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Lithospheric flexure and the tectonic evolution of the Betic Cordilleras (SE Spain)

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ABSTRACT

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We present flexural and gravity models along four profiles across the Betic Cordilleras (SE Spain). The flexural response of the Iberian lithosphere to the loading exerted by the thrust sheets of the internal zone of the orogen is modelled for a broken plate with an elastic rheology of the lithosphere. The conspicuously low value for the flexural rigidity (e.e.t. = 10 km) required to produce a fit of the deflection suggests that a heating event predated the thrusting, and that decoupling takes place along crustal discontinuities. The present topographic load is insufficient to produce the observed deflection and an additional subsurface load is required, which can be explained by overthrusting of a pre-existing rifted margin. The model predicts a thickness of the overthrust units of approximately 10 km.

In the western Betics, the gravity signature results from flexure of the Iberian plate combined with crustal thinning and flank uplift in the Alborán basin and the internal zone. In the eastern Betics, the flexural response is completely overprinted by post-thrusting extensional events. Rheological models of the Iberian lithosphere confirm the results of the flexural and gravity models. Lateral variations in plate rheology, primarily caused by differences in thermal structure, can explain the observed variation in tectonic configuration along strike of the Betic orogen and the eastward disappearance of the Guadalquivir foreland basin. The modelling demonstrates that at least two thermo-tectonic events must have affected the Betics during their orogenic evolution. The first event took place after the Mesozoic extension and development of a rifted margin and before the main phase of overthrusting. This phase reflects most likely the Oligocene–Early Miocene extension which is widely recognised in the Valencia trough and surrounding areas. The second phase, which produced the present configuration of the Betics, corresponds to the Tortonian–Recent extension centred in the Valencia trough and Alborán basin.

Introduction

Although the Betic Cordilleras in southeastern Spain have been the subject of geological research for nearly a century, the study of lithosphere dynamics responsible for their tectonic evolution by means of quantitative geodynamic modelling has only recently begun (e.g., Van Wees et al., 1992; Peper and Cloetingh, 1992). During the last decade, flexural and gravity models have been widely used in the study of collisional orogens around the world (e.g., Quinlan and Beau-

mont, 1984; Moretti and Royden, 1988; Zoete-meijer et al., 1990). The rheology of the lithosphere, which is of prime importance for its response to the emplacement of thrust loads in an orogenic belt (McNutt et al., 1988), is generally inherited from its pre-orogenic evolution. Pre-orogenic phases of extension are commonly recognised in many orogenic belts and found to be of great importance to the orogenic phase of their evolution (Stockmal et al., 1986; Desegaulx et al., 1991). The Betic Cordilleras underwent a complex history of repeated extensional and compressional phases and are, therefore, of particular interest to study the dynamics of the rift–foreland basin transition in polyphase foreland basin evolution. The Betics are also well suited to study

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the dynamics of extensional basin formation in a regime of overall convergence.

In the present study, the orogenic phase of the Betic Cordilleras is studied by means of flexural, gravity and rheological modelling. The flexural response of the Iberian lithosphere to the load emplaced by the thrust sheets of the Betic Cordilleras is studied along four profiles through this orogen. The model is constrained by seismic and well data from the Guadalquivir basin and stratigraphic field data in the eastern Betics. The results of the flexural model are further constrained by gravity modelling. Subsequently, rheological models for the lithosphere are constructed for different parts of the Betic Cordilleras and compared to estimates of the mechanical properties inferred from the flexural and gravity modelling. The results of the modelling demonstrate that extensional events exert a dominant control

on the orogenic evolution of the Betic Cordilleras. At least two spatially and temporally distinct extensional events have affected the orogenic phase of the evolution of the Betic Cordilleras and both have left their tectonic imprint on it.

Tectonic setting

The Betic Cordilleras in southeastern Spain, together with the northern Moroccan Rif, comprise the westernmost part of the circum-Mediterranean Alpine range, and constitute the southernmost deformed margin of the Iberian plate. Their Meso–Cenozoic evolution is associated with the opening of the Atlantic Ocean and the convergence between the Iberian and African plates (Vegas and Banda, 1982; Dercourt et al., 1986). During Late Triassic–Middle Cretaceous times, a left-lateral transtensional movement between

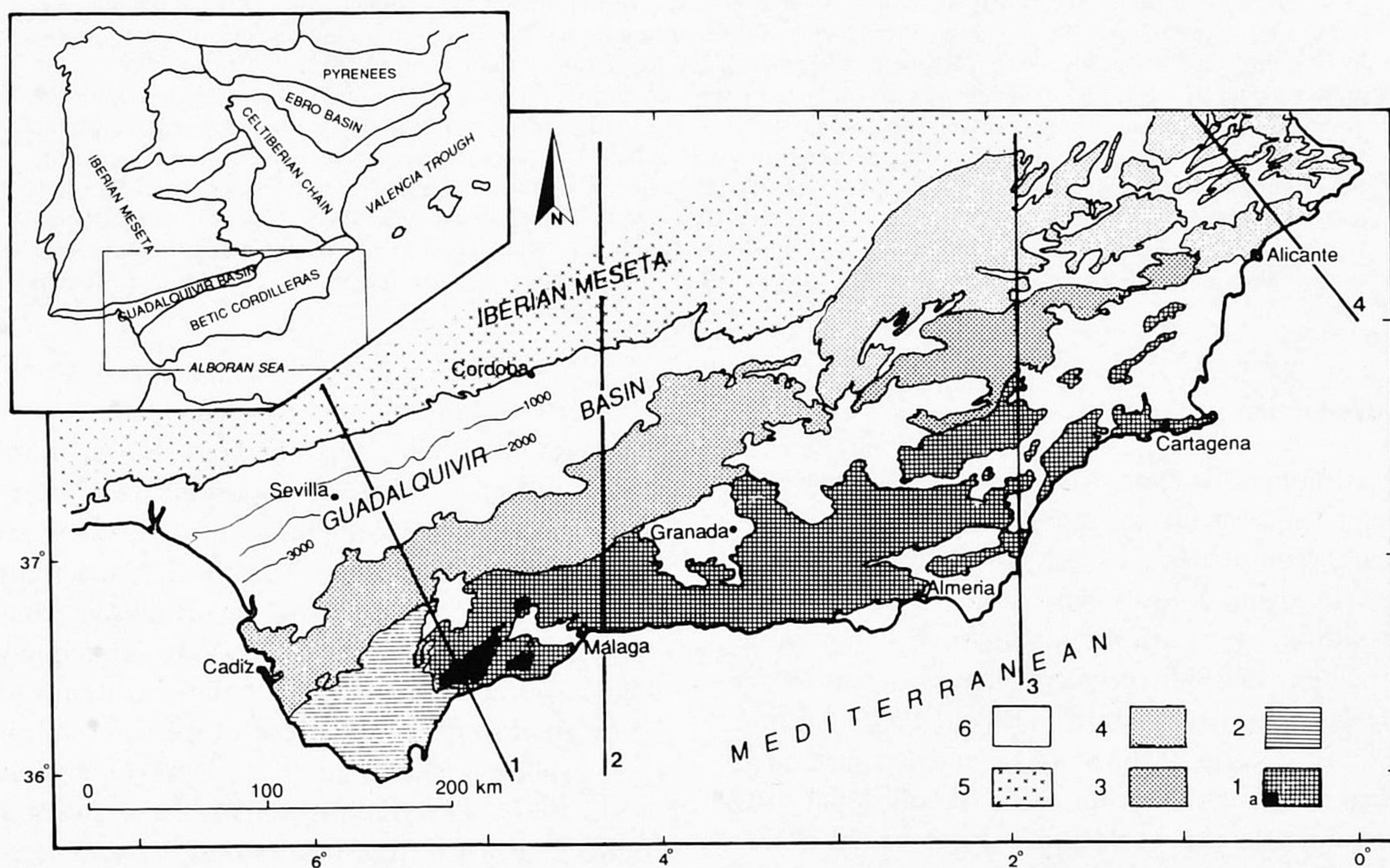


Fig. 1. Generalised geological map of the Betic Cordilleras showing the location of the selected profiles for the flexural modelling: 1 = internal zone (1a = Ronda peridotites); 2 = Campo de Gibraltar flysches; 3 = Subbetic; 4 = Prebetic; 5 = Iberian meseta; 6 = Neogene basins. Isobaths of Pre-Miocene basement (IGME, 1987) are indicated for the Guadalquivir basin. Inset: tectonic subdivision of the Iberian Peninsula (after Vegas and Banda, 1982).

Africa and Eurasia–Iberia gave rise to an extensional tectonic regime and a rifted margin evolved on the southeastern border of the Iberian continent. During and after Late Cretaceous times, convergence of the two plates induced a regional compressive regime which continues up to the present (cf. Bufo et al., 1988). Within this large-scale collisional setting, spatially and temporally localised extension led to the initiation of a number of rifted basins in the western Mediterranean of which the Gulf of Lions, Valencia trough and Alborán basin are prominent examples.

As in many Alpine chains, the Betic Cordilleras can be subdivided into a nonmetamorphic external zone and a metamorphic and highly deformed internal zone (Fig. 1). The internal zone of the Betic Cordilleras is made up of mainly pre-Mesozoic and Triassic terrains. The palaeogeography of these units is not well known, but they are clearly allochthonous to the Iberian continent (Vegas and Banda, 1982; Mäkel, 1985). Three main units (thrust complexes) can be distinguished in the internal zone on the basis of differences in their tectono-metamorphic evolution (Egeler and Simon, 1969; Torres-Roldán, 1979). The contacts between these units represent major thrusts.

The external zones of the Betic Cordilleras represent the Mesozoic–Palaeogene rifted continental margin of Iberia (García-Hernández et al., 1980). The external zones are subdivided into the Prebetic domain, which represents the shelf of the margin, and the Subbetic slope and basin. Triassic syn-rift deposits (evaporites and redbeds) indicate the initiation of rifting. During the Early Jurassic, break-up of the platform occurred and the differentiation into Prebetic and Subbetic took place. From the Jurassic onward to Eocene times, thick sequences of shallow (Prebetic), respectively deep (Subbetic) marine sediments, mainly in calcareous facies, were deposited. Quantitative subsidence analysis and forward modelling of the Subbetic stratigraphy (Peper and Cloetingh, 1992) indicates that three consecutive rifting episodes have affected the external zones during Mesozoic times, leading to a mature rifted margin at the end of the Palaeogene.

The earliest deformation event that is recognised in the Betic realm is a phase of high-pressure–low-temperature metamorphism and associated crustal underthrusting in the internal zone (Bakker et al., 1989). This deformation episode has been tentatively dated as Late Cretaceous (De Jong, 1990). During this period, tectonic activity is absent in the eastern external zones (García-Hernández et al., 1980; Geel et al., 1992). In the western part of the Cordilleras, Upper Cretaceous–Oligocene flysches indicate involvement in early Alpine overthrusting (Balanyá and García-Dueñas, 1987). In palaeogeographic reconstructions, the origin of the internal zone (Alborán domain) is sought east of its present position (Rehault et al., 1984; Dercourt et al., 1986), having moved westwards during Tertiary times (Balanyá and García-Dueñas, 1987; Comas et al., 1990). Together with the westward expulsion of the Alborán domain, subduction of the African plate beneath the Alborán domain and Iberia took place. The dip direction of the subduction zone has long been debated (Vegas and Banda, 1982; Mäkel, 1985; Torres-Roldán et al., 1986), but a northwardly subducted slab of African lithosphere under the Alborán and Iberian plates seems now well established from seismological observations (Bufo et al., 1988), surface wave analyses (Marillier and Mueller, 1985) and seismic tomography (Spakman, 1990). Thus the external zone of the Betic Cordilleras can be interpreted as a continental retro-arc foreland fold and thrust belt.

