crustal segment may evolve through different metamorphic facies series during its orogenic evolution.

From 1970 onwards, the static earth models were quickly replaced by the dynamic concept of plate tectonics. Plate tectonic theory predicts that mountain building takes place at converging plate boundaries in either compressional/non-collisional (Andean type) or collisional (Himalayan type) settings (Burchfiel 1980). Continent-continent collision involves the subduction or underthrusting of continental crust during early stages of the collision event. In the collision zone the continental crust is telescoped and thickened at depth into a series of nappes. After initial collision has taken place, continuing convergence between the interacting lithosphere plates, results in overprinting of the collision zone by several phases of ductile deformation. Recent papers by Platt (1986) and Dewey (1988) suggest that underthrusting and thickening of the continental crust in the collision zone is followed by uplift and gravitational instability of the previously thickened crustal domain, which is then thinned by gravitational spreading along low-angle normal faults during extensional collapse of the orogenic belt. The final stages of orogeny take place under decreasing temperatures and confining pressures as indicated by an increasingly brittle behaviour of the continental crust during deformation. Within several orogenic belts continuing convergence of the interacting lithospheric plates is accommodated along major faults, often with a large component of strike-slip movement. During these final stages the more external parts of the mountain belt become deformed too, as the deformation front spreads out towards the foreland.

Geological outline

This paper analyses the Betic Cordilleras of southern Spain as an example of a collision orogenic belt and discusses the main tectonic phases that evolved during the process of orogeny.

The Betic Cordilleras are situated within the zone of interaction between the African and European-Iberian plates in the westernmost part of the Alpine orogenic belt of southern Europe.

The mountain belt does not form a continuous range of high mountains, like the Alps, but is a discontinuous chain of more or less isolated sierras, with altitudes up to more than 3000 m in the Sierra Nevada, separated by Miocene-Quaternary basins with mainly flat-lying or slightly tilted, but locally strongly deformed sediments. The outcrop pattern of the orogenic belt continues into the north-African Rif and Kabylian Ranges, thus forming an arcuate loop that surrounds the Late Oligocene-Miocene extensional basin of the Alboran Sea (Fig. 1).

The Betic Cordilleras are subdivided into two main domains. The External Zone in the north consists of non-metamorphic rocks that were deposited on the Late Jurassic-Paleogene passive continental margin of Iberia in roughly parallel NE-SW trending facies belts (Garcia Hernandez et al. 1980, Geel 1991, De Ruig et al. 1991). The northernmost facies belt (External Prebetic), represents the extensive shallow marine carbonate platform of the Iberian shelf, characterised by lagoonal limestones and dolomites. More to the south, the slope of the Iberian margin shows a rather complete and continuous stratigraphy in open marine platform facies (Internal Prebetic). Still further south the Subbetic Zone is characterised mainly by pelagic deep water basinal sediments in a facies deepening to the south-east (De Ruig et al. 1991). Subsidence patterns and sudden lateral facies changes demonstrate differential subsidence of tectonic blocks during the Late Jurassic-Paleogene history, indicating that the margin was repeatedly influenced by extensional tectonics (Kenter et al. 1990). From the earliest Miocene onwards, the External Zone became involved in the collisional process and the former passive margin evolved in a thin-skinned foreland fold-thrust belt (Kenter et al. 1990, Geel 1991, De Ruig 1992).

The Internal Zone in the south is formed by a stack of metamorphic nappes (Egeler & Simon 1969), representing the deep structure of an unroofed Late Cretaceous subduction complex, where elements of the African plate have been thrusted underneath the over-

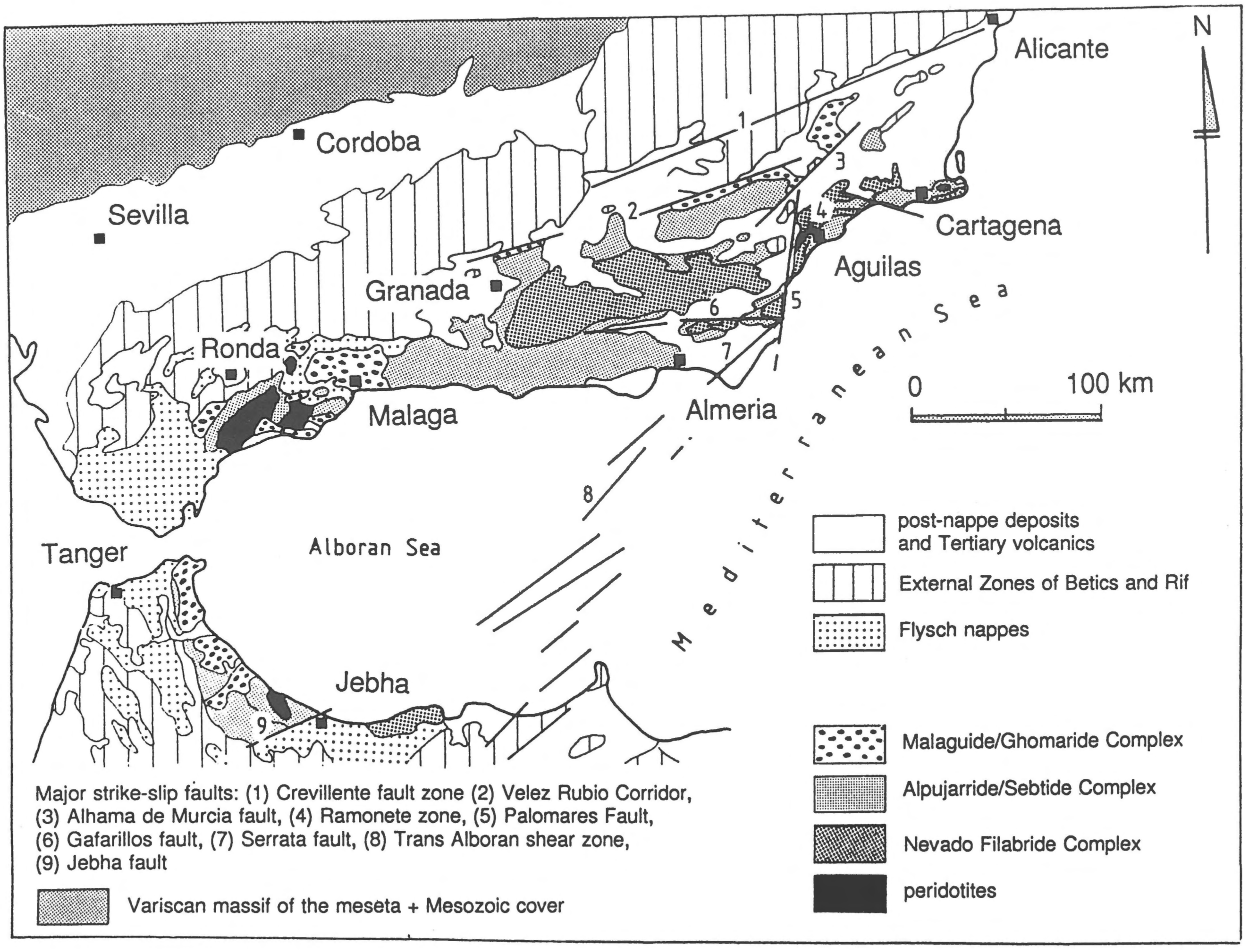

Fig. 1. Tectonic sketch map of the Betic Cordilleras and Rif Mountains (after Mäkel 1985).

lying Iberian plate (Bakker et al. 1989; De Jong 1991). The deepest structural elements are the metamorphic thrust sheets of the Veleta and Mulhacen Nappe Complexes, made up of Triassic and older metamorphic rocks (Puga & Diaz de Federico 1978). Together they form the Nevado-Filabride Complex (Egeler 1964), named after the high mountainous area of the Sierra Nevada and Sierra de los Filabres (Fig. 1). The Nevado-Filabride thrust slices are overlain by a large number of relatively thin metamorphic thrust sheets of the classical Alpujarride nappes, comprising deformed and metamorphosed Middle to Late Triassic platform carbonates with underlying Triassic and Palaeozoic clastics (Kozur et al. 1985, Simon 1987). High-pressure metamorphism has only been recorded in the Nevado-Filabride (Nijhuis 1964, Bakker et al. 1989) and Alpujarride (Goffé et al. 1989) nappes of the Internal Zone of the Betic Cordilleras.

