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Crustal structure and the mechanical properties of extended continental lithosphere in the Valencia trough (western Mediterranean)

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Abstract: Seismic refraction and reflection, gravity and well data have been used to constrain the structure and evolution of the Valencia trough: a young extensional basin in the western Mediterranean. Seismic refraction data reveal the present-day crustal structure, while backstripping of seismic reflection and well data reveal the structure *at the time of rifting*. By comparing the seismically constrained crustal structure to that inferred from backstripping, we have been able to evaluate whether the Moho is in the expected position, given the tectonic subsidence/uplift and sediment loading history of the trough. The best overall fit between the predicted and seismically constrained crustal structure in the trough is for a model in which the extended continental crust and lithosphere has acquired little or no strength following rifting. A low strength is in accord with the observed free-air gravity anomaly as well as results from other types of rift basins such as the North Sea and the continental margin off the east coast of the USA. There are discrepancies between the observed and predicted Moho depths beneath the flanks of the Valencia trough, but they are attributed to either the removal of crustal material during or following rifting (Iberian peninsula), or to orogenic loading of previously extended continental crust (Mallorca). The largest discrepancy (up to 10 km) correlates with the Ebro delta which has prograded into the trough by more than 60 km since the Pliocene. The discrepancy cannot be explained by the same large-scale processes that modified the Moho beneath the trough flanks and requires that the lithosphere beneath the Ebro delta region is associated with a high ($60 < T_e < 80$ km), rather than low, flexural strength. We do not believe, however, that the high values indicate that extended continental lithosphere is fundamentally strong. Rather, we attribute it to incomplete stress relaxation of the lithosphere during loading.

During the past decade there have been a number of advances in our understanding of the evolution of sedimentary basins, especially rift-type basins which originate in extensional settings. This advance has come about primarily because of the increase in availability of seismic reflection, as well as refraction and other geophysical data over basins together with the development of 'backstripping' techniques to quantitatively analyse their subsidence/uplift history using biostratigraphical data from deep commercial wells.

Recent studies (e.g. Barton & Wood 1984; Royden & Horvath 1988; Dyment 1990; Watts & Torné in press) have adopted approaches that *integrate* seismic, gravity and geoid, well and heat flow data into self-consistent models for basin structure. In the case of crustal thickness, however, these data sets give complementary, but different types of information. Seismic refraction, for example, reveals the present day crustal structure of the basin, whereas backstripping of seismic reflection and well data yield the structure *at the time of rifting*. The crustal structure determined seismically will not necessarily be the same as that deduced from backstripping, even when account is taken of sediment loading. This is because the crustal structure may be modified prior to, during or following a rifting event (e.g. Fig. 1). For example, magmatic underplating (e.g. Cox 1980) may add crustal material to the base of the crust following rifting, while 'tectonic erosion', due for example, to a previous rifting event or excessive heating during rifting, could remove it.

In rift-type basins, where both seismic and well data exist, the crustal structure may be compared with that predicted by backstripping, crustal restoration and sediment loading. By comparing the two types of crustal structure it should be possible to evaluate whether the crust beneath a particular basin has the 'correct' configuration for its rifting history, or whether it has been modified prior to, during or following rifting.

Unfortunately, with the exception of the North Sea, the NW European margin and the Baltimore Canyon Trough, there are only a few localities where sufficient data coverage exist to enable a detailed comparison between the different types of crustal structure to be made. Barton & Wood (1984) compared the depth to the Moho based on seismic refraction data to that expected from backstripping of well data across the North Sea, and concluded that a uniform extension model (McKenzie 1978), when combined with sediment loading of a relatively weak lithosphere, explains the present day seismic structure, *provided* that a pre-Permian stretching event was included prior to mid-Jurassic to early Cretaceous rifting. Similarly, Watts (1988) found using data from the LASE (1984) experiment that the refraction Moho beneath the Baltimore Canyon Trough (east coast, US) was in the expected position given its rifting and loading history. In contrast, Dyment (1990) has argued, using the BIRPS data set, that the reflection Moho beneath the Celtic basin is deeper than the Moho deduced from backstripping. He concluded that the seismic Moho may be quite young due either to ductile flow of the lower crust or the addition of continental material to the base of the crust following rifting.