During Oligocene–Early Miocene times, rifting took place in the western Mediterranean and oceanic crust was formed in the Ligurian, Sardinian–Balearic and North Algerian basins (Rehault et al., 1984). In the internal zone of the Betics, extension took place together with high-temperature–low-pressure metamorphism and intrusion of ultramafic rocks (Bakker et al., 1989). In the external zones, this rifting event is recognised as a period of block-faulting and accelerated subsidence (Ott d'Estevou et al., 1988; Kenter et al., 1990). Contemporaneous with or shortly after the extensional episode in the Alborán domain, nappes of the internal zone started to overthrust the Iberian margin to the north (Mäkel,

1985; Sanz de Galdeano, 1990). The emplacement of the thrust nappes of the internal zone led to deformation in the foreland, with thin-skinned thrusting in the Subbetic and north-vergent folding in the Prebetic. The Guadalquivir foreland basin started to develop during this time (Perconig and Martinez Diaz, 1977).

Following the emplacement of the internal zone thrust complexes, a period of strike-slip faulting affected the Betic–Rif orogen. The onset of strike-slip tectonics is envisaged to be caused by the blocking of thrust movements and continued convergence of Africa and Iberia (Sanz de Galdeano, 1990). These strike-slip movements caused the formation of a number of “post-orogenic” intramontane basins and also led to renewed subsidence of the Alborán basin (Dillon et al., 1980; Comas et al., 1992). After the main tectonic events in the Betic Cordilleras, the region underwent rapid uplift from the late Tortonian–Pliocene onward.

Geological and geophysical data

The analysis of the flexural response has been conducted primarily on the basis of the geometry of the Guadalquivir foreland basin. The depth to basement in this basin is constrained by a large amount of seismic and well data (Perconig and Martinez Diaz, 1977; IGME, 1987). These data indicate that the basement is deflected towards the south under the basin. The Guadalquivir basin started to develop during the Late Miocene and is filled with Neogene–Recent shallow marine and alluvial sediments, and Upper Miocene gravity flows of Subbetic material. The IGME (1987) compilation map of isobaths of the pre-Miocene basement serves as the main constraint for the study of the flexural response of the Iberian lithosphere to loading.

The tectonic configuration changes along strike in the foreland of the Betic Cordilleras. The Guadalquivir basin narrows towards the east and terminates near 3°W in the foreland fold- and thrust-belt of the eastern Betics. The Neogene stratigraphic evolution of the eastern Betics is discussed by, among others, Ott d’Estevou et al. (1988), Kenter et al. (1990) and Geel et al. (1992).

Burdigalian–Serravallian syntectonic deposits occur in rapidly subsiding synclinal basins in a zone about 50 km wide, in line with the Guadalquivir basin (Fig. 1), and overstep to the north. These sediments consist of marls intercalated with mass-flow deposits. The estimated palaeo-water-depth is around 200 m, while the maximum thickness of the sediments is estimated at 1000–1500 m, showing large lateral variation. Kenter et al. (1990) and Geel et al. (1992) suggest that the phase of rapid subsidence in the Serravallian, coincident with the formation of the Guadalquivir basin, was caused by the loading of advancing thrust masses from the south. A foreland trough may have existed in the Burdigalian–Serravallian. Seismic reflection studies performed in the Valencia trough indicate that this fossil foreland trough continues to the east (Roca and Desgaulx, 1991). Subsequent Serravallian strike-slip tectonics and Tortonian extensional events in the Valencia trough may be the cause of the structural reorganisation and uplift of this trough (De Ruig, 1990). Tortonian closure of the inferred foreland trough is also reflected in the sedimentary record of the western Mediterranean basins (e.g., Dillon et al., 1980).

An important constraint imposed on the results of the flexural and gravity analyses is also given by the information from numerous seismic refraction profiles shot in the Betic Cordilleras (Hatzfeld, 1976; Banda and Ansorge, 1980; Banda et al., 1983, and submitted; Zeyen et al., 1985; Medialdea et al., 1986; Banda, 1988; Fig. 2a). These studies show that a crustal root with a maximum Moho depth of about 38 km exists under the western part of the Betic Cordilleras. This root is displaced to the north with respect to the highest topography. The crust thins rapidly towards the Alborán Sea and the crustal thickness at the coastline is only about 24 km. In the eastern part of the Betic chain, no crustal root seems to exist and the crust thins gently from 30 km under the Meseta to less than 20 km in the Valencia trough (Banda and Ansorge, 1980; Zeyen et al., 1985).

Gravity data used in this study (Fig. 2b) were supplied by the Bureau Gravimétrique International (B.G.I.) in Toulouse, France. This database

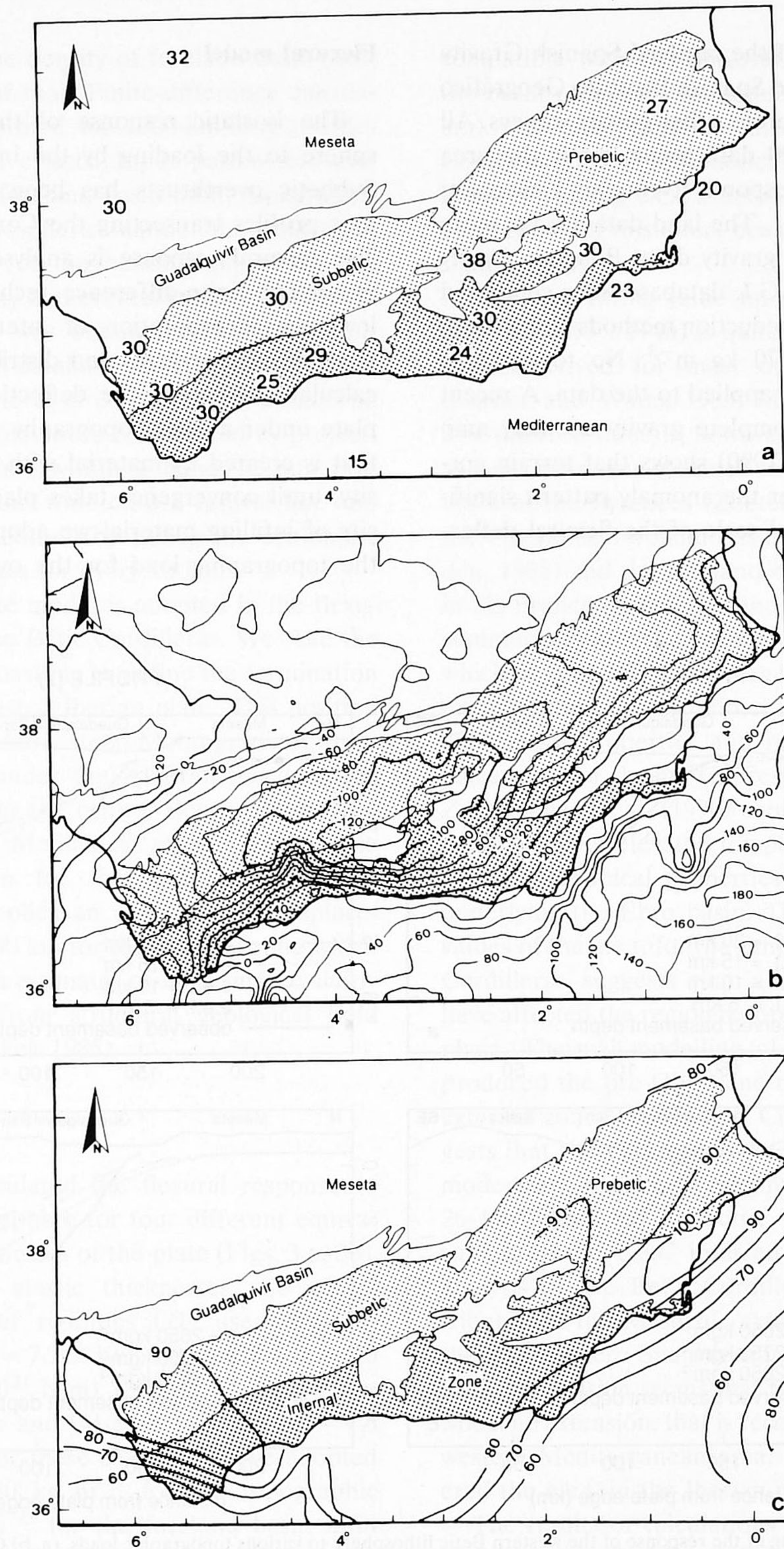


Fig. 2. Geophysical data sets used in the tectonic modelling, plotted on a simplified tectonic map of southern Spain. The area of the Betic Cordilleras is indicated; light shading denotes external zones and darker shading internal zone. (a) Crustal thickness map of the Betic Cordilleras showing the Moho depth in km. Modified from Banda (1988). (b) Simplified Bouguer gravity anomaly map of the Betic Cordilleras, compiled from the B.G.I. database. Contour interval is 20 mGal. Density reduction: 2670 kg m^{-3} . (c) Simplified heat flow map of the Betic Cordilleras. Contour interval is 10 mW m^{-2} . Modified from Albert-Beltrán, 1979b.

includes the data of the regional Spanish Gravity Net, collected by the Spanish Instituto Geográfico Nacional, and numerous other gravity surveys. All together about 8000 data points cover the area studied, which corresponds to roughly 1 measurement every 30 km^{-2} . The land data are far more dense than marine gravity data. Bouguer gravity anomalies in the B.G.I. database were computed using the standard reduction methods and a mean rock density of 2670 kg m^{-3} . No topographic correction has been applied to the data. A recent compilation of a complete gravity anomaly map (Casas and Carbo, 1990) shows that terrain corrections do not alter the anomaly pattern significantly on the spatial scale of the flexural deflection.