The top of the thrust stack is formed by non-metamorphic thrust slices of the Malaguide Complex composed of Palaeozoic to Paleogene strata (Mäkel 1985, Roep & Mac Gillavry 1962). In the eastern Betic Cordilleras (amongst others in the Sierra Almagro, Sierra de Carascoy and Sierra de Orihuela), the Alpujarride and Malaguide nappes are situated directly upon very low-grade metamorphic rocks of the Almagride Complex, which represents the southern continuation of the Subbetic Zone in a position underneath the higher Betic nappes (Besems & Simon 1982).

The first main tectonic event in the Betic Cordilleras occurred during the Late Cretaceous-Paleogene when elements of the African plate subducted underneath, and collided with, Iberia. The subduction process resulted in the understacking of nappes in a deep crustal environment, accompanied by HP-metamorphism and polyphase ductile deformation. The subduction com-

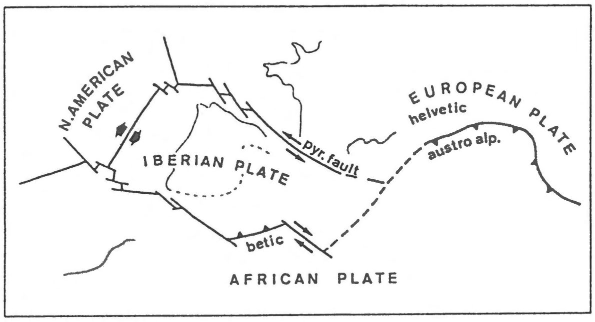

Fig. 2. Tentative reconstruction of Late Cretaceous boundaries between the interacting African, Iberian and European plates.

plex was subsequently uplifted by buoyancy, possibly assisted by active movement of the crustal segment over the footwall of the crustal subduction zone.

The second main phase in the tectonic development of the orogenic belt took place during the Late Oligocene-earliest Miocene and started with an important regional phase of extension.

During an abrupt change to subsequent compression, parts of the previously formed collision zone were thrusted upon the Iberian continental margin. This phase of overthrusting marks the onset of deformation in the External Zone that developed into a foreland fold-thrust belt characterised by cover deformation overlying a detachment in Triassic evaporites.

The third and last deformational episode is characterised by large-scale strike-slip deformation during continued convergence between the African and Iberian plates from the Late Burdigalian onwards. Strike-slip deformation has been observed in both the Internal and External Betic Cordilleras. Deformation during this third tectonic phase is dominated by basin subsidence and uplift of the pre-Neogene substratum and took place during reorientation of the main compressive stress.

Cretaceous subduction and collision

During the Late Jurassic-earliest Cretaceous, Iberia and Africa were separated by the Gibraltar-Azores Fracture Zone, a sinistral transtensional transform fault zone that connected the Jurassic Central Atlantic Basin with the opening Ligurian Ocean east of Iberia. Local oceanic pull-apart basins, characterised by newly formed oceanic crust, presently exposed in ophiolite complexes in North Africa, indicate the transtensional character of the transform fault zone. A similar strike-slip system connecting oceanic pull-apart basins, formed during Triassic and Early Jurassic rifting and transform faulting, was present at the northeastern side of the small Ligurian Ocean basin in the Piemont-Penninic domain (Lemoine & Trümpy 1987). Further east the system of thinned continental and locally oceanic crust continued in the Pienides and Transylvanide Carpathian area (Sandulescu 1988).

During the Late Cretaceous (115–80 Ma) oceanic spreading in the Atlantic Ocean propagated north (Mauffret et al. 1989, Malod & Mauffret 1990) and activated the North Pyrenean transform fault (Savostin et al. 1986, Srivastava et al. 1990, Malod & Mauffret 1990), that became the new plate boundary between Africa-Iberia and Eurasia (Fig. 2). In the deactivated transform fault between Africa and Iberia compression was initiated by a counter clock-wise rotation of $27° \pm 12°$ of Iberia, that resulted from spreading in the Bay of Biscay. As a result, the former transform fault zone developed into a subduction zone, where elements of the African plate, broken up along the previous strike-slip faults, subducted underneath the overlying plate of Iberian continental crust (De Jong 1991). This Late Cretaceous phase of subduction in the westernmost part of Alpine-Tethyan subduction system led to the development of the main Betic nappe units.

The early structural and metamorphic evolution has been unravelled by detailed field and laboratory analyses in the eastern Sierra de los Filabres (Bakker et al. 1989, De Jong 1991). The early history of the metamorphic nappes shows polyphase ductile deformation during HP-metamorphism characterised by an increase in temperature at constant pressure (Fig. 3).

In the Mulhacen Complex the main deformation structure is a D_2 transposition foliation (S_2). S_2 is parallel to the axial plane of centimetre to metre-scale isoclinal folds and is penetratively developed in all lithologies with the exception of some mafic relicts, which have either undeformed cores or contain a penetrative older S_1 foliation.

S_2-foliations further enclose intrafolial folds, boudins and augen structures. In mafic rocks boudins wrapped by S_2 contain S_1 glaucophane-epidote-mica foliations, intrafolial folds and tension gashes filled with quartz and aragonite. Evidence for pre-D_2 deformation in micaschists is usually restricted to local relicts of D_1 foliations within microlithons and the internal fabrics of pre- and syn-D_2 porphyroblasts.

Synkinematic D_1-metamorphism is characterised by the occurrence of glaucophane *s.l.*, garnet, epidote