The Valencia trough is a small basin of Neogene age in the western Mediterranean. There are now large amounts of geological and geophysical data available (e.g. Lanaja 1987) from the trough, due mainly to an intense program of exploration by the oil and gas industry over the past decade. Previous seismic, gravity, heat flow and well data studies (e.g. Watts *et al.* 1990; Torné *et al.* 1992; Foucher *et al.* 1992; Maillard *et al.* 1992;

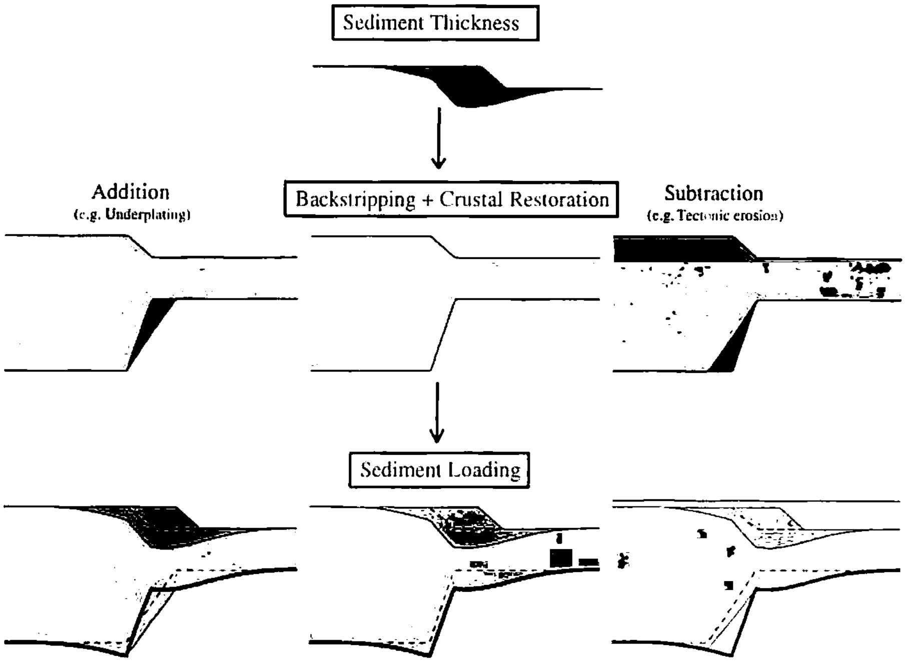

Fig. 1. Schematic diagram comparing the expected position of the Moho (heavy solid line) based on a simple model of crustal thinning and sediment loading (centre column) to the Moho position expected if material had been added to the crust (left-hand column) or subtracted from it (right-hand column) prior to, during or following rifting. The method normally used to compute the simple model (which involves compiling the sediment thickness, backstripping, crustal restoration and sediment loading) is also shown in the central column.

Watts & Torné, in press) indicate that the central part of the trough is underlain by continental crust that has been heated and thinned by as much as a factor of 3. Plate kinematic studies (e.g. Dewey *et al.* 1989) suggest that the extension developed within a region of overall convergence between the Eurasian and African plates, possibly behind a now extinct island arc–trench system. It is therefore of geodynamic interest to examine whether the extensional processes that operated in the trough resemble in any way those of normal Atlantic-type continental margin basins.

The purpose of this paper is to use seismic, well and gravity data to understand better the structure and origin of the Valencia trough. In particular, 3-D flexural backstripping techniques will be used to restore the crustal structure that would be expected based on the tectonic subsidence/uplift and sediment loading history of the trough. By comparing the predicted crustal structure to that measured seismically we aim to improve our understanding of (a) the extent to which a simple rifting and loading model can explain extension in the basin, (b) the extent to which the crust beneath the basin may have been modified since rifting, and (c) the long-term thermal and mechanical properties of stretched continental lithosphere.