Flexural model

The isostatic response of the Iberian lithosphere to the loading by the internal zone and Subbetic overthrusts has been modelled along four profiles transecting the Cordilleras (Fig. 1). The flexural response is analysed using a 2-dimensional finite-difference technique which allows the incorporation of lateral variations in mechanical properties and distributed loads. We calculate iteratively the deflection of an elastic plate under a given topography, filling the space that is created by material with a specified density, until convergence takes place. For the density of infilling material we adopt the density of the topographic load for the overthrust part of

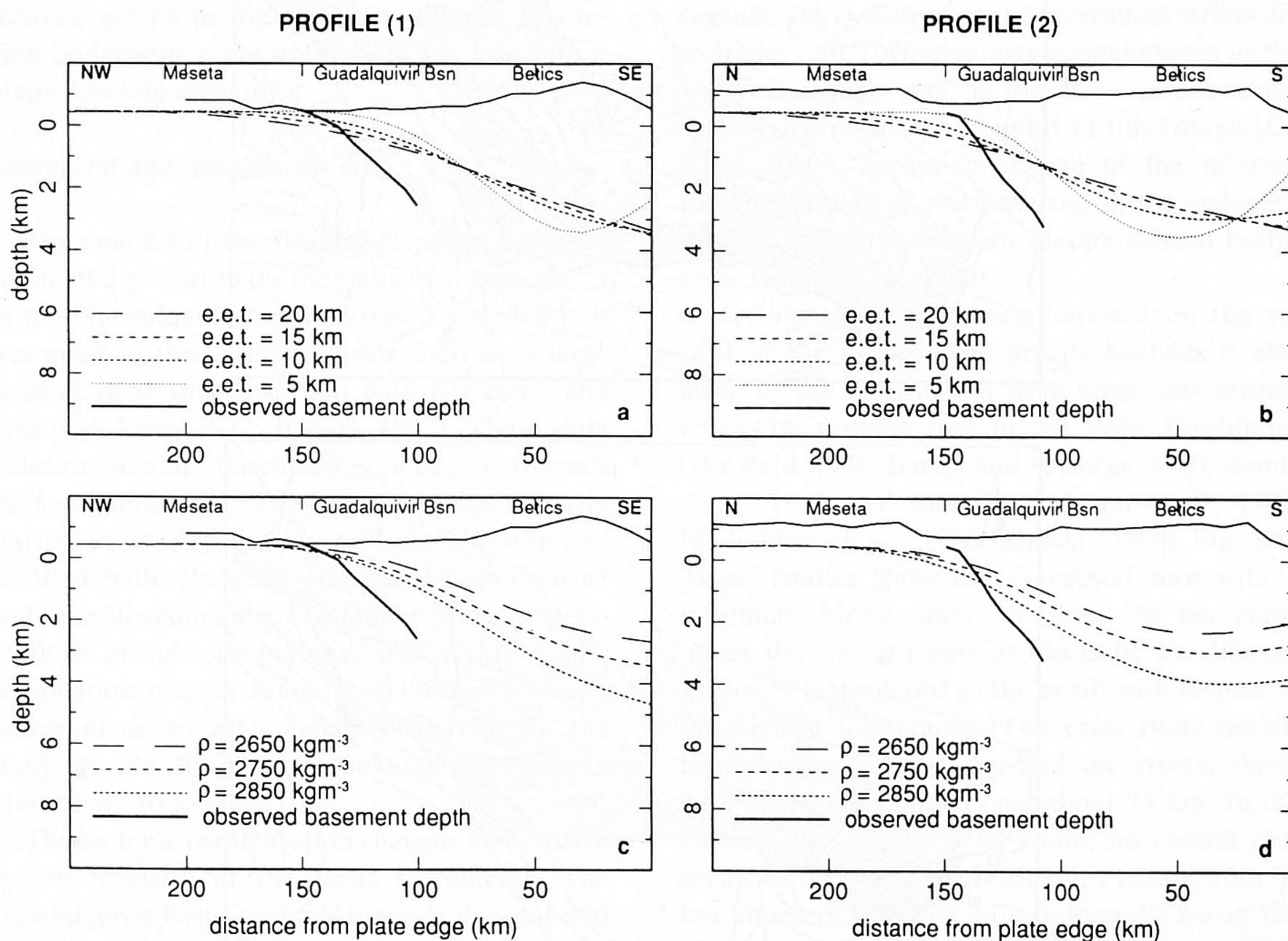


Fig. 3. Flexural modelling of the response of the western Betic lithosphere to various topographic loads. (a, b) Calculated deflection for different equivalent elastic thicknesses of the plate ranging from e.e.t. = 5 km (flexural rigidity $D = 7.2 \times 10^{20} \text{ N m}$) to e.e.t. = 20 km ($D = 4.8 \times 10^{22} \text{ N m}$), for profiles (1) and (2), respectively. Density of topographic load ρ is 2750 kg m^{-3} . (c, d) Calculated deflection for different densities of the topographic load ρ : 2650 kg m^{-3} , 2750 kg m^{-3} and 2850 kg m^{-3} and an e.e.t. = 10 km ($D = 5.7 \times 10^{21} \text{ N m}$), for profiles (1) and (2), respectively.

the plate, and the density of foreland basin sediments in front of that. Finite-difference calculations were performed for 1000-km-long profiles to minimise end effects. Input parameters are: the densities of foreland basin infill, topographic load and asthenospheric mantle, the equivalent elastic thickness of the bending plate, the observed topography and eventual subsurface loads. The topography was obtained from the B.G.I. gravity anomalies database which gives the height above mean sea level of each gravity station. The topographic profiles were constructed by projecting all data points within a band 12 min (approximately 20 km) wide onto a central line and subsequently smoothed by taking the arithmetic mean of the values for every 10 km.

A broken-plate model is adopted in the flexural analysis of the Betic Cordilleras. We take the position of the coastline to define the termination of the underthrust Iberian plate. This position coincides with a fairly steep Moho gradient, from around 17 km under the Alborán Sea to more than 35 km under the central Betics (Banda and Ansorge, 1980; Medialdea et al., 1986). The adopted location for the termination of the Iberian plate implies an overthrusting displacement of about 100 km for the internal zone which is consistent with estimates of near-surface shortening inferred from structural geological field studies (e.g., Mäkel, 1985).

Western Betics

We have calculated the flexural response of the Iberian lithosphere for four different equivalent elastic thicknesses of the plate (Figs. 3 a, 3b). The equivalent elastic thicknesses and corresponding flexural rigidities (D) used are; (1) e.e.t. = 5 km ($D = 7.2 \times 10^{20}$ N m); (2) e.e.t. = 10 km ($D = 5.7 \times 10^{21}$ N m); (3) e.e.t. = 15 km ($D = 2.0 \times 10^{22}$ N m); and (4) e.e.t. = 20 km ($D = 4.8 \times 10^{22}$ N m). For these calculations we adopted densities of 2750 kg m^{-3} for the topographic load, 2400 kg m^{-3} for the foreland basin infill and 3300 kg m^{-3} for the asthenospheric mantle, respectively.

Large values of the e.e.t. (15 and 20 km) produce a very wide and shallow basin that is not

compatible with the observations. The width of the basin requires a low value of the e.e.t. (5–10 km). This value, however, fails to produce a sufficiently deep basin. Numerical experiments using laterally varying e.e.t.'s demonstrate that the incorporation of this effect does not improve the fit significantly.

The best fitting value for the e.e.t. of 10 km ($D = 5.7 \times 10^{21}$ N m) is quite low compared with values derived for most other Alpine chains (Karner and Watts, 1983; McNutt et al., 1988). The inferred value is, however, in the same range as e.e.t. estimates obtained for the Ebro foreland basin of the Pyrenees (Zoetemeijer et al., 1990), the Apennine foreland basin (Moretti and Royden, 1988) and the foreland of the Western Alps in SE France (S. Guellec and P. Desegaulx, pers. commun., 1990). The rheology of the lithosphere, which is controlled to a large extent by its thermal structure, is of prime importance for the mechanical properties and response to the emplacement of loads (expressed in the e.e.t.). Zoetemeijer et al. (1990) showed that the weakening effect of elevated temperatures is reflected in the mechanical properties of the lithosphere underlying the Ebro basin. The existence of low values of the e.e.t. for the lithosphere of the Betic Cordilleras suggests that a heating event may have affected the region before overthrusting took place. Thermal modelling of the extension that produced the pre-Oligocene rifted margin of the external zones (Peper and Cloetingh, 1992) suggests that the Palaeogene e.e.t. was lowered only moderately, reaching a minimum value of around 26 km under the Subbetic. Therefore, the low values of the e.e.t. inferred from the flexural analysis of the Betic Cordilleras cannot be attributed to the Mesozoic extension that affected the southeastern Spanish margin. The most plausible explanation seems that the Oligocene–Early Miocene extension, that is recognised in the whole western Mediterranean area, has drastically lowered the e.e.t. in the Iberian plate.

The results of calculations for an elastic plate with an e.e.t. of 10 km deflected under the topographic load of the Betic Cordilleras, for different densities attributed to the load, are shown in Figures 3c and d. Inspection of these figures

demonstrates that none of the adopted density distributions can produce a basin that is sufficiently deep to fit the observed depth of the basement under the Guadalquivir basin. Clearly, the density of a complex thrust belt is not a unique or well constrained value and can only be approximated to a first order by a mean density. A value of 2850 kg m^{-3} for the load is the upper limit of what is geologically feasible, even when the incorporation of large masses of ultramafic rock (Ronda peridotites) in the thrust sheets is taken into account. Furthermore, about one third of the volume of the overthrusting units is occupied by Subbetic sediments with relatively low

densities. Therefore, it is evident that the topographic load alone cannot account for the deflection of the basement observed under the Guadalquivir basin.

If the observed topography is insufficient to account for the observed deflection of the basement two approaches can be taken to improve the fit: (1) introducing shear forces and/or bending moments at the free end of the plate by changing the boundary conditions (cf. Royden and Karner, 1984), or (2) introducing a (number of) subsurface load forces that may act anywhere on the plate (cf. Karner and Watts, 1983). The introduction of forces and bending moments at