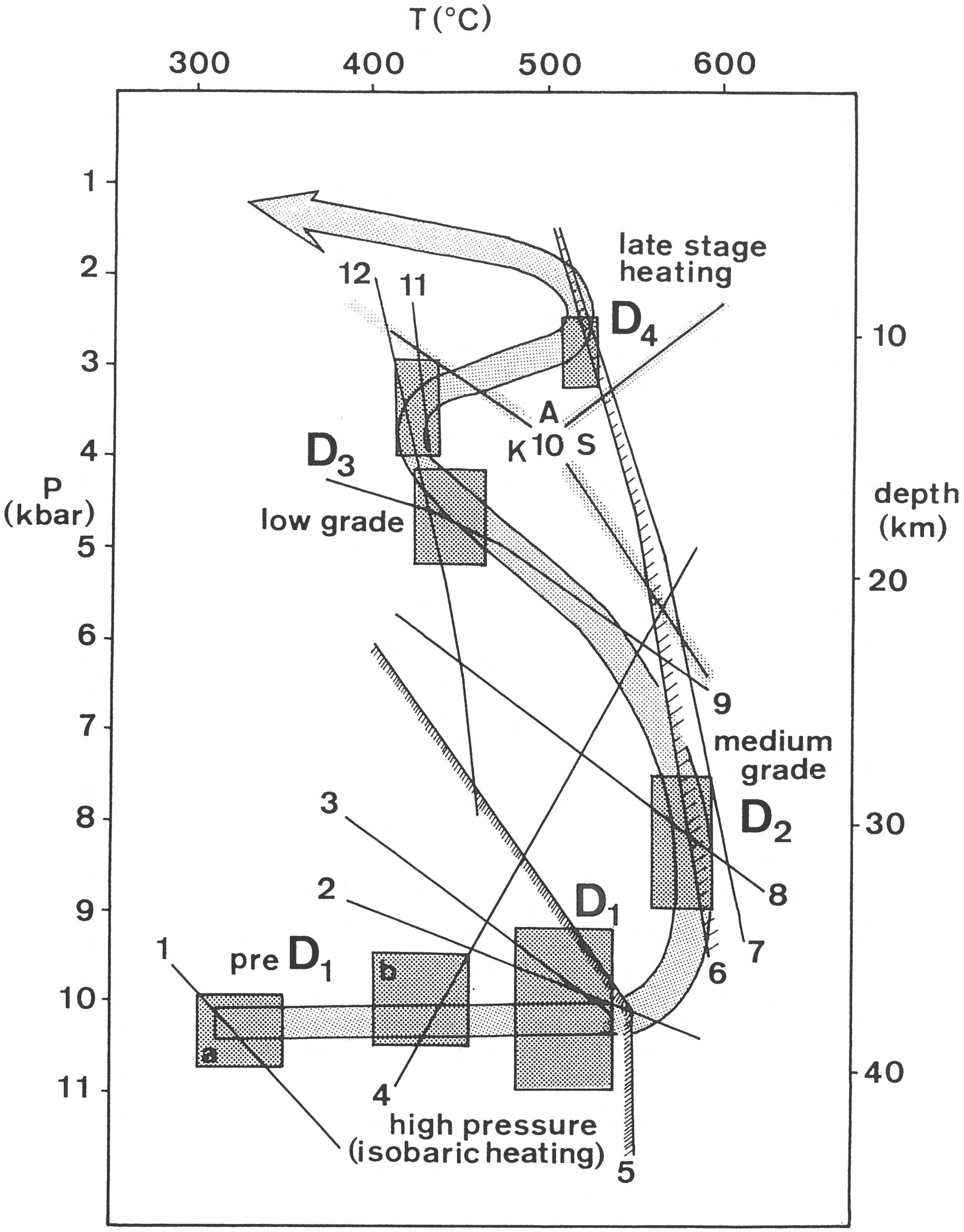

Fig. 3. Pressure-temperature-time diagram for the Mulhacèn nappe in the eastern Sierra de los Filabres. (1) albite = jadeite + quartz (Newton & Kennedy 1968); (2) rutile + almandine = ilmenite + kyanite + quartz (Bohlen et al. 1983); (3) aragonite-in (Boettcher & Wyllie 1967); (4) glaucophane + chloritoid = paragonite + chlorite (Kienast & Triboulet 1972); (5) glaucophane stability (Maresch 1977); (6) staurolite-in (Hoschek 1969); (7) antigorite = forsterite + talc + H_2O (Evans et al. 1976); (8) anorthite + H_2O = kyanite + zoisite + quartz (Newton & Kennedy 1963); (9) barroisite stability (Ernst 1979); (10) Al-silicate triple point (Holdaway 1971); (11) stilpnomelane + phengite = biotite + chlorite + quartz (Nitsch 1970); (12) pyrophyllite = Al-silicate + quartz + H_2O (Chatterjee et al. 1984). Boxes indicate the established P-T conditions for deformation phases D. (Modified after Bakker et al. 1989).

s.l., pure low-albite, paragonite and locally omphacite. Further high-pressure indicators are Mg-rich chloritoid, almandine-rutile intergrowth, aragonite stability and Si-values and b_0-values of phengites. Geothermobarometry indicates metamorphic conditions of 475–525°C at 9–11 kbar (Bakker et al. 1989).

The D_1-mineral parageneses postdate even older metamorphic assemblages (pre-D_1) characterised by incipient eclogitisation of dolerites and garnet-hedenbergite skarns, by crystallisation of quartz-jadeite, quartz-aragonite pairs, and by the formation of albite- and epidote-bearing omphacites at the contacts of mafic igneous rocks with carbonate layers (Helmers 1982). Metamorphic conditions for pre-D_1 parageneses indicate 350–400°C at 9.5–10.5 kbar (box a in Fig. 3) or, taking into account the Ca-contents of garnet (usually approximately Gr_{30}), a slightly higher temperature range (400–460°, box b).

During D_2, omphacite, glaucophane and crossite in mafic rocks recrystallise to blue-green amphibole, usually accompanied by albite growth. Locally blue-green amphibole or taramite replaces glaucophane. S_2 is further characterised by the occurrence of clinozoisite, zoisite and paragonite. Aragonite occurring in carbonate veins in mafic rocks has recrystallised to calcite. In micaschists the D_2 mineral assemblage consists of phengite, quartz, chloritoid, kyanite, staurolite, almandine and (clino-)zoisite. In gneisses taramite was locally formed from jadeite-acmite, concurrent with the recrystallisation of K-feldspar. High Si-phengite has remained stable. In calcitic and dolomitic marbles the parageneses tremolite + zoisite and colourless mica + quartz + Mg-rich chlorite have been formed. In ultramafic rocks chlorite + antigorite are stable. Neither forsterite nor pseudomorphs after this mineral have been observed. Geothermobarometry indicates that P-T conditions during D_2 are 8.25 kbar at approximately 570 °C.

Metamorphic conditions indicate that the rocks of the Mulhacen Complex have been subjected to a phase of rapid tectonic burial to a depth of approximately 37 km. This process is considered to represent a phase of crustal underthrusting of relatively cold, continental crustal material, which caused deformation of the original pattern of (sub)horizontal isotherms in the upper part of the lithosphere.

Thrusting of the main Betic nappe units is interpreted as a process of sequential delamination from the subducting plate during crustal understacking. It is assumed that initial stacking of the thrust slices of the Mulhacen Complex started during D_1. D_1 fabrics, although often strongly obliterated by severe D_2 deformation, appear to be present throughout the pile of thrust sheets. Furthermore there is evidence that initial thrusting predates D_2 as no D_2 strain or metamorphic gradient is present across the contacts between the thrust slices of the Mulhacen Complex (De Jong 1991).

D_1 stretching lineations associated with thrusting show WNW-NW orientation (De Jong 1990). The central part of the Macael-Chive gneiss body south of Lubrin for example contains a penetrative D_1 foliation with WNW-ESE trending stretching lineations. Asymmetric tails around feldspar porphyroclasts in these gneisses indicate WNW-ESE directed shear. Randomly oriented glaucophane sheaves in mafic bodies were progressively rotated into a pronounced WNW-ESE trending mineral lineation during formation of S_1 foliations.

After rapid tectonic burial the rocks were subjected to an increase in temperature at approximately constant pressure indicating that the rocks of the Mulhacen Complex were heated at constant depth. The isobaric heating pattern is interpreted to have been caused by cessation of the rapid crustal-scale underthrusting, enabling the start of the restoration of the thermal structure of the disturbed lithosphere (Bakker et al. 1989). Termination of the crustal-scale underthrusting is probably due to the fact that continental crust is only allowed to descend to a limited depth during subduction as the forces associated with underthrusting are balanced by the buoyancy of the underthrust segment.