Seismic Moho

The principal seismic refraction studies that have been carried out of the Valencia trough region are summarized in Fig. 2a. These studies are based mainly in land areas on Banda *et al.* (1980), Zeyen *et al.* (1985) and the ECORS group (e.g. Bois *et al.* 1990) and in sea areas on the results of the VALSIS group (e.g. Watts *et al.* 1990).

The continental crust beneath the northwestern border of the Iberian peninsula and the Balearic promontory is characterized by seismic P-wave velocities in the range 6.1 to 6.9 km s^{-1} (Gallart *et al.* 1990; Zeyen *et al.* 1985; Banda *et al.* 1980).

Along the Ebro Platform (Fig. 2a) and the central part of the Valencia trough the Neogene and Quaternary sediments are characterized by a succession of layers with velocities of 1.6–4.8 km s^{-1}. Beneath the flanks of the trough and possibly the axis, a 5.1–5.9 km s^{-1} velocity layer has been identified which probably represents pre-rift Mesozoic sediments. The upper and middle crust are characterized by velocities which increase gradually from about 6.0–6.4 km s^{-1} in the upper crust to about 6.4–7.0 km s^{-1} in the lower crust. Both the flank and trough axis are underlain by anomalously low P_n velocities that range from 7.2–7.9 km s^{-1}.

Towards the NE and SW extremities of the trough, the velocity–depth structure shows some marked differences from the central region. In the NE, sediment velocities are similar to those described above. The crustal velocity structure, however, can be characterized by 3 velocity-gradient layers ranging from 4.8–6.1, 6.1–6.7 to 6.7–7.2 km s^{-1}, which resembles that of normal oceanic crust (Pascal *et al.* 1992; Torné *et al.* 1992). Below these layers a 1.5 km thick Moho with velocities in the range from 7.2–7.9 km s^{-1} marks the transition into the uppermost mantle. In the SW, dipping reflections are observed in the upper crust which suggest a locally thick Mesozoic sequence. The base of this reflective region is marked by a bright reflector with a gentle northwestward slope that soles out at a depth of about 11 km. The deep reflector may correspond either to the base of the Mesozoic basin or to a detachment surface that cuts across the upper–middle crust.

The seismic data indicate that the crust thins from about 30–35 km beneath the Iberian peninsula to 20–25 km beneath the Valencia shelf and to about 15 km beneath the centre of the trough (Zeyen *et al.* 1985; Gallart *et al.* 1990; Banda *et al.* 1980; Pascal *et al.* 1992; Torné *et al.* 1992). Gravity and geoid modelling (Watts *et al.* 1990; Watts & Torné in press) indicate that the transition between unstretched continental crust and stretched crust takes place over a horizontal distance of about