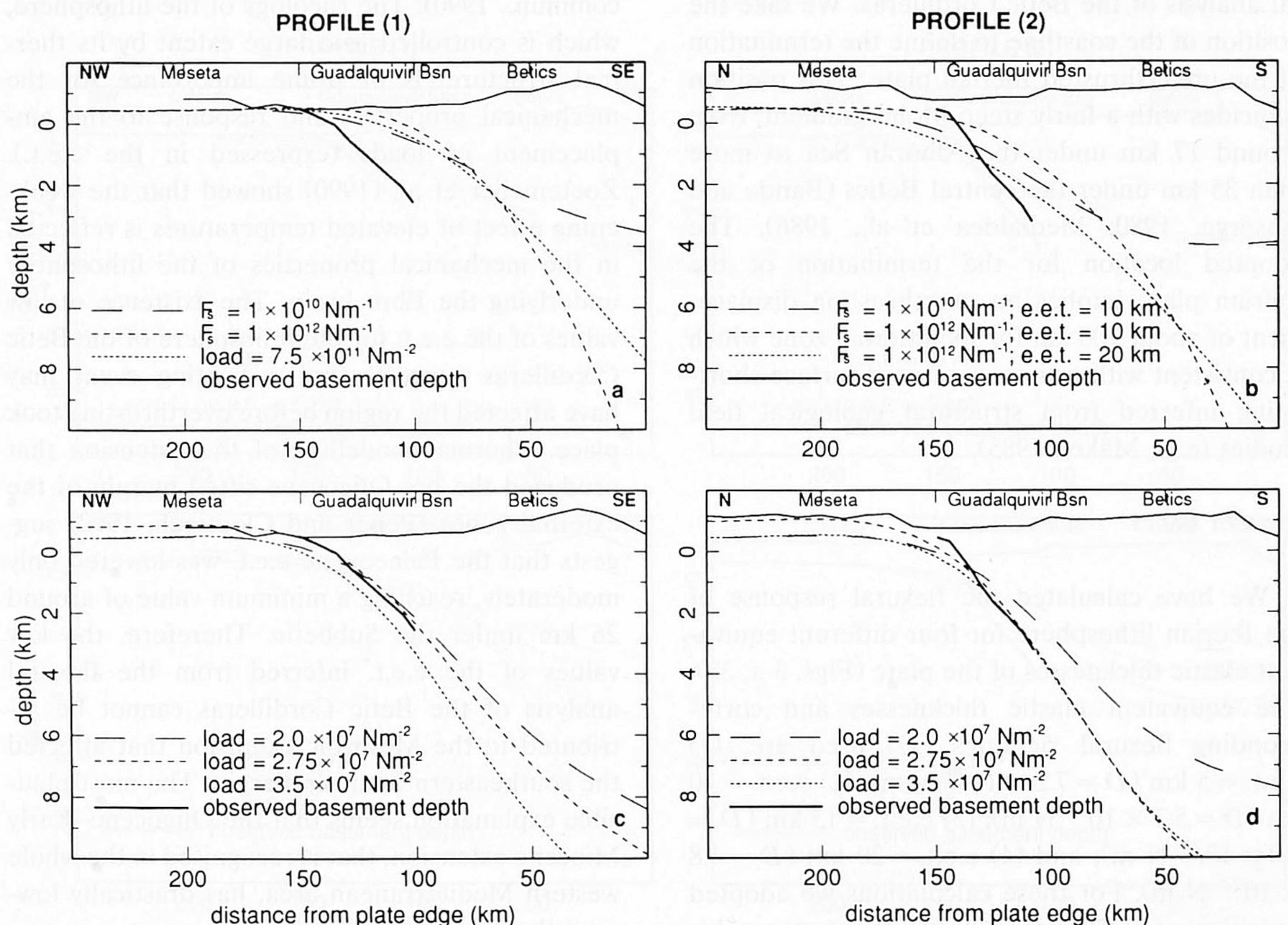


Fig. 4. Flexural modelling of the response of the western Betic lithosphere to various tectonic loads. (a) Calculated deflection for profile (1) incorporating shear forces (F_s) at the free end of the plate and an e.e.t. of 10 km, and result including a concentrated load force of $7.5 \times 10^{11} \text{ N m}^{-2}$ at $x = 15 \text{ km}$ (modelled effect of peridotitic body extending down to mantle depth). (b) Calculated deflection for profile (2) incorporating shear forces (F_s) at the free end of the plate and varying e.e.t.'s of 10 and 20 km. (c, d) Calculated deflection for profiles (1) and (2), respectively, including distributed subsurface load forces acting over the first 100 km of the profiles. Other parameters are: e.e.t. = 10 km; density of topographic load ρ is 2750 kg m^{-3} .

the end of the plate can improve the fit of the model but the tectonic mechanism by which they are created is neither clear nor unique. Lyon-Caen and Molnar (1989) showed analytically that any geometry can be fitted by manipulating the boundary conditions. Therefore, apart from finding a combination of shear force and bending moment that fits the model, it is crucial to investigate whether this combination results in a testable tectonic mechanism.

The deflection of profile (1) has been modelled by applying a shear force at the free end of the plate (Figs. 4a, 4b). A shear force alone, however, cannot account for the deflection of the basement under the Guadalquivir basin. The effect of applying a shear force at $x = 0$ is mainly that the deflection increases under the internal zone located near the free end of the plate. The area for which the deflection increases, can be widened by adopting a larger e.e.t. (Fig. 4b), but the steep gradient of the basement under the Guadalquivir basin, relatively far from the free end of the plate cannot be induced by a shear force applied there. This would require an excessively large shear force and an equally large but counteracting (!) bending moment (cf. Lyon-Caen and Molnar, 1989). A concentrated load may possibly act on the plate as a result of the involvement of mantle material (Ronda peridotites) in thrust slices. Several authors (e.g., Bonini et al., 1973) have argued that a peridotitic body extending down to the mantle exists in the area. We have modelled the deflection caused by such a body, corresponding to a concentrated load of $7.5 \times 10^{11} \text{ N m}^{-1}$ located at $x = 15 \text{ km}$ (Fig. 4a). Since the Ronda peridotites are located in the internal zone close to $x = 0$, the result is very similar to the effect of applying a shear force at the free end of the plate. The additional load necessary to fit the deflection of the Iberian lithosphere cannot be explained by the loading effect of the Ronda peridotites.

In the light of the above, the only possibility to successfully model the observed deflection is by applying a distributed additional vertical load to the plate. The results of modelling the deflection incorporating a distributed load acting over the total width of the overthrust area are depicted in

Figures 4c and 4d. A good fit to the observed deflection of the basement under the Guadalquivir basin is obtained for an additional distributed load of $2.75 \times 10^7 \text{ N m}^{-2}$ acting over the overthrust area. Such a load is consistent with thrusting of the internal zone of the Betic Cordilleras onto a rifted margin constituted by the external zones. As demonstrated by Stockmal et al. (1986) and Nunn et al. (1987), overthrusting of a rifted margin can result in a load that is significantly larger than the topographic load. This is because pre-existing loads inherited from the rifted margin stage may still contribute to the loading in the orogenic stage and because the overthrust is emplaced onto a negative (palaeo)topography. The value for an additional distributed load of $2.75 \times 10^7 \text{ N m}^{-2}$, inferred from our analysis, would imply overthrusting onto a pre-existing rifted margin with a 1 km deep bathymetry, in the absence of pre-existing load forces. The configuration and bathymetry of this pre-existing Betic rifted margin is consistent with palaeogeographic reconstructions of the external zones of the Betic Cordilleras (García-Hernández et al., 1980).

Our modelling predicts a maximum thickness of overthrusting units in the internal zone of the Betics in the order of 10 km. The cumulative thicknesses of the internal zone thrust units are in agreement with this value. Also, recent seismic experiments (Banda et al., submitted) indicate that a prominent reflector, which is interpreted as a detachment surface, exists at this depth.

Eastern Betics

The absence of a foreland basin and a crustal root in the eastern Betics leaves no direct constraint for a flexural analysis of the present-day tectonic situation there. For this reason, we have modelled the Serravallian palaeogeography using the reconstructions discussed previously. The modelling of the deflection under the topographic load exerted by the Subbetic and internal zones for profile (3) (Fig. 5a) demonstrates that also in this case the observed topographic load fails to explain the depth of the reconstructed basin. A model incorporating the same distributed subload of $2.75 \times 10^7 \text{ N m}^{-2}$ over the

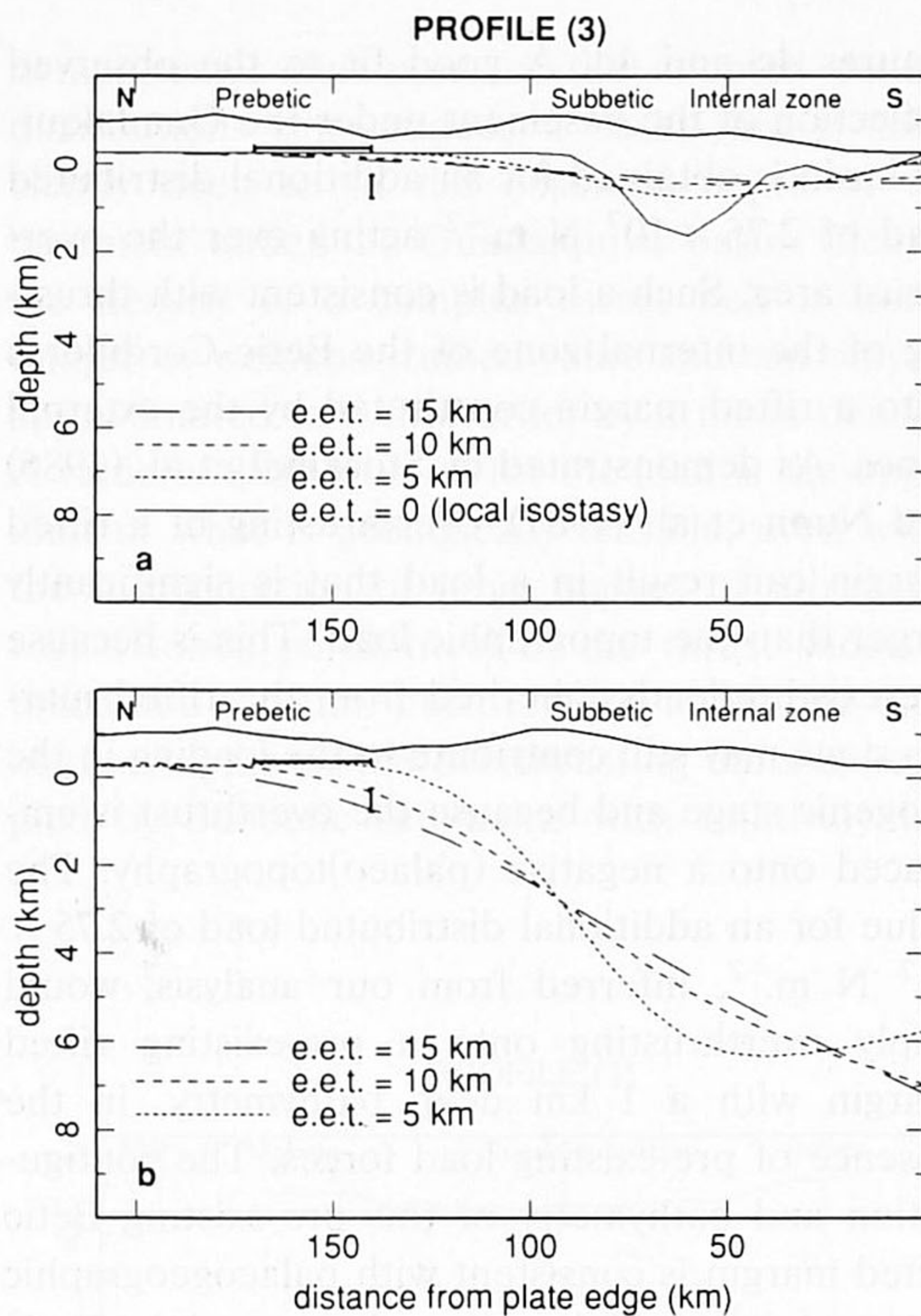


Fig. 5. Calculated deflection of profile (3) for different equivalent elastic thicknesses of the plate ranging from 5 km (flexural rigidity $D = 7.2 \times 10^{20}$ N m) to 20 km ($D = 4.8 \times 10^{22}$ N m). (a) Zero subload; horizontal and vertical bars indicate reconstructed width and depth range of the inferred Serravallian foreland basin. (b) Effect of incorporating a distributed subsurface load force that equals 2.75×10^7 N m⁻² acting over the first 100 km of the profile (same model as for western Betics).

total width of the overthrust zone does provide a good fit for an e.e.t. of 10 km (Fig. 5b).

The situation is less clear for profile (4) located in the Alicante area. This is primarily the result of the fact that the overthrusting units are not exposed on land in this area. These units have, however, been recognised in seismic profiles offshore in the Valencia trough, and underwent subsidence since their emplacement over the Prebetic, probably as a result of later extensional events affecting the Valencia trough (Roca and Desegaulx, 1992). We have modelled the deflection under a hypothetical "wedge" with a maximum height of 500 m and extending the first 25 km of the profile. The modelling suggests that

the basin here could be the result of a similar flexural response as in the other profiles.

We, therefore, conclude that the postulated existence of a Serravallian foreland trough of the inferred dimensions in the eastern Betics is supported by a flexural model assuming the same parameters for the mechanical properties of the lithosphere and distribution of the applied load as in the western Betics. It seems that a foreland basin may have extended from east to west along the entire length of the Betic orogen and has disappeared in the eastern part as a result of later tectonic events.

Gravity model

We have used the results of the flexural analysis described above to calculate synthetic gravity anomaly profiles along the selected transects. These profiles were constructed by calculating theoretical Bouguer anomalies for 2-dimensional crustal structure models based on the method of Talwani et al. (1959).