Temperature increase lasted up to D_2. This prograde trajectory is characterised by the recrystallisation of several HP-minerals to intermediate pressure parageneses. D_2 has been associated with an upward movement of the affected crustal segment towards a higher level in the crust. The P-T-t path for the Mulhacen Complex indicates that D_2 took place at a depth of 28–34 km and that the rocks were uplifted some 7 km while the temperature had increased by approximately 70 °C. As this phase proceeded under peak temperature conditions it has resulted in the most penetrative and homogeneous deformation.

The uplift of high-pressure metamorphic rocks during D_2 towards a higher level in the crust may be explained by a combination of buoyancy and active transport of the rocks over the footwall of the previously formed crustal subduction zone. Structural analysis indicates dominant non-coaxial deformation and suggests an important component of lateral transport during D_2. Quartz c-axis fabrics indicate W to NW directed shear during D_2. It is therefore possible that D_1 and D_2 are related to a similar process and represent a continuum during progressive continental collision. In such a model D_2 is considered to represent continued imbrication of a crustal segment unable to descend any further because of buoyancy (Bakker et al. 1989, De Jong 1991).

Late Oligocene and Early Miocene extension and compression

During the Late Oligocene-earliest Miocene the orogenic evolution of the Betic Cordilleras is dominated by a major extensional tectonic phase. As the result anomalous thinning of continental crust occurred in the Betic-Alboran region, decreasing in thickness from 30–40 km in the northern to 24 km in the southern Internal Zone (Banda & Ansorge 1980, Banda 1988). The thickness of the continental crust further decreases towards the south to 17–20 km in the Alboran Sea (Fig. 7). Thinning of the continental crust in the Betic-Alboran region was compensated by updoming of upper mantle lithosphere, reflected by the pattern of positive Bouguer gravity anomalies centred in the Alboran domain.

Extension in the nappe pile of the Internal Zone was accommodated along low-angle normal faults (Platt & Vissers 1989). Thinning of the continental lithosphere caused an anomalous increase in temperature that is well documented in the metamorphic evolution of the Internal Zone (Fig. 3). Metamorphic parageneses and prograde mineral reactions in these thrust sheets indicate renewed heating, disturbing the overall retrograde metamorphic evolutionary trend related to uplift of the deep-seated nappe complex (Bakker et al. 1989, De Jong 1991).

In Early Miocene times an abrupt transition from regional tension to compression is associated with the final thrust emplacement of elements of the Betic nappe complex. Crustal shortening is evidenced by the superposition of higher grade rocks on top of lower grade rocks and by ductile thrusting of slices of ultramafic rocks. Kinematic indicators show NNW to NNE directed thrusting in the mylonite zone between the Nevado-Filabride and overlying Alpujarride Complex (De Jong 1991). The final emplacement of the nappes of the Internal Zone during the Early Miocene had an immediate effect on the structural evolution of the External Zone, that changed from a passive margin into a foreland fold and thrust belt. Sequential development of folds in the eastern Prebetic took place during the Late Aquitanian-Burdigalian (Beets & De Ruig 1992). Palaeostress data indicate that the direction of maximum compression was oriented N-S to NNW-SSE at the onset of folding. The location and orientation of major anticlines was partly controlled by ramps along pre-existing normal faults.

Several models have been presented to explain the associated extension and thrusting in the Betic Cordilleras. Model 1 (Fig. 4a) is the classical mantle diapir model (Van Bemmelen 1969, Torres Roldan 1979, Mäkel 1985, Weyermars 1985). According to the model the Alboran Sea is the site of a former mantle diapir causing uplift by the thermal anomaly in the upper mantle. Uplift results in nappe shedding of upper crustal segments, driven by gravity gliding or gravity spreading. This tectonic unroofing left an area with thinned continental crust in the centre of the orogenic area (Fig. 4a). Subsequent cooling of the diapir and isostatic adjustment leads to subsidence of the Alboran Sea.

Model 2 involves extensional collapse of previously thickened continental crust (Dewey 1988, Platt & Vissers 1989). According to this model, detachment or convective removal of the lithospheric root of the orogen causes rapid uplift of the collision zone. During uplift the crust then collapses in extension when the gravitational forces in the highly elevated orogen overcome the forces generated at the convergent plate boundary (Fig. 4b). Collapse of the orogenic belt thus leads to radial thrusting towards the periphery of the orogenic zone, and deep-seated metamorphic rocks of the former subduction complex (the previously thickened wedge) become exposed at the surface.

The model seems attractive but does not explain the driving process of lithospheric root detachment. One possibility is that removal of the lithospheric root was caused by detachment of part of the subducted lithosphere slab underneath the former collision zone (Fig. 4b). A problem that remains however, is that collapse of the orogenic wedge should end when the extended domain has sunk topographically lower than the zone of thrusting. Subsidence of the Alboran Basin started in the Late Aquitanian-Early Burdigalian (Jurado & Comas 1992), indicating that thrusting in the Betic Cordilleras was contemporaneous with marine sedimentation in the Alboran Basin.

Several authors explain the Late Oligocene-earliest Miocene Alboran Basin and other extensional basins in the western Mediterranean as back-arc basins on