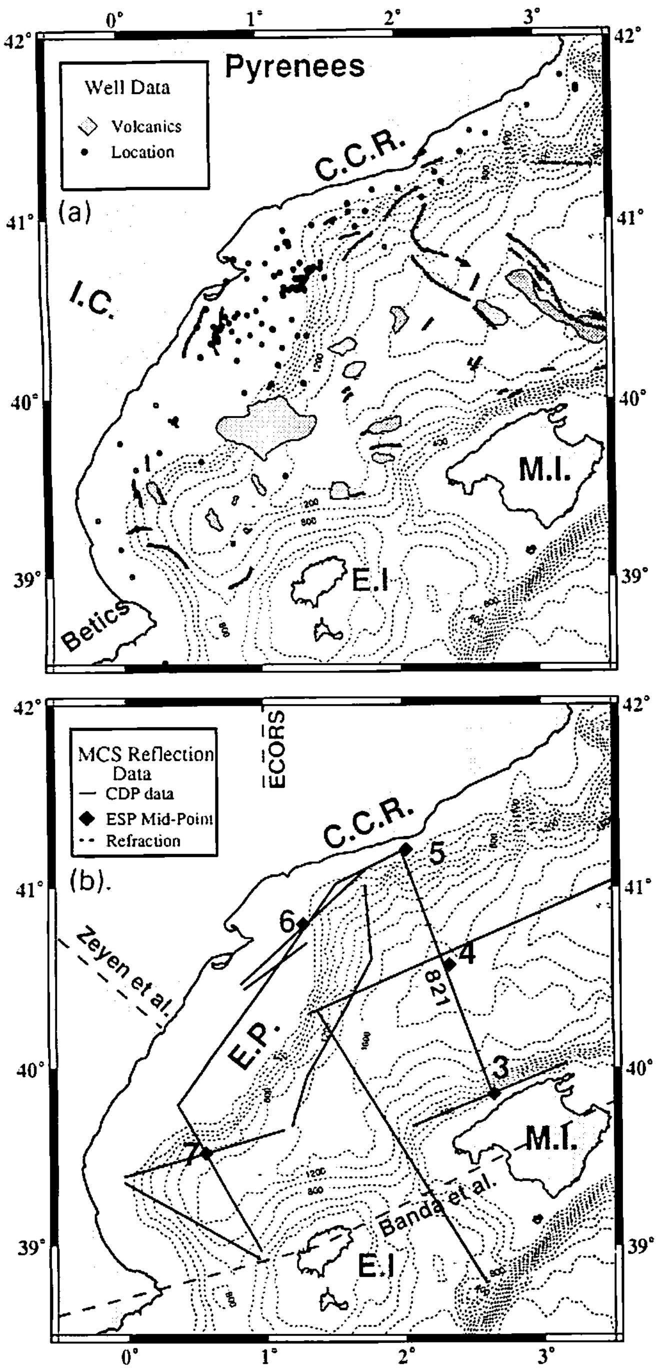

Fig. 2. Summary map showing the distribution of (a) seismic reflection (solid lines) and refraction (solid diamonds and dashed lines) data and (b) well data. The shaded regions show the distribution of volcanic material and the heavy lines are faults (after Maillard *et al.* 1992). The bathymetry is shown at 200 m intervals and is based on the IOC chart (1988). I.C., Iberian Chain; C.C.R., Catalan Coastal Ranges; M.I., Mallorca Island; E.P., Ebro Platform; E.I., Eivissa Island.

50 km. This makes the transition beneath the Iberian margin one of the steepest to have been reported from a passive continental margin and, as will be shown later, it is difficult to explain in terms of current extensional models. Beneath the Mallorca margin the transition is of more typical width (100–200 km), but this has been attributed (Watts & Torné in press) to the formation of the margin by orogenic loading of weak, thin (stretched ?) crust in the Balearic promontory.

Sediment thickness

Detailed maps of the thickness of the Neogene beneath the Valencia trough have been published by IGME (Lanaja 1987), based on industry well and seismic reflection profile data sets. Maillard *et al.* (1992) have modified the IGME map so as to include the seismic reflection profile data obtained by the VALSIS group (e.g. Fig. 2). Their map, which extends from the Barcelona graben in the north to the Mallorca margin in the south, did not include the shelf regions mapped previously by IGME.

Figure 3 is a compilation of the depth to the base of the Neogene sediments which was obtained by combining the IGME and Maillard *et al.* (1992) data sets with additional data from the Neogene basins of the Catalan Coastal Ranges. The map shows a generally symmetric distribution of sediments in the Valencia trough with a rapid thickening from the Iberian peninsula and the Balearic promontory towards the trough centre. Maximum depths (> 4.5 km) are reached in the extreme NE (Fig. 3), but large depths are also encountered in the axis of the trough and beneath its SW extremity. The rapid thickening in the Neogene generally correlates with the present day outer shelf break in slope, except in the region of the Ebro delta where it is located close to the inner shelf.

Backstripping

The main factors that are responsible for the subsidence of extensional basins are crustal thinning at the time of rifting and sedimentary loading. Various mechanisms have been proposed to explain the origin of the thinning. One possibility is that it occurs by plastic necking in response to extension as suggested by Artemjev & Artyushkov (1971) and applied by McKenzie (1978) to explain stretching of the crust and lithosphere. Other mechanisms involve extension of the mantle over a broader area than the crust (Rowley & Sahagian 1986; White & McKenzie 1988) and creep due to lower crustal flow (Bott 1971).