Gravity profiles were extended into the Alborán Sea to include the gravity effects of the Alborán basin (cf. Fig. 2b). The geometry of the Alborán basin is relatively well known from seismic, DSDP and commercial well data (Dillon et al., 1980; IGME, 1987; Comas et al., 1992). We have used the IGME (1987) isopach map to constrain the thickness of Neogene sediments in the basin and calculated associated crustal thicknesses assuming local isostasy. This assumption is probably a simplification since the dynamics of extension in the Alborán basin are not taken into account. However, the inferred crustal thicknesses of 17–20 km are in good qualitative agreement with seismic refraction results. Densities assigned to the various bodies are: 2400 kg m⁻³ for the sediments of the Guadalquivir basin, 2600 kg m⁻³ for the Prebetic sediments, 2650 kg m⁻³ for the Subbetic overthrusts, 2200 kg m⁻³ for the Alborán basin sediments, 2850 kg m⁻³ for the internal zone and Alborán crust, 2850 kg m⁻³ for the Iberian crust, 3250 kg m⁻³ for the Alborán subcrustal lithosphere and 3300 kg m⁻³ for the Iberian subcrustal lithosphere. In order to obtain a fit to the gravity data, a separation of the

overthrust load into Subbetic and internal zone units is required. As the mean density of these two units equals the overthrust density adopted in the flexural model, this modification does not significantly alter the results of the flexural model. The adopted low density of the subcrustal lithosphere under the Alborán basin is consistent with estimates from analyses of P_n velocities and gravity anomalies (Hatzfeld, 1976; Banda, 1988).

Western Betics

In a first attempt, we have taken the results of the flexural analysis as input for the gravity calculations, with a constant crustal thickness for the (underthrust) Iberian crust of 30 km. Figures 6a and 6b demonstrate that the fit of the gravity anomalies for this model is not satisfactory. The model fails to reproduce the coastal gravity peak

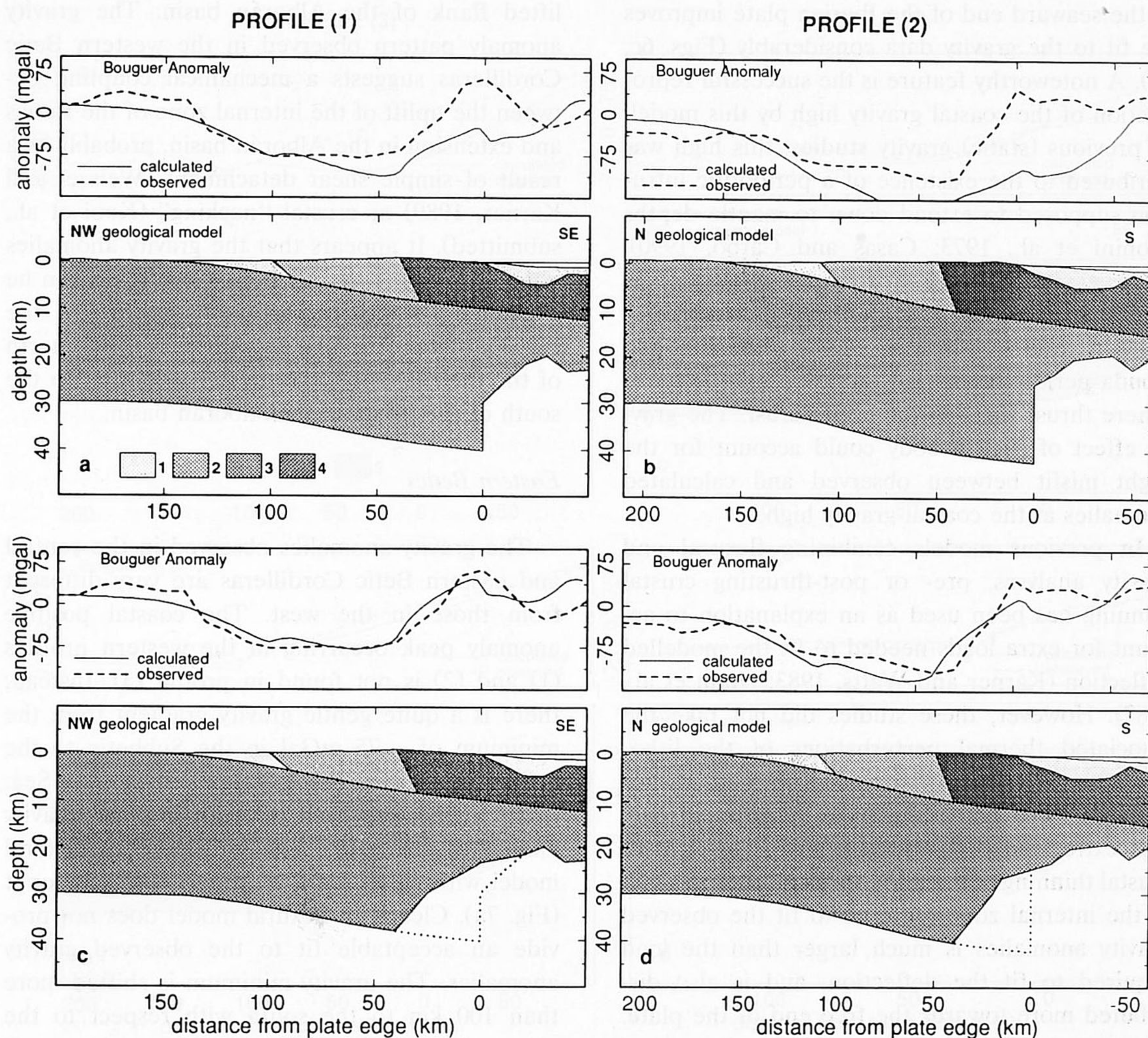


Fig. 6. Gravity models for the western Betic Cordilleras. (a, b) Gravity anomalies calculated for a model of flexure with a constant crustal thickness of the underthrust Iberian plate, for profiles (1) and (2), respectively. Shading: 1 = Guadalquivir ($\rho = 2400 \text{ kg m}^{-3}$) and Alborán ($\rho = 2200 \text{ kg m}^{-3}$) basins; 2 = Subbetic ($\rho = 2650 \text{ kg m}^{-3}$); 3 = Iberian crust ($\rho = 2850 \text{ kg m}^{-3}$); 4 = internal zone and Alborán crust ($\rho = 2850 \text{ kg m}^{-3}$). (c, d) Gravity anomalies calculated for a model of flexure and crustal attenuation in the internal zone and Alborán basin, for profiles (1) and (2), respectively. Shading and assigned densities as in (a) and (b). Stippled lines indicate crustal configuration of (a) and (b), respectively.

and produces anomalies that are too low for the whole internal zone and Subbetic. The model is also not compatible with seismic refraction data for this part of the Betics (Medialdea et al., 1986; Banda, 1988) as these data indicate that the maximum crustal thickness is not located at the coastline, under the internal zone, but rather somewhat to the north of it. Clearly, crustal thinning was not confined to the Alborán basin but also affected the underthrust Iberian plate.

Incorporation of the effect of crustal thinning in the seaward end of the Iberian plate improves the fit to the gravity data considerably (Figs. 6c, 6d). A noteworthy feature is the successful reproduction of the coastal gravity high by this model. In previous (static) gravity studies, this high was attributed to the existence of a peridotitic intrusion supposed to extend down to mantle depths (Bonini et al., 1973; Casas and Carbo, 1990). However, it appears from structural observations (Tubia and Cuevas, 1986) and recent seismic refraction results (Barranco et al., 1990) that the Ronda peridotites are an unrooted mantle lithosphere thrust sheet in the upper crust. The gravity effect of such a body could account for the slight misfit between observed and calculated anomalies at the coastal gravity high.

In previous models combining flexural and gravity analyses, pre- or post-thrusting crustal thinning has been used as an explanation to account for extra loads needed to fit the modelled deflection (Karner and Watts, 1983; Nunn et al., 1987). However, these studies did not take the associated thermal perturbations of the lithosphere into account, which result into an upward directed force and reduce the subsurface load. The extra load that results from the amount of crustal thinning of the underthrust Iberian plate in the internal zone required to fit the observed gravity anomalies is much larger than the load required to fit the deflection, and is also distributed more towards the free end of the plate. The extra load created by crustal thinning ranges from $2.97 \times 10^7 \text{ N m}^{-2}$ at $x = 30 \text{ km}$ to $7.3 \times 10^7 \text{ N m}^{-2}$ at $x = 0$ in profile (1) and from $2.84 \times 10^7 \text{ N m}^{-2}$ at $x = 30 \text{ km}$ to $8.7 \times 10^7 \text{ N m}^{-2}$ at $x = 0$ in profile (2) (cf. Fig. 6). We postulate, therefore, that the dynamics of crustal thinning are much

more complex than a pure geometrical reduction of the thickness of the underthrust Iberian plate. It is interesting to note that the internal zone of the Betic Cordilleras has experienced an uplift of at least 200 m since the Tortonian, simultaneous with extension and subsidence of the Alborán basin (Estévez and Sanz de Galdeano, 1980; Weijermars et al., 1985; Cloetingh et al., 1992). It is likely that this uplift is associated with attenuation in the internal zone of the Betics which has thereby become an uplifted flank of the Alborán basin. The gravity anomaly pattern observed in the western Betic Cordilleras suggests a mechanical coupling between the uplift of the internal zone of the Betics and extension in the Alborán basin, probably as a result of simple shear detachment (Weissel and Karner, 1989) or crustal "necking" (Kooi et al., submitted). It appears that the gravity anomalies observed in the western Betic Cordilleras can be explained by the interaction of flexure of the Iberian plate (with an e.e.t. = 10 km) to the north of the internal zone and crustal extension to the south of this zone, in the Alborán basin.

Eastern Betics

The gravity anomalies observed in the central and eastern Betic Cordilleras are very different from those in the west. The coastal positive anomaly peak occurring in the western profiles (1) and (2) is not found in profile (3). Instead, there is a quite gentle gravity gradient from the minimum of -75 mGal in the Subbetic to the positive peak of $+100 \text{ mGal}$ in the Alborán Sea.

As a first approach to modelling the gravity anomalies associated with profile (3), a flexural model with an e.e.t. of 10 km was used as input (Fig. 7a). Clearly, a flexural model does not provide an acceptable fit to the observed gravity anomalies. The gravity minimum is shifted more than 100 km to the south with respect to the observed anomalies. The incorporation of crustal thinning in the internal zone as described in profiles (1) and (2) would result in a gravity peak at the coastline, which is not observed here. The fit is remarkably better for a model of local (Airy) isostatic response (Fig. 7b). This model results in

a crustal thickness of 35 km under the Subbetic, which accords reasonably well with seismic refraction results, and a gradual shallowing of the Moho towards the Alborán Sea as a combined result of local isostasy and attenuation. The small-wavelength negative anomaly superimposed on this signal in the Subbetic can be explained by the effect of large intramountainous basins (Casas and Carbo, 1990).