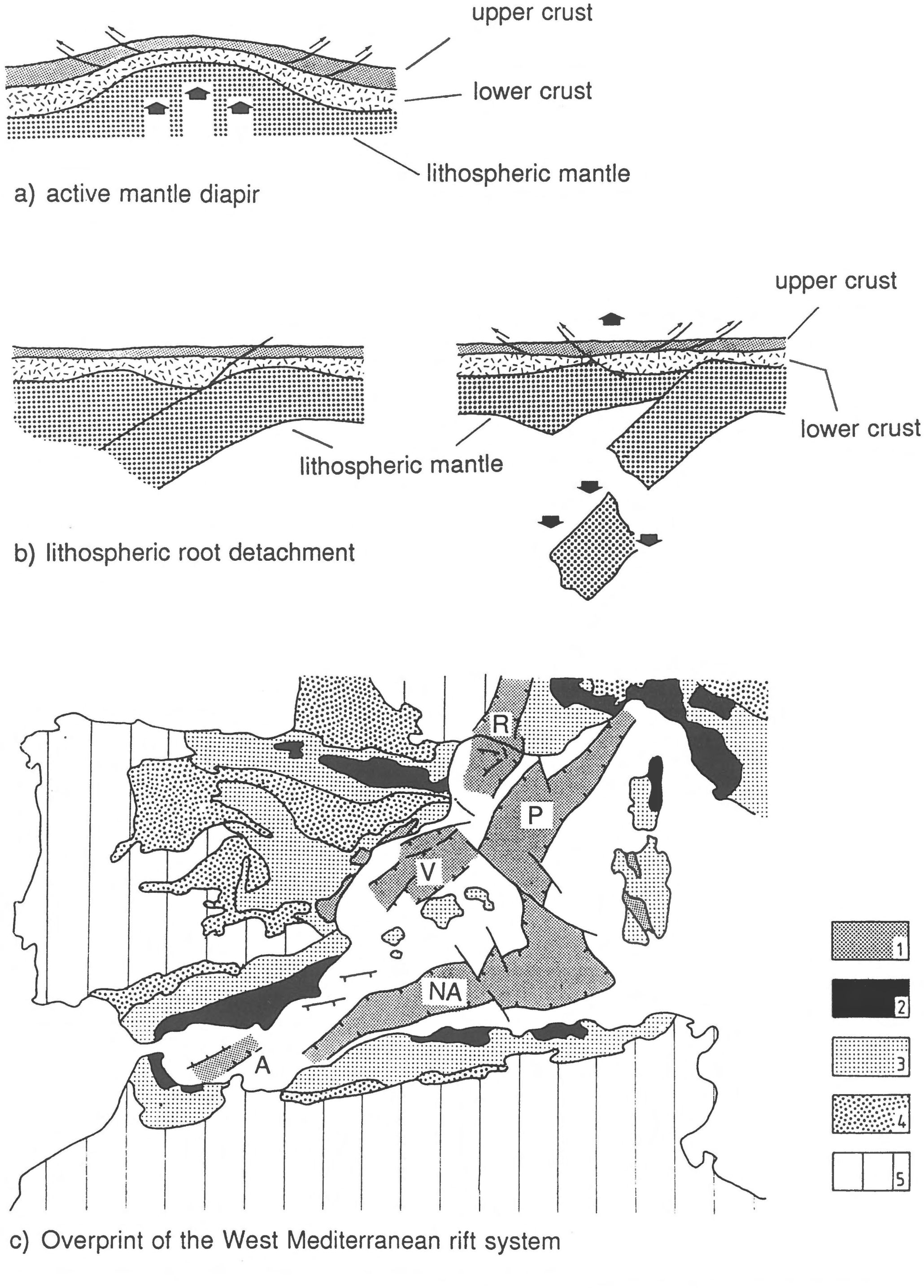

Fig. 4. Large-scale models for Late Oligocene-Early Miocene extension in the Betic Cordilleras. (a) Classical mantle diapir model: the actively uprising mantle diapir causes crustal extension, heating and uplift. Subsequent radial thrusting by gravitational spreading from the uplifted domain towards the periphery of the orogenic belt. (b) Lithospheric root detachment: detachment of the lithospheric slab causes rapid uplift and gravitational instability of the former collision zone. Crustal thinning by nappe shedding. (c) Regional overprint of the West Mediterranean rift system: overprint on the Betic Cordilleras causes crustal extension in the Betic-Alboran domain. Legend: (1) West Mediterranean rift system; (2) Internal Zones of Betics, Rif, Kabylian Ranges, Pyrenees, Apennines, Western Alps and Corsica, and oceanic remnants; (3) Deformed Mesozoic continental margins and inverted Mesozoic continental basins (Iberian chain); (4) Foreland basins; (5) European and African foreland. P = Provençal Basin; V = Valencia Trough; NA = North Algerian Basin; A = Alboran Basin; R = Rhône Graben. Modified after De Jong (1991) and De Ruig (1992).

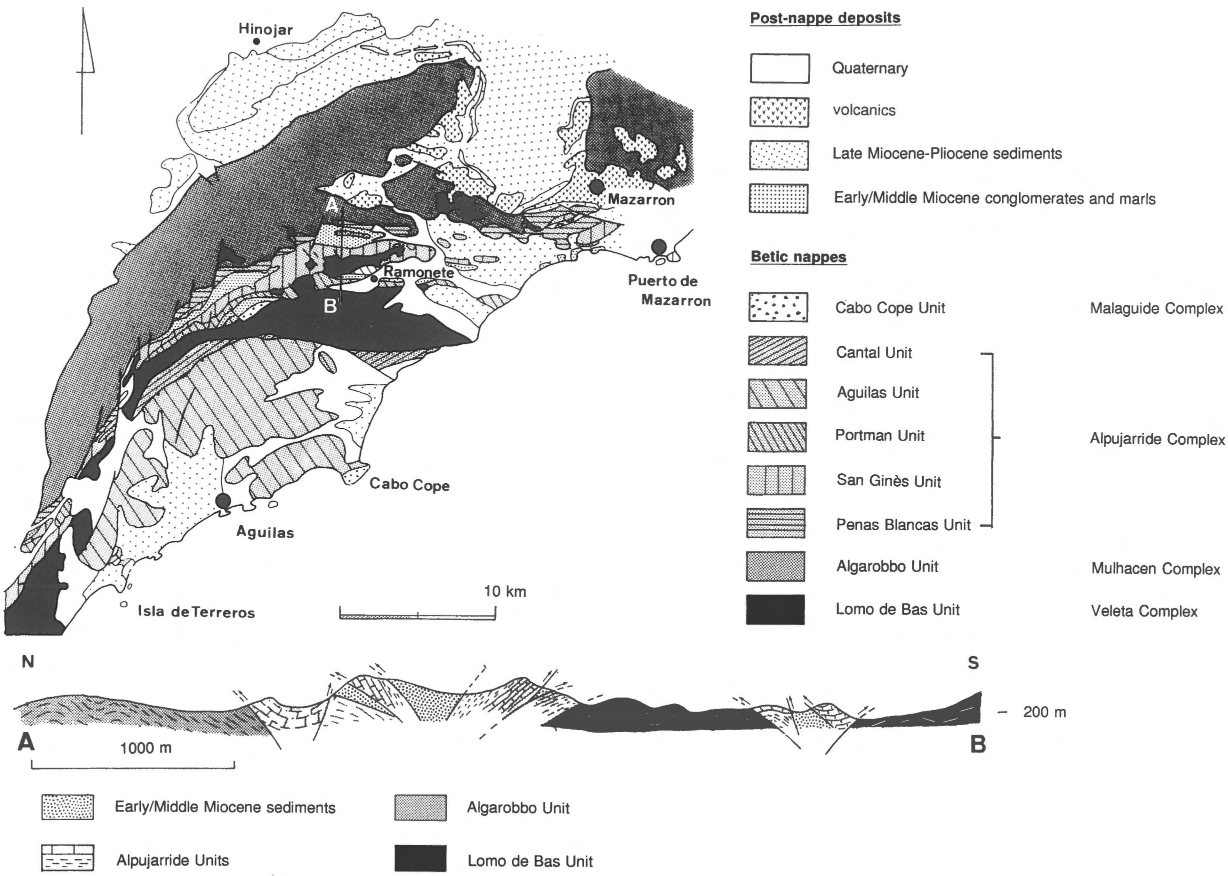

Fig. 5. Geological map of the Aguilas–Mazarron area. The Ramonete Zone is represented by the elongate structure of Middle Miocene basins and pre-Neogene substratum. Cross section AB illustrates the complicated structure of the Ramonete Zone.

the overlying plate above an active subduction system. They emphasise that the Mediterranean Alpine orogenic belts form arcuate loops that partially surround extensional basins. The basins are characterised by extension and thinning of the continental crust. Extension in the western Meditèrranean area in this period is demonstrated by rifting in the Valencia Trough and the Provençal Basin and by limited spreading and formation of new oceanic crust in the Ligurian, Sardino-Balearic, Alboran and North Algerian Basins (Fig. 4c).

There are however serious problems with such an interpretation. Firstly, there is no evidence that active subduction continued during the Miocene, when the Alboran Basin was formed. According to P-T-t data from the Internal Zone of the Betics, the nappe structure of the Upper Cretaceous subduction complex had already been formed, uplifted and re-equilibrated into lower greenschist facies conditions before extension started (Fig. 3). Secondly, one of the factors controlling extension in back-arc basins is the force exerted by the ocean-directed backward migration (roll-back) of the trench. Such a mechanism cannot operate however in the Betic-Alboran domain as Iberia and Africa were sealed and space for roll-back was not available.

The author suggests that extension in the Betic Cordilleras is neither fundamentally and genetically linked nor restricted to the specific collision zone in the Betic-Alboran domain, but is related to a much wider regional Cenozoic extensional system in western Europe (Ziegler 1988) that can be followed from the Rhône Graben southwards into the extensional basins in the western Mediterranean (Fig. 4c). Based on crustal thickness data (Fig. 7) this extensional system effected the eastern Betic Cordilleras in particular.