In McKenzie's (1978) model, there is an initial subsidence, due to crustal thinning followed by a thermal subsidence as the lithosphere cools. The crustal thinning is calculated assuming a Pratt-type model of isostasy in which a column of stretched crust is balanced with an unstretched column (e.g. Cochran 1981) which, in turn, is balanced with a mid-ocean ridge. The mid-ocean ridge consists of a 2.5 km water column, a 5 km thick oceanic crust and a 125–7.5 = 117.5 km thick mantle with densities for water, crust and mantle of 1030, 2800 and 3180 kg m^{-3} respectively. By column balancing it is possible, using the McKenzie (1978) model, to compute the water depth that results from a particular amount of stretching. Alternatively, if the water depth is given, say by backstripping, then the amount of thinning can be deduced.

In practice, of course, subsiding basins are filled to some extent by sediments. At the east coast of the USA for example, the margin has built upward by several kilometres and outwards by a few hundred kilometres over extended continental crust so the present water depth gives little indication of how thick the crust is. Only in 'starved' basins such as the Bay of Biscay margin will the present-day water depth give a reasonably accurate guide to the amount of thinning.

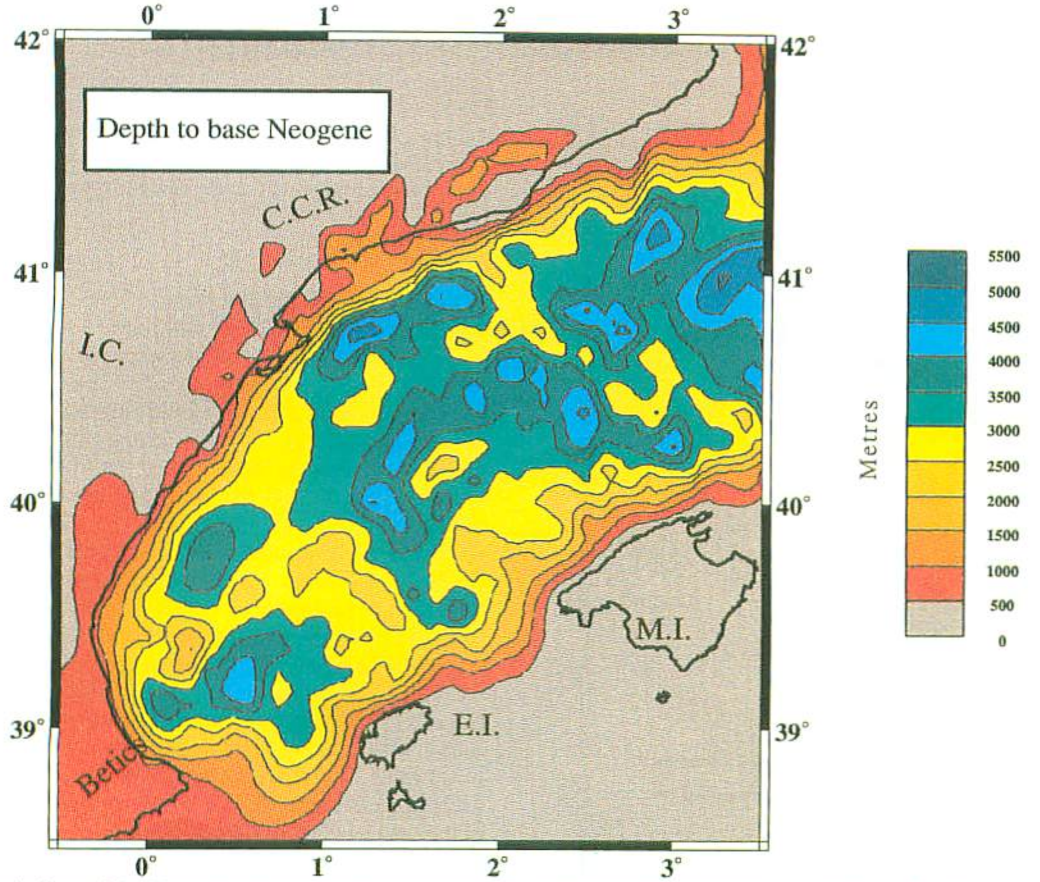

Fig. 3. Depth to the base of the Neogene based on seismic and well data. Modified from IGME (Lanaja 1987) and Maillard *et al.* (1992).