As the modelling of the gravity anomalies observed at profile (3) supports the existence of local isostasy, the same approach was taken in the first model for profile (4) (Fig. 7c). The resulting model, however, is not compatible with the gravity anomalies observed at profile (4). In contrast with profile (3), there is no clear minimum. Instead, the anomaly profile grades gently from around +50 mGal offshore to around -50 mGal

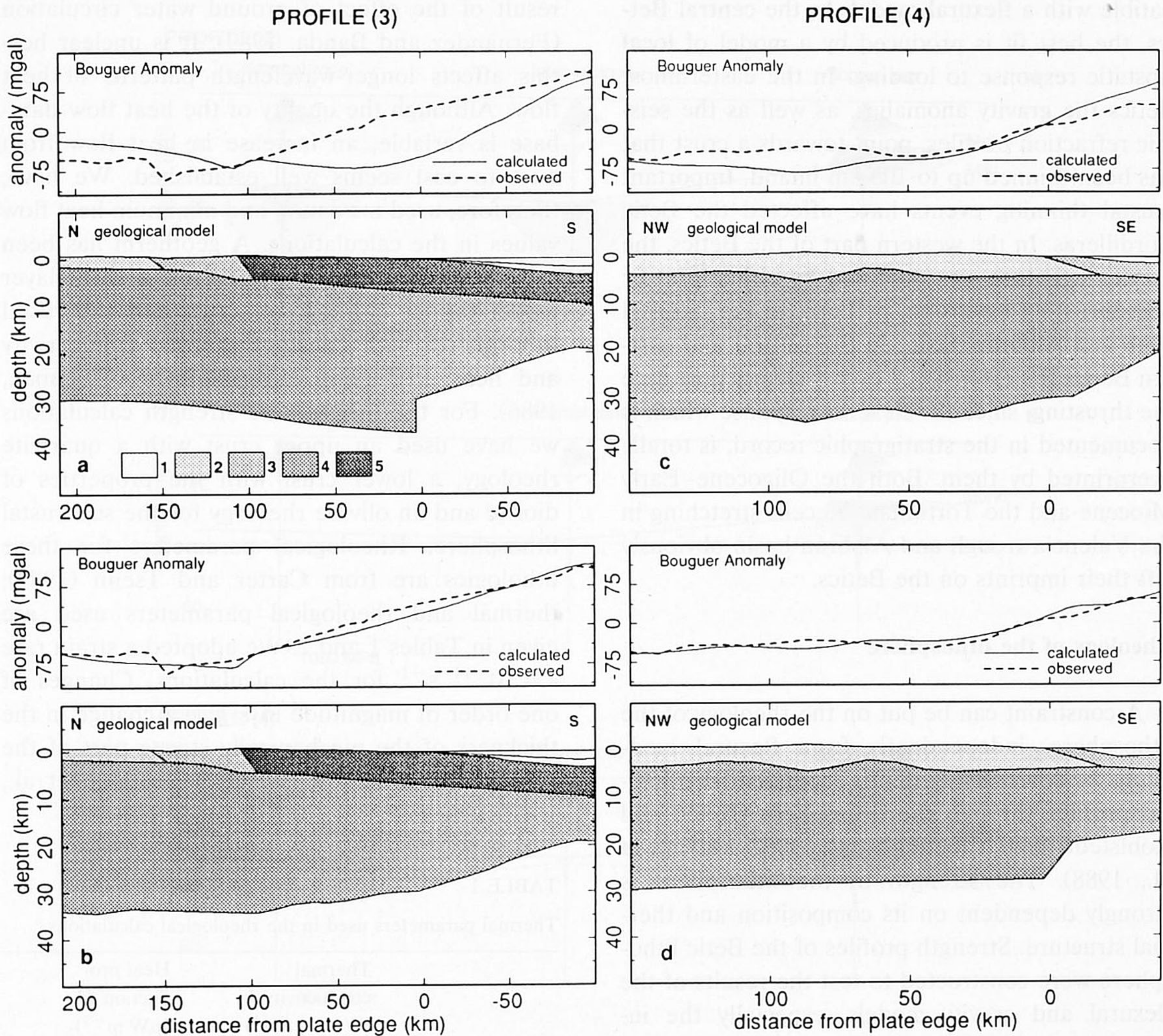


Fig. 7. Gravity models for the eastern Betic Cordilleras. (a) Model of flexure (e.e.t. = 10 km) and crustal thinning (local isostatic) under the Alborán basin for profile (3). Shading: 1 = Alborán basin sediments ($\rho = 2200 \text{ kg m}^{-3}$); 2 = Prebetic ($\rho = 2600 \text{ kg m}^{-3}$); 3 = Subbetic ($\rho = 2650 \text{ kg m}^{-3}$); 4 = Iberian crust ($\rho = 2850 \text{ kg m}^{-3}$); 5 = internal zone and Alborán crust ($\rho = 2850 \text{ kg m}^{-3}$). (b) Model of local isostatic response and crustal thinning under the Alborán basin for profile (3). Shading and assigned densities as in (a). (c) Model of local isostatic response and crustal thinning offshore for profile (4). Shading and assigned densities as in (a). (d) Model incorporating crustal attenuation up to 100 km inland for profile (4). Shading and assigned densities as in (a).

100 km inland, being quite flat on the inland portion. Seismic refraction profiling (Zeyen et al., 1985) has shown that in the Alicante area the crust is thinned in areas located as far as 100 km inland. This crustal thinning can be related to the extension in the Valencia trough. A gravity model assuming such an amount of crustal thinning strongly improves the fit to the observed gravity anomalies (Fig. 7d).

In the central and eastern parts of the Cordilleras, the gravity signature is thus not compatible with a flexural model. In the central Betics, the best fit is produced by a model of local isostatic response to loading. In the easternmost Betics the gravity anomalies, as well as the seismic refraction profiles, point towards a crust that has been thinned up to 100 km inland. Important crustal thinning events have affected the Betic Cordilleras. In the western part of the Betics, the gravity signature was modified by either pre- or post-thrusting extension, or both, in the internal zone and Alborán basin. In the central and eastern Betics the extensional events clearly post-date the thrusting, since the flexural response which is documented in the stratigraphic record, is totally overprinted by them. Both the Oligocene–Early Miocene and the Tortonian–Recent stretching in the Valencia trough and Alborán basin obviously left their imprints on the Betics.

Rheology of the lithosphere

A constraint can be put on the rheology of the lithosphere, independently from flexural modelling, by constructing depth-dependent rheological profiles through the lithosphere (Brace and Kohlstedt, 1980; Cloetingh et al., 1982; McNutt et al., 1988). The strength of the lithosphere is strongly dependent on its composition and thermal structure. Strength profiles of the Betic lithosphere were constructed to test the results of the flexural and gravity models, especially the inferred conspicuously low value of the e.e.t. and the along-strike variation in mechanical properties.

The strength profiles have been constructed in a two-step procedure in which we first analytically calculate a geotherm from a given surface

heat flow and then use this result, together with a given petrological model of the lithosphere, to calculate the strength at certain depth intervals. Heat flow values used (Fig. 2c) were taken from the map published by Albert-Beltrán (1979b). These values are high (in excess of 70 mW m^{-2}) for the whole area of the Betic Cordilleras, but reach extremely high values ($90\text{--}100 \text{ mW m}^{-2}$) in the Gulf of Cadíz in the west and the Alicante area and Valencia trough in the east (Albert-Beltrán, 1979a, b). Locally high values may be the result of the effect of ground water circulation (Fernández and Banda, 1989). It is unclear how this affects longer-wavelength patterns of heat flow. Although the quality of the heat flow database is variable, an increase in heat flow from west to east seems well established. We have, therefore, used maximum and minimum heat flow values in the calculations. A geotherm has been calculated from these values for a three-layer model (upper crust, lower crust and subcrustal lithosphere) with constant thermal conductivity and heat production in each layer (Chapman, 1986). For the lithospheric strength calculations we have used an upper crust with a quartzite rheology, a lower crust with the properties of diorite and an olivine rheology for the subcrustal lithosphere. Rheological parameters for these lithologies are from Carter and Tsenn (1987); thermal and rheological parameters used are given in Tables 1 and 2. We adopted a strain rate $\dot{\epsilon} = 10^{-14} \text{ s}^{-1}$ for the calculations. Changes of one order of magnitude in $\dot{\epsilon}$ give a change in the thickness of the mechanically strong part of the lithosphere of only a few km (Cloetingh et al.,

TABLE 1

Thermal parameters used in the rheological calculations^a

	Thermal conductivity k ($\text{W m}^{-1} \text{K}^{-1}$)	Heat production H ($\mu\text{W m}^{-3}$)
Upper crust	2.51	2.5
Lower crust	2.1	0.42
Lithospheric mantle	3.1	0

^a parameters used are the same as those of Albert-Beltrán (1979a, b).

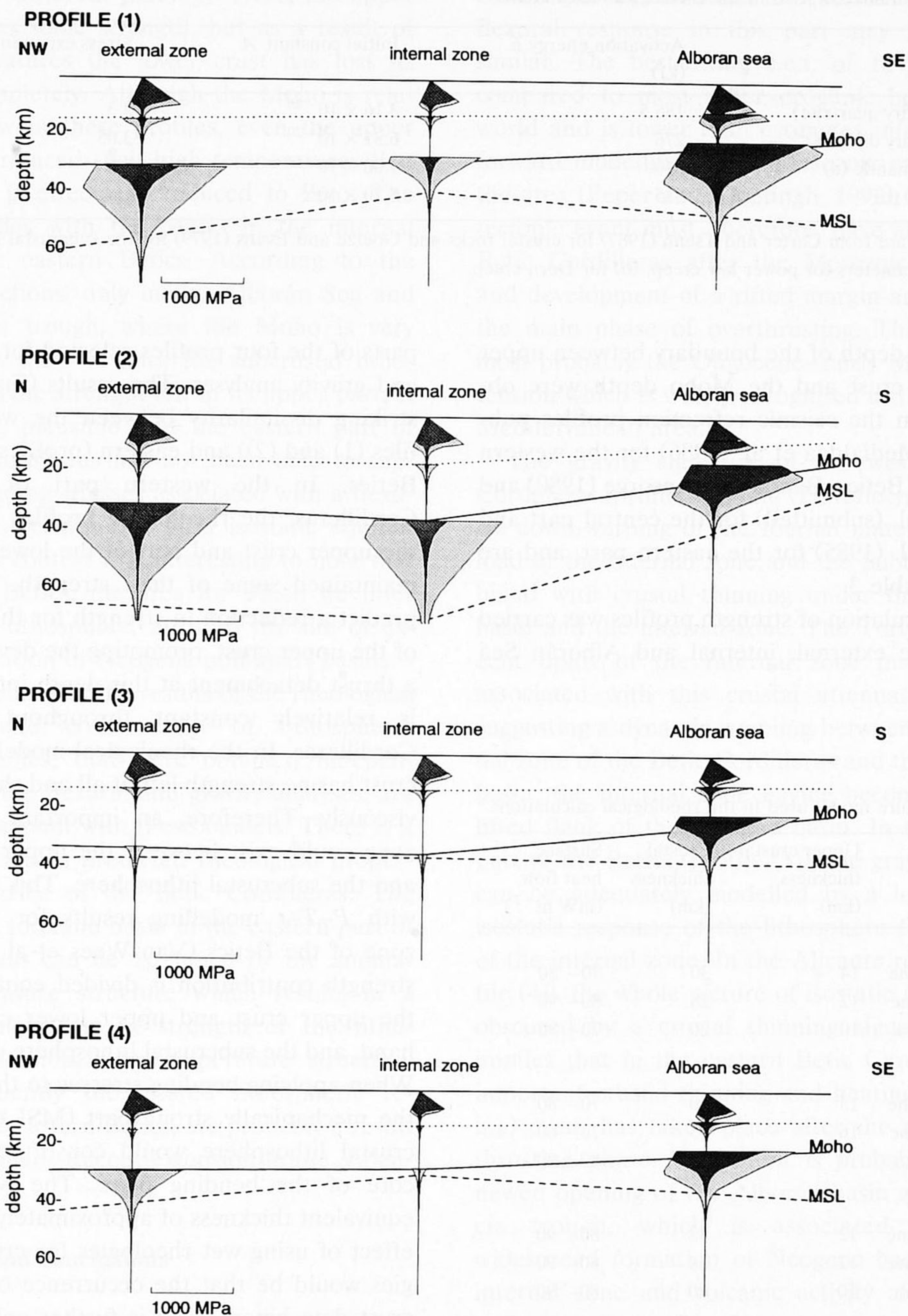


Fig. 8. Lithospheric strength profiles calculated for the external zone of the Betics, internal zone of the Betics and Alborán Sea along the four selected profiles. Light and darker shading indicate strength profiles corresponding to minimum and maximum heat flow estimates, respectively. Heat flow estimates range from 70–90 mW m^{-2} in the western Betics to 80–100 mW m^{-2} in the eastern Betics (cf. Table 3). A layered lithospheric model was used with a quartzite upper crust, diabase lower crust and olivine subcrustal lithosphere. Thermal and rheological parameters used are given in Tables 1 and 2. Also shown are upper/lower crust boundary, Moho depth and the lower boundary of the mechanically strong part of the lithosphere (MSL), corresponding to a ductile strength of 50 MPa in the subcrustal lithosphere.