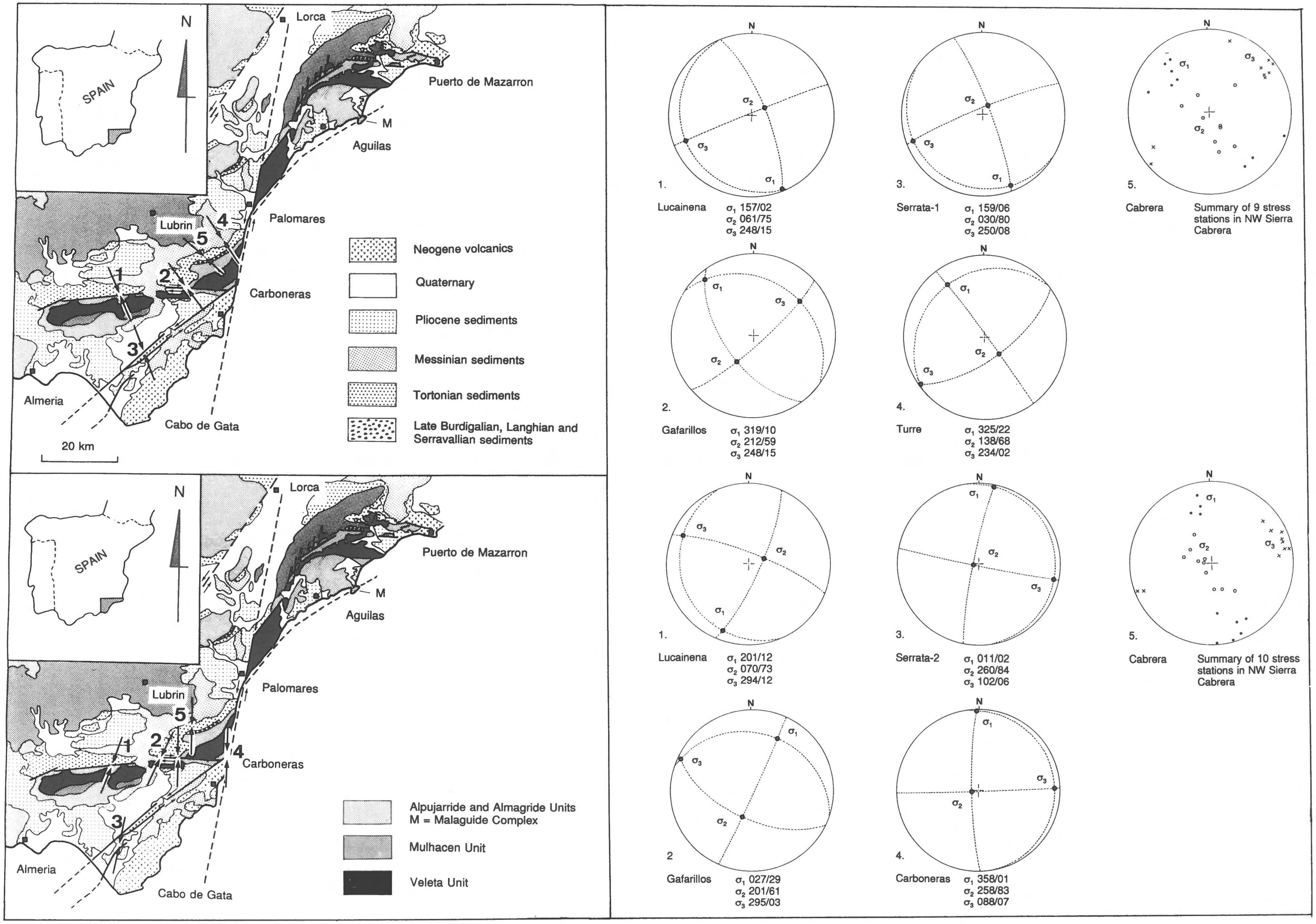

Fig. 6. Stress analyses in the south-eastern Betic Cordilleras. The upper map shows the location of stress measurements (1–5) and the orientation of the approximately NW-SE directed main principal stress (σ_1) during the Tortonian. The upper stereographic projection diagrams 1–5 (equal area, lower hemisphere) show the orientation of the calculated stress tensor for each stress station. The lower map and stereographic projection diagrams 1–5 illustrate the N-S directed Messinian-Pliocene stress field. Stress tensors represented in diagrams 5 are from Drenth et al. (1993).

Post-nappe strike-slip tectonics

Strike-slip deformation became an important mechanism in the Internal Zone during the Middle and Late Miocene. The major fault zones in SE Spain are indicated in Fig. 1. Detailed investigations have been carried out in recent years on the NE-SW-trending left-lateral Serrata Fault (Boorsma 1993), the Ramonete Fault Zone (Bon et al. 1989), and the right-lateral Gafarillos Fault (Miedema 1991, Drenth et al. 1993).

The tectonic scenery during the Neogene is one of rising sierras and rapidly subsiding intramontane basins. The Neogene sedimentary basins show structural and sedimentological evidence for a strike-slip pull-apart origin. Bon et al. (1989) have presented data on the age, depositional environment and deformation of Middle Miocene sediments within the ENE-WSW trending Ramonete Fault Zone, that transsects the nappe structure of the eastern Internal Betic Cordilleras in the Aguilas-Mazarron area. (Fig. 5). Transtensional deformation along the fault zone during the Middle Miocene caused the formation of small (kilometre-scale) elongated basins that were largely filled by sedimentary mass flows derived from nearby Malaguide and Alpujarride basement uplifts. The occurrence of high-energy deposits and the proximal provenance of the basin-fill sediments indicates the close association of areas of rapid subsidence and deposition with areas of local basement uplift and erosion, that is characteristic for a strike-slip environment (Christie-Blick & Biddle 1985).

In the earliest Tortonian the basins telescoped during transpressive deformation along the Ramonete Fault Zone (Bon et al. 1989). Structural evidence for strike-slip fault movement is presented by (sub)horizontal striations on steeply dipping fault planes and by small-scale overthrusting and imbrication of the Middle Miocene basin sediments with rocks of the underlying Betic thrust sheets. Thrusting took place in both a northern and southern direction. The thrusts have been interpreted as faults that branch off from a principal displacement zone at depth in a major flower structure (Bon et al. 1989).

The Neogene stratigraphical succession within the intramontane basins reveals that activity along major fault zones and the associated formation of sedimentary basins did not occur simultaneously throughout the Internal Betics. During the Late Neogene a new generation of basins developed in the eastern Internal Zone, that partly overlie the previously active Middle Miocene fault zones. The Ramonete Fault Zone for example forms part of a more extensive fault system that is partly covered by younger basins of Tortonian age. The fault zone disappears north of Mazarron underneath Tortonian sediments of the Mazarron Basin, but it reappears on the eastern side of this basin in the mountains south of Cartagena. In the south the Ramonete Zone is cut off by the NNE-SSW trending strike-slip faults of the Palomares Fault System which offset the Betic chains in a left-lateral sense for more than 20 km during the Middle Messinian.

Quantitative subsidence analyses were carried out on several relatively undeformed younger Neogene basins (Cloetingh et al. 1992). Synthetic subsidence curves have been constructed on the basis of a non-instantaneous stretching model (Cochran 1983), modified to incorporate effects of lateral heat flow and depth-dependent stretching. The curves were further based on a thickness of the continental crust of less than approximately 20 km, as inferred from seismic profiles (Banda 1988). The synthetic subsidence curves have subsequently been compared with back-stripped curves. The analysis indicates that the Late Neogene basins in the Internal Zone are either strike-slip pull-apart or extensional basins. Rapid initial subsidence is in accordance with strike-slip pull-apart generation. Breaks in the subsidence of the basins are often recorded indicating compressive or transpressive deformation and uplift phases.