TABLE 2

Rheological parameters used in the rheological calculations ^a

	Activation energy <i>E</i> (kJ)	Initial constant <i>A</i>	Stress exponent <i>n</i>
Upper crust (dry quartzite)	186.5	3.16×10^{-26}	3.3
Lower crust (dry diabase)	276	6.31×10^{-20}	3.05
Lithospheric mantle (a)	510	7.00×10^{-14}	3.0
(dry olivine) (b)	535	5.70×10^{11}	

^a parameters are from Carter and Tsenn (1987) for crustal rocks and Goetze and Evans (1979) for the subcrustal lithosphere; (a) denotes parameters for power law creep, (b) for Dorn creep.

1982). The depth of the boundary between upper and lower crust and the Moho depth were obtained from the seismic refraction profiles published by Medialdea et al. (1986) for the western part of the Betics, Banda and Ansorge (1980) and Banda et al. (submitted) for the central part and Zeyen et al. (1985) for the eastern part, and are given in Table 3.

The calculation of strength profiles was carried out for the external, internal and Alborán Sea

parts of the four profiles selected for the flexural and gravity analyses. The results (Fig. 8) show a striking dissimilarity between the western (profiles (1) and (2)) and eastern (profiles (3) and (4)) Betics. In the western part of the Betic Cordilleras, the rheological profiles suggest that the upper crust and part of the lower crust have maintained some of their strength. The models predict a reduction in strength for the lower part of the upper crust, promoting the development of a thrust detachment at this depth interval, which is relatively constant throughout the Betic Cordilleras. In the rheological models, the lower crust has no strength left at all and should behave viscously. Therefore, an important detachment zone could exist between the upper lower crust and the subcrustal lithosphere. This is consistent with *P-T-t* modelling results for the internal zone of the Betics (Van Wees et al., 1992). The strength contribution is divided equally between the upper crust and upper lower crust on one hand, and the subcrustal lithosphere on the other. When applying bending stresses to these profiles, the mechanically strong part (MSL) of the subcrustal lithosphere would constitute the elastic core of the bending plate. The MSL has an equivalent thickness of approximately 10 km. The effect of using wet rheologies for crustal lithologies would be that the occurrence of the lower-crust detachment zone is further enhanced. The strength of the upper crust would not be altered significantly, since it is mainly limited by brittle failure. Thus, it seems that the rheological strength profiles calculated for the western part of the Betic Cordilleras are consistent with the flexural model that predicts an e.e.t. of 10 km.

TABLE 3

Crustal structure model used in the rheological calculations ^a

	Upper crustal thickness (km)	Crustal thickness (km)	Surface heat flow (mW m ⁻²)
Profile (1)			
external zone	14	30	70– 80
internal zone	12	28	80– 90
Alborán Sea	10	20	70– 80
Profile (2)			
external zone	15	30	70– 80
internal zone	16	35	70– 80
Alborán Sea	7	15	80– 90
Profile (3)			
external zone	12	32	80– 90
internal zone	12	34	80– 90
Alborán Sea	10	20	80– 90
Profile (4)			
external zone	14	28	80– 90
internal zone	12	27	90– 100
Alborán Sea	10	20	80– 90

^a crustal structure was obtained from seismic refraction data (Banda and Ansorge, 1980; Banda et al., 1983, and submitted; Zeyen et al., 1985; Medialdea et al., 1986).

The eastern profiles through the Betic point to a completely different rheology. Here, the upper crust still has some strength, but as a result of high temperatures the lower crust has lost its strength completely. Although the Moho is relatively shallow in these profiles, even the upper mantle experienced such high temperatures that its strength is effectively reduced to zero. The MSL coincides with the Moho in the internal zone of the eastern Betics. According to the model predictions, only in the Alborán Sea and the Valencia trough, where the Moho is very shallow (less than 20 km), the subcrustal lithosphere has some strength left in its upper part. It is, therefore, plausible that the eastern part of the Betic Cordilleras has not been able to support any bending stresses associated with a flexural response, resulting in local isostatic equilibrium. In this context it is interesting to note that the eastern Betics, the area for which we infer the weakest lithosphere, are also the site of extensive formation of Neogene pull-apart basins.

It thus seems that the results of the rheological modelling and construction of lithospheric strength profiles, that were obtained independently from the flexural and gravity analyses, are in good agreement with these models. There is a strong variation in predicted rheological properties along strike of the Betic Cordilleras. The absence of a foreland basin in the eastern part of the Cordilleras can be explained by the anomalous temperature structure which results in a nearly complete loss of strength of the lithosphere. This anomalous temperature structure, and consequently the altered lithospheric response in the eastern Betics, are probably a result of the Tortonian–Recent extension in the Valencia trough.

Discussion and conclusions

The flexural response of the western part of the Betic Cordilleras can be explained by the loading effect of the thrust sheets of the internal zone if this loading took place onto a pre-existing rifted margin of Iberia. The corresponding equivalent elastic thickness of the Iberian lithosphere is estimated at 10 km (corresponding to a flexural

rigidity $D = 5.7 \times 10^{21}$ N m). The analysis of the eastern part of the Betics indicates that the initial flexural response in this part may have been similar. The best-fitting e.e.t. of 10 km is low compared to most other orogenic belts in the world and is lower than estimates inferred from forward modelling of the Mesozoic extension of the area (Peper and Cloetingh, 1992). A thermo-tectonic event must, therefore, have affected the Betic Cordilleras after the Mesozoic extension and development of a rifted margin and prior to the main phase of overthrusting. This event is most probably the Oligocene–Early Miocene extension which is widely recognised in the western Mediterranean area.

The gravity signature of the western Betic Cordilleras [profiles (1) and (2)] reflects the flexural down-warping of the Iberian plate under the load of the internal zone and the Subbetic, combined with crustal thinning under the Alborán basin and the internal zone. The Tortonian–Recent uplift of the internal zone may well be associated with this crustal attenuation event, suggesting a dynamic coupling between the internal zone of the Betic Cordilleras and the Alborán basin, the internal zone having become the uplifted flank of the Alborán basin. In the central part of the Betics (profile (3)), the gravity profile can be adequately modelled by a local (Airy) isostatic response of the lithosphere to the load of the internal zone. In the Alicante region [profile (4)], the whole picture of isostatic response is obscured by a crustal thinning signature. This implies that in the eastern Betic Cordilleras an important crustal thinning and heating (weakening) event has taken place after the main overthrusting phase. This event is probably the renewed opening of the Alborán basin and Valencia trough, which is associated with the widespread formation of Neogene basins in the internal zone and volcanic activity along major strike-slip faults. The lateral variation in structural characteristics of these Neogene basins, with a strong contrast in spatial scale from approximately 20 km wide pull-aparts in the west to approximately 5 km wide pull-aparts in the east, could also be controlled by the lateral variation in plate rheology (Cloetingh et al., 1992).

Rheological models confirm the results obtained from the flexural and gravity analyses. We observe a striking dissimilarity between the eastern and western Betics. The western part shows a detachment zone between the upper lower crust and the upper mantle, with the upper mantle constituting the elastic core of the down-flexed Iberian plate. An e.e.t. of 10 km is consistent with the predictions from the strength profiles. In the eastern Betics, however, the lithosphere has apparently lost its strength completely as a result of the high temperatures encountered here. The crustal thinning associated with the opening of the Valencia trough seems to be the major cause of this feature.

These findings impose important constraints on geodynamic models for the Betic Cordilleras. A crucial feature is the occurrence of two major extensional events which have affected the Cordilleras during the orogenic phase of their evolution. Both left their imprint on the rheology. The extension and heating during the Oligocene–Early Miocene lowered the e.e.t. in the Iberian plate considerably, leaving a hot and weakened crust to be overthrust by the internal zone thrusts. The Tortonian–Pliocene extension and uplift modified the picture again, producing flank uplift and a further thinning of the underthrust plate in the southwest. In the eastern part of the Betics, the combined effect of extension in the Alborán basin and the Valencia trough led to a complete loss of strength of the lithosphere and the attainment of local isostasy. The lateral variation of mechanical properties along strike of the Betic Cordilleras, inferred from our analysis, are important for understanding the dynamics of extensional basin formation in an overall convergent setting.

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References

- Albert-Beltrán, J.F., 1979a. Heat flow and temperature gradient data from Spain. In: V. Čermak and L. Rybach (Editors), *Terrestrial Heat Flow in Europe*. Springer, Berlin, pp. 261–266.
- Albert-Beltrán, J.F., 1979b. El mapa español de flujos caloríficos. Intento de correlación entre anomalías geotérmicas y estructura cortical. *Bol. Geol. Min.*, 90: 36–48.
- Bakker, H.E., De Jong, K., Helmers, H. and Biermann, C., 1989. The geodynamic evolution of the Internal Zone of the Betic Cordilleras (southeast Spain): a model based on structural analysis and geothermobarometry. *J. Metamorph. Geol.*, 3: 359–381.
- Balanyá, J.C. and García-Dueñas, V., 1987. Les directions structurales dans le Domaine d'Alboran de part et d'autre du Détroit de Gibraltar. *C.R. Acad. Sci. Paris, Sér. II*, 304: 929–933.
- Banda, E., 1988. Crustal parameters in the Iberian Peninsula. *Phys. Earth Planet. Inter.*, 51: 222–225.
- Banda, E. and Ansorge, J., 1980. Crustal structure under the central and eastern part of the Betic Cordillera. *Geophys. J.R. Astron. Soc.*, 63: 515–532.
- Banda, E., Udías, A., Mueller, St., Mezcuca, J., Boloix, M., Gallart, J. and Aparicio, A., 1983. Crustal structure beneath Spain from deep seismic sounding experiments. *Phys. Earth Planet. Inter.*, 31: 277–280.
- Banda, E., Gallart, J., García-Dueñas, V., Dañobeita, J.J. and Makris, J., submitted. Lateral variation of the crust in the Iberian Peninsula. New evidence from the Betic Cordillera. *Tectonophysics*.
- Barranco, L.M., Ansorge, J. and Banda, E., 1990. Seismic refraction constraints on the geometry of the Ronda peridotitic massif (Betic Cordilleras, Spain). *Tectonophysics*, 184: 379–392.
- Bonini, W.E., Loomis, T.P. and Robertson, J.D., 1973. Gravity anomalies, ultramafic intrusions, and the tectonics of the region around the Strait of Gibraltar. *J. Geophys. Res.*, 78: 1372–1382.
- Brace, W.F. and Kohlstedt, D.L., 1980. Limits on lithospheric stress imposed by laboratory experiments. *J. Geophys. Res.*, 85: 6248–6252.
- Bufo, E., Udías, A. and Colombás, M.A., 1988. Seismicity, source mechanisms and tectonics of the Azores–Gibraltar plate boundary. *Tectonophysics*, 152: 89–118.
- Carter, N.L. and Tsenn, M.C., 1987. Flow properties of continental lithosphere. *Tectonophysics*, 136: 27–63.
- Casas, A. and Carbo, A., 1990. Deep structure of the Betic Cordillera derived from the interpretation of a complete Bouguer anomaly map. *J. Geodyn.*, 12: 137–147.

- Chapman, D.S., 1986. Thermal gradients in the continental crust. In: J.B. Dawson, D.A. Hall and K.H. Wedepohl (Editors), *The Nature of the Lower Continental Crust*. Geol. Soc. Spec. Publ., 24: 63–70.
- Cloetingh, S.A.P.L., Wortel, M.J.R. and Vlaar, N.J., 1982. Evolution of passive continental margins and initiation of subduction zones. *Nature*, 297: 139–142.
- Cloetingh, S., van der Beek, P.A., van Rees, D. Roep, Th.B., Biermann, C. and Stephenson, R., 1992. Flexural interaction and the dynamics of Neogene extensional basin formation in the Alborán-Betic region. *Geomar. Lett.*, in press.
- Comas, M.C., García-Dueñas, V. and Jurado, M.J., 1992. Neogene tectonic evolution of the Alborán Basin from MCS data. *Geomar. Lett.*, in press.
- De Jong, K., 1990. Alpine tectonics and rotation pole evolution of Iberia. *Tectonophysics*, 184: 279–296.
- Dercourt, J., Zonenshain, L.P., Ricou, L.-E., Kazmin, V.G., Le Pichon, X., Knipper, A.L., Grandjacquet, C., Sbortschikov, I.M., Geysant, J., Lepvrier, C., Pechersky, D.H., Boulin, J., Sibuet, J.-C., Savostin, L.A., Sorokhtin, O., Westphal, M., Bazhenov, M.L., Lauer, J.P. and Bijou-Duval, B., 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. *Tectonophysics*, 123: 241–315.
- De Ruig, M.J., 1990. Extensional tectonics in the eastern Prebetic, Alicante region, SE Spain (abstract). *Proc. Symp. Valencia Trough: Geology and Geophysics. Terra Abstr.*, 2: 10.
- Desegaulx, P., Kooi, H., Cloetingh, S. and Moretti, I., 1991. Consequences of foreland basin development on thinned continental lithosphere: application to the Aquitaine basin (SW France). *Earth Planet. Sci. Lett.*, 106: 116–132.
- Dillon, W.P., Robb, J.M., Greene, H.G. and Lucena, J.C., 1980. Evolution of the continental margin of southern Spain and the Alboran Sea. *Mar. Geol.*, 36: 205–226.
- Egeler, C.G. and Simon, O.J., 1969. Sur la tectonique de la zone Bétique (Cordillères Bétiques, Espagne). *Verh. K. Ned. Acad. Wetensch., Afd. Natuurkd., Reeks 1*, 25: 1–90.
- Estévez, A. and Sanz de Galdeano, C., 1980. Néotectonique du secteur central des Chaînes Bétiques (Bassins de Guadix-Baza et de Grenade). *Rev. Géogr. Phys. Géol. Dyn.*, 21: 23–34.
- Fernández, M. and Banda, E., 1989. An approach to the thermal field in northeastern Spain. *Tectonophysics*, 164: 259–266.
- García Hernández, M., Lopez-Garrido, A.C., Rivas, P., Sanz de Galdeano, C. and Vera, J.A., 1980. Mesozoic paleogeographic evolution of the External Zone of the Betic Cordilleras. *Geol. Mijnbouw*, 59: 155–368.
- Geel, T., Roep, Th.B., Ten Kate, W. and Smit, J., 1992. Miocene stratigraphic turning points in the Alicante region (SE Spain): reflections of the Western Mediterranean plate-tectonic reorganizations. *Sediment. Geol.*, 75: 417–433.
- Hatzfeld, D., 1976. Étude sismologique et gravimétrique de la structure profonde de la mer d'Alboran: mise en évidence d'un manteau anormal. *C.R. Acad. Sci. Paris, Sér. D*, 283: 1021–1024.
- IGME, 1987. Contribución de la exploración petrolífera al conocimiento de la Geología de España. Instituto Geológico y Minero de España, Madrid, 465 pp.
- Karner, G.D. and Watts, A.B., 1983. Gravity anomalies and flexure of the lithosphere at mountain ranges. *J. Geophys. Res.*, 88: 10449–10477.
- Kenter, J.A.M., Reymer, J.J.G., Van der Straaten, H.C. and Peper, T., 1990. Facies patterns and subsidence history of the Jumilla-Cieza region (SE Spain). *Sediment. Geol.*, 67: 263–280.
- Kooi, H., Burrus, J. and Cloetingh, S., submitted. Lithospheric necking and regional isostasy at extensional basins, Part 1. Subsidence and gravity modeling with an application to the Gulf of Lions margin (SE France). *J. Geophys. Res.*
- Lyon-Caen, H. and Molnar, P., 1989. Constraints on the deep structure and dynamic processes beneath the Alps and adjacent regions from an analysis of gravity anomalies. *Geophys. J. Int.*, 99: 19–32.
- Mäkel, G.H., 1985. The geology of the Malaguide complex and its bearing on the geodynamic evolution of the Betic-Rif orogen (southern Spain and northern Morocco) (Ph.D. Thesis). *GUA Pap. Geol. Ser.*, 1, 22, 263 pp.
- Marillier, F. and Mueller, St., 1985. The Western Mediterranean region as an upper-mantle transition zone between two lithospheric plates. *Tectonophysics*, 118: 113–120.
- McNutt, M., Diament, M. and Kogan, M.G., 1988. Variations of elastic plate thickness at continental thrust belts. *J. Geophys. Res.*, 93: 8825–8838.
- Medialdea, T., Suriñach, E., Vegas, R., Banda, E. and An-sorge, J., 1986. Crustal structure under the western end of the Betic Cordillera (Spain). *Ann. Geophys.*, 4: 457–464.
- Moretti, I. and Royden, L., 1988. Deflection, gravity anomalies and tectonics of doubly subducted continental lithosphere: Adriatic and Ionian seas. *Tectonics*, 7: 875–893.
- Nunn, J.A., Czerniak, M. and Pilger, R.H., Jr., 1987. Constraints on the structure of Brooks Range and Colville Basin, Northern Alaska, from flexure and gravity analysis. *Tectonics*, 6: 603–617.
- Ott d'Estevou, P., Montenat, C., Ladure, F. and Pierson d'Autrey, L., 1988. Evolution tectono-sédimentaire du domaine prébétique oriental (Espagne) au Miocène. *C.R. Acad. Sci. Paris, Sér. II*, 307: 789–796.
- Peper, T. and Cloetingh, S., 1992. Lithosphere dynamics and tectono-stratigraphic evolution of the Mesozoic Betic rifted margin (southeastern Spain). In: E. Banda and P. Santanach (Editors), *Geology and Geophysics of the Valencia Trough, Western Mediterranean*. *Tectonophysics*, 203: 345–361.
- Perconig, E. and Martinez Diaz, C., 1977. Perspectivas petrolíferas de Andalucía Occidental. *Bol. Geol. Min.*, 88: 417–433.
- Quinlan, G.M. and Beaumont, C., 1984. Appalachian thrusting, lithospheric flexure, and the Paleozoic stratigraphy of

- the Eastern Interior of North America. *Can. J. Earth Sci.*, 21: 973–996.
- Rehault, J.P., Boillot, G. and Mauffret, A., 1984. The western Mediterranean basin geological evolution. *Mar. Geol.*, 55: 447–477.
- Roca, E. and Desegaulx, P., 1992. Geological evolution and vertical movement analysis of the Valencia Trough area (western Mediterranean). *Mar. Pet. Geol.*, in press.
- Royden, L. and Karner, G.D., 1984. Flexure of lithosphere beneath Apennine and Carpatian foredeep basins: evidence for an insufficient topographic load. *Am. Assoc. Pet. Geol. Bull.*, 68: 704–712.
- Sanz de Galdeano, C., 1990. Geologic evolution of the Betic Cordilleras in the Western Mediterranean, Miocene to the present. *Tectonophysics*, 172: 107–119.
- Spakman, W., 1990. Tomographic images of the upper mantle below central Europe and the Mediterranean. *Terra Nova*, 2: 542–553.
- Stockmal, G.S., Beaumont, C. and Boutelier, R., 1986. Geodynamic models of convergent margin tectonics: the transition from rifted margin to overthrust belt and the consequences for foreland basin development. *Am. Assoc. Pet. Geol. Bull.*, 70: 181–390.
- Talwani, M., Worzel, J.L. and Landisman, M., 1959. Rapid gravity computation of two dimensional bodies with application to Mendocino submarine fracture zone. *J. Geophys. Res.*, 64: 95–137.
- Torres-Roldán, R.L., 1979. The Tectonic subdivision of the Betic Zone (Betic Cordilleras, southern Spain): its significance and one possible geotectonic scenario for the westernmost Alpine belt. *Am. J. Sci.*, 279: 19–51.
- Torres-Roldán, R.L., Poli, G. and Peccerillo, A., 1986. An Early Miocene arc–tholeiitic magmatic dike event from the Alborán Sea—Evidence for precollisional subduction and back-arc crustal extension in the westernmost Mediterranean. *Geol. Rundsch.*, 75: 219–234.
- Tubia, J.M. and Cuevas, J., 1986. High-temperature emplacement of the Los Reales peridotite nappe (Betic Cordillera, Spain). *J. Struct. Geol.*, 8: 473–482.
- Van Wees, J.D., De Jong, K. and Cloetingh, S., 1992. Two dimensional $P-T-t$ modelling and the dynamics of extension and compression in the Betic zone (SE Spain). In: E. Banda and P. Santanach (Editors), *Geology and Geophysics of the Valencia Trough, Western Mediterranean*. *Tectonophysics*, 203: 305–324.
- Vegas, R. and Banda, E., 1982. Tectonic framework and Alpine evolution of the Iberian Peninsula. *Earth Evol. Sci.*, 4: 320–343.
- Weijermars, R., Roep, Th.B., Van den Eeckhout, B., Postma, G. and Kleverlaan, K., 1985. Uplift history of a Betic fold nappe inferred from Neogene–Quaternary sedimentation and tectonics (in the Sierra Alhamilla and Almería, Sorbas and Tabernas Basins of the Betic Cordilleras, SE Spain). *Geol. Mijnbouw*, 64: 397–411.
- Weissel, J.K. and Karner, G.D., 1989. Flexural uplift of rift flanks due to mechanical unloading of the lithosphere during extension. *J. Geophys. Res.*, 94: 13919–13950.
- Zeyen, H.J., Banda, E., Gallart, J. and Ansorge, J., 1985. A wide angle seismic reconnaissance survey of the crust and upper mantle in the Celtiberian Chain of eastern Spain. *Earth Planet. Sci. Lett.*, 75: 393–402.
- Zoetemeijer, R., Desegaulx, P., Cloetingh, S., Roure, F. and Moretti, I., 1990. Lithospheric dynamics and tectonic–stratigraphic evolution of the Ebro basin. *J. Geophys. Res.*, 95: 2701–2711.