Stress analysis

In the northern Sierra Cabrera and along the northern boundary of the Sierra Alhamilla, the Gafarillos Fault Zone forms a linear dextral strike-slip zone that continues towards the east into the Alpujarran Corridor between the Sierra Nevada and Sierra de Gador. The fault zone separates the Alpujarride and Nevado-Filabride nappes of the Sierra Cabrera and Sierra Alhamilla from the Tortonian sediments of the Sorbas and Tabernas Basins. Within the fault zone highly deformed Langhian-Serravallian and Early Tortonian sediments are intensely mixed with Alpujarride and Nevado-Filabride basement. Movement along the fault zone influences Tortonian sediments in the Sorbas and Tabernas Basins, which are folded in en-echelon folds with NE-oriented fold axes in agreement with dextral strike-slip movement along the master fault.

Activity along the major fault zones depends on fault orientation in relation to orientation of the stress

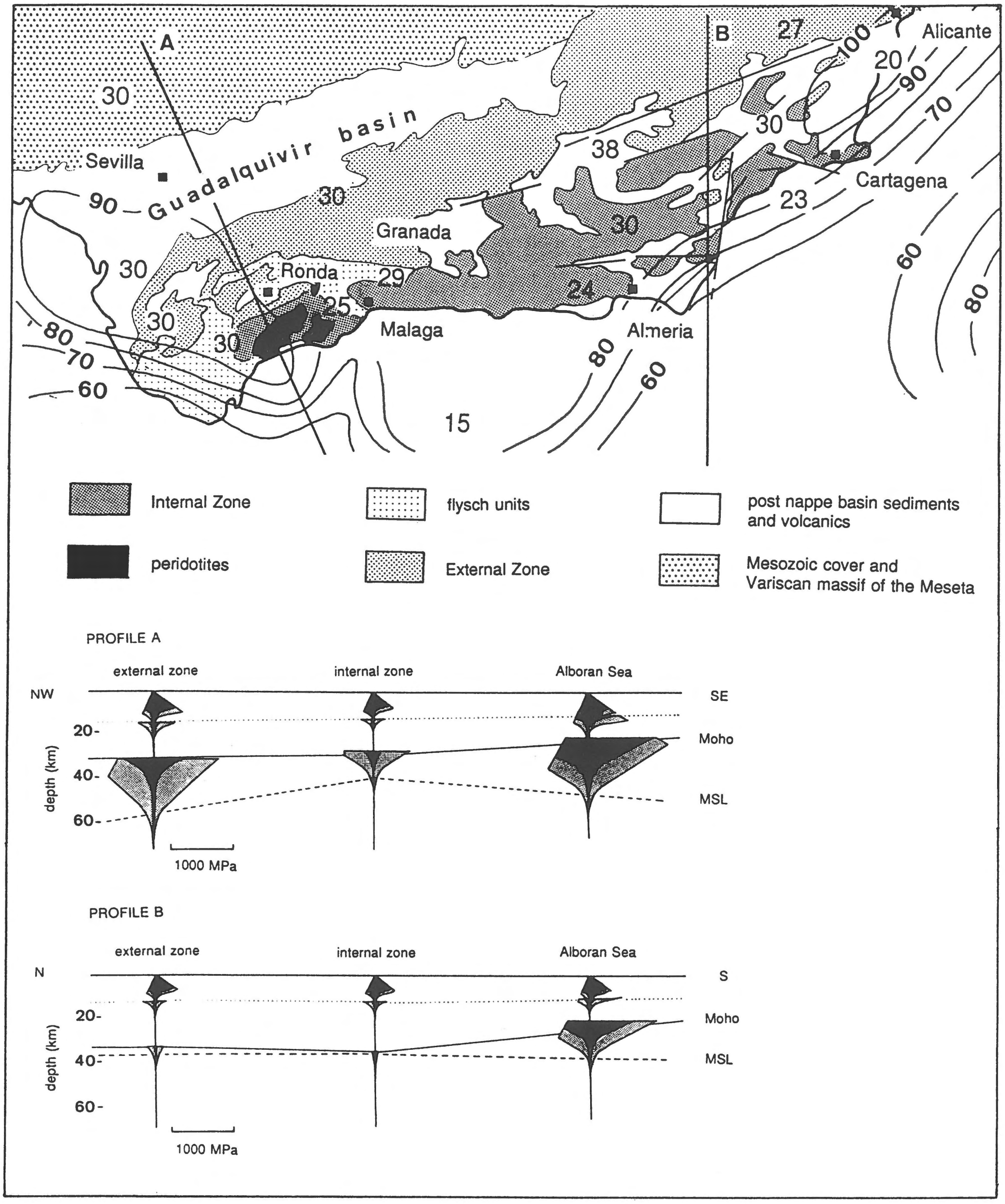

Fig. 7. Simplified structural map of the Betic Cordilleras: The contour pattern represents heat flow data from Albert-Bertrán (1979). Heat flow estimates range from 70–90 mWm^{-2} in the western Betics to 80–100 mWm^{-2} in the eastern Betics. Isolated numbers in the map indicate Moho-depth in km (after Banda 1988). Lithospheric strength profiles along profiles A and B: Layered lithospheric model with quartzitic composition of the upper crust, diabase composition of the lower crust and olivine composition of the subcrustal lithosphere. Grey and black shading indicate strength profiles corresponding to minimum and maximum heat flow estimates, respectively. Tensional strength is shown on the left-hand side of the diagrams, and compressional strength on the right-hand side. The lower boundary of the mechanically strong part of the lithosphere (MSL) corresponds to a ductile strength of 50 MPa in the subcrustal lithosphere. Modified after Van der Beek & Cloetingh (1992) and Cloetingh et al. (1992).

field. During field studies a large number of fault planes and fault striations have been measured in a limited number of stress stations. The data have subsequently been submitted to computer analysis, developed by Smid & Heijke (1987), based on the inversion method of Michael (1984). The program enables the calculation of the best fitting stress tensor for a population of striated fault planes with a minimum size of 6. The results indicate a clock-wise rotation of the stress field during the Neogene (Fig. 6).

Figure 6 (upper map and diagrams) summarises the orientation of the Tortonian stress tensor. Along the Gafarillos Fault Zone, dextral strike-slip along the principal displacement zone was controlled by a NW-SE directed stress field with a horizontal ($\sigma_1 - \sigma_3$) plane and vertical σ_2 (diagrams 1: Lucainena and 2: Gafarillos). A similar orientation of the stress tensor is obtained from measurements in Tortonian sediments in the Serrata Fault Zone (diagram 3: Serrata-1) and from 9 stress stations in pre-Messinian rocks in the NE Sierra Cabrera (diagram 5; Drenth et al. 1993). The youngest sediments producing this tensor orientation are of earliest Messinian age in the Vera Basin (diagram 4: Turre).

Figure 6 (lower map and diagrams) summarises stress tensors from presumably Late Miocene-Pliocene age. The main principal axis of the stress tensor (σ_1) has an approximately horizontal N-S orientation; the (σ_1–σ_3)-plane is nearly horizontal and σ_2 nearly vertical. This tensor orientation has been measured in Pliocene sediments in the Serrata Fault Zone (diagram 3: Serrata-2); it was obtained from a stress analysis in the Serrata Fault near Carboneras (diagram 4) and was interpreted as a second phase in stress stations Lucainena (diagram 1) and Gafarillos (diagram 2), and in 10 stress stations in the NE Sierra Cabrera (diagram 5; Drenth et al. 1993).

The results of the stress field analysis presented here are comparable to those of Boorsma (1993), who carried out a detailed investigation of the orientation of the stress field during the Neogene in the Serrata Fault Zone (Fig. 1). The fault zone is expressed at the surface by a straight topographic ridge running NE-SW from the coast south of the Sierra Cabrera to the Bay of Almeria, and continues into the NE-SW trending fault system of the Alboran Sea. The fault zone forms a complex anastomosing fault system in the sedimentary cover overlying a major NE-SW trending basement fault, that has been episodically active from the latest Early Miocene up to the present time.

Boorsma (1993) described the relationship between fault movement and sedimentation and the complicated relationship between repeated fault movement and the changing orientation of the regional stress field. He analysed a horizontal WNW-ESE orientation of the largest principal stress (σ_1) during the Langhian and Serravallian, that subsequently rotated into a NW-SE orientation during the Early Tortonian and into N-S orientation during the Late Tortonian to Recent period.

The stress analysis also confirms earlier results published by De Larouzière et al. (1987, 1988) and Montenat et al. (1987), who showed that the Tortonian-Pliocene period is characterised by changing orientations of the main stress field. According to their results the principal compressive stress axis rotates from NW-SE during the Tortonian to N-S in the Late Tortonian-Pliocene and then back to NW-SE during the Late Pliocene-Holocene.

Discussion

Neogene strike-slip deformation and the associated formation of sedimentary basins took place on a continental crust with rheological properties that were severely influenced by the previous phase of Late Oligocene-Early Miocene extension and heating.

Strength profiles of the lithosphere have been calculated by Van der Beek & Cloetingh (1992) along a number of profiles across the eastern and western Betic Cordilleras for the External Zone, Internal Zone and the Alboran Sea. Two of these profiles are shown in Fig. 7. The strength of the lithosphere depends on the composition and the thermal structure of the lithosphere. In their calculations Van der Beek & Cloetingh (1992) used a model with depth-dependent rheology with quartzite composition for the upper crust, diabase composition for the lower crust and olivine composition for the subcrustal lithosphere. Recent heat flow data (Albert-Bertrán 1979) are represented by the contour lines in Fig. 7. The thickness of the continental crust is also indicated. The heat flow pattern indicates considerably higher heat flow in the eastern Betics, coinciding with a thinner continental crust.

The theoretical strength profiles indicate that in the western Betics the upper crust and the upper part of the lower crust have preserved their strength. The lower part of the upper crust is characterised by decrease in strength and this level probably represents a suitable detachment level for thrusting. The lower crust has

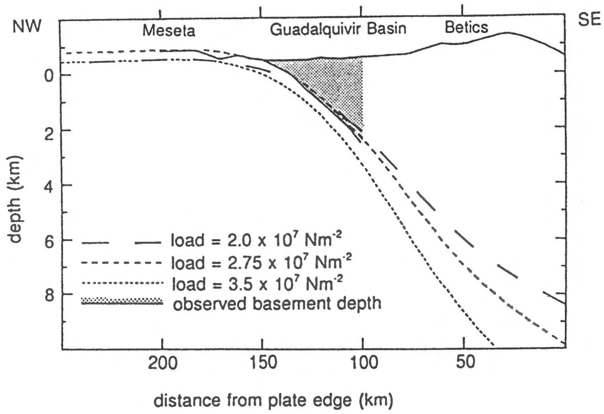

Fig. 8. Two-dimensional finite-difference modelling of the flexural response of the western Betic lithosphere as a function of tectonic loading (profile A in Fig. 7). Broken plate model incorporating distributed subsurface load forces acting over the first 100 km of the profile. Density of topographic load (ρ) = 2750 kg m^{-3}; effective elastic thickness = 10 km. After Van der Beek & Cloetingh (1992) and Cloetingh et al. (1992).

no strength and is characterised by viscous behaviour. The model further predicts a mechanically strong layer (MSL), approximately 10 km thick, in the subcrustal lithosphere that will act as the elastic core during flexure or bending of the plate under stress. The strength profiles of the eastern Betics show a totally different rheological structure. In the Alboran Sea, where Moho depths are very shallow, the upper part of the subcrustal lithosphere preserved its strength. In the External and Internal Zones the upper crust is equally strong but the elevated temperatures have reduced the strength of the lower crust and subcrustal lithosphere to zero. Finally, the mechanically strong layer in the Internal and External Zones is very thin making it improbable that the plate will react by bending or flexure under stress.

It is suggested that strength differences in the lithosphere between the eastern and western Betic Cordilleras, caused by differences in thermal structure and crustal thickness, may well explain the differences in large-scale behaviour between both areas. These differences are an inherited effect of the Late Oligocene-earliest Miocene extensional phase that influenced the eastern part of the Betic Cordilleras in particular (Fig. 4c).

For the western Betic Cordilleras tectonic modelling predicts bending of the lithosphere and development of the Guadalquivir foreland basin under the load of the nappes formed during the final emplacement in the Early Miocene. Van der Beek & Cloetingh (1992) carried out a flexure analysis, modelling the isostatic response of the Iberian lithosphere to the load of the nappes in the Internal Zone and the Subbetic thrust sheets that formed during the final emplacement of the nappes in the Early Miocene. They calculated the down-bending of the Iberian lithosphere underneath the Guadalquivir Basin using an equivalent elastic thickness of 10 km (corresponding to the MSL in the strength profiles) and a flexural rigidity $D = 5.7 \times 10^{21}$ Nm. Using a load of 2.75×10^7 Nm working over the overthrust area, which corresponds to a realistic maximum thickness of 10 km of overthrust units, the calculated curve fits well with the real down-bending of the basement underneath the Guadalquivir Basin as it is known from seismic and well data (Fig. 8).

For the eastern Betics, the lithospheric strength profiles demonstrate a predominantly brittle upper crust on top of a viscous lower crust and subcrustal lithosphere. During compression the upper crust reacts brittly and deforms along strike-slip faults with associated development of pull-apart basins and basement uplifts.

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This paper was presented in a lecture on the orogenic evolution of the Betic Cordilleras at the Symposium ‘Orogenesis, a time odyssey’ that was organised to celebrate the 60th anniversary of the Leidse Geologische Vereniging. It is a summary of work, that has been carried out in recent years by a large number of staff members, PhD-students and students from the Vrije Universiteit in Amsterdam and of the former Geological Institute of the University of Amsterdam. The work is amongst others based on the detailed analysis of the structural and metamorphic evolution of the eastern Sierra de los Filabres that was carried out initially by Wim Zevenhuizen, Koen de Jong and Henk Bakker, under supervision of Otto Simon, Henk Helmers and the author. This work was continued and finished during PhD-research by Koen de Jong. Other data used in this paper concern studies on the strike-slip tectonics of the Ramonete Zone with Otto Simon, Anna Bon and Dik Koenen, Paul Houweling, Jeroen Vastenhouw, Kees de Leeuw and Hans Veldkamp; in the Sierra Cabrera-Sorbas area with Tom Roep, Anne Fortuin, Gerard Miedema, Martin Drenth, Rikkert Moeys and Gerco Stapel and on the Serrata Fault with Leopold Boorsma. The author has further used data on the External Zone of the Betics published by Menno de Ruig, Kay Beets, Tini Geel, Warner ten Kate, Jan

Smit, Jeroen Kenter, John Reymer, Rieks van Straaten and Tim Peper. Data on the tectonic modelling of the Betics are based on research carried out by Sierd Cloetingh, Randell Stephenson, Henk Kooi, Douwe van Rees, Peter van der Beek and Jan Diederik van Wees. This paper is NSG-contribution 950406.

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