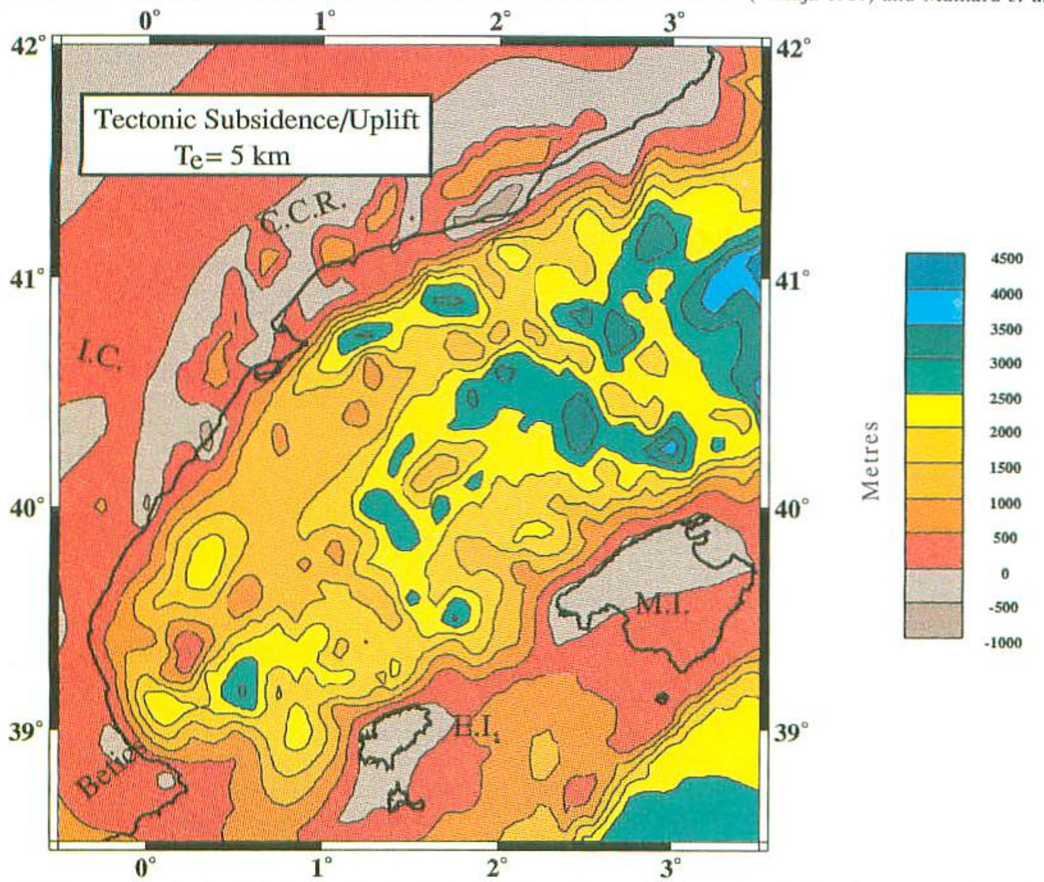

Fig. 5. Tectonic subsidence/uplift obtained by flexural backstripping of the sediment thickness data in Fig. 2. A T_e = 5 km and densities of the water, sediment and mantle of 1030, 2400 and 3300 kg m^{-3} respectively have been assumed. The map shows a broad region of subsidence associated with the trough which is flanked by a broad region of uplift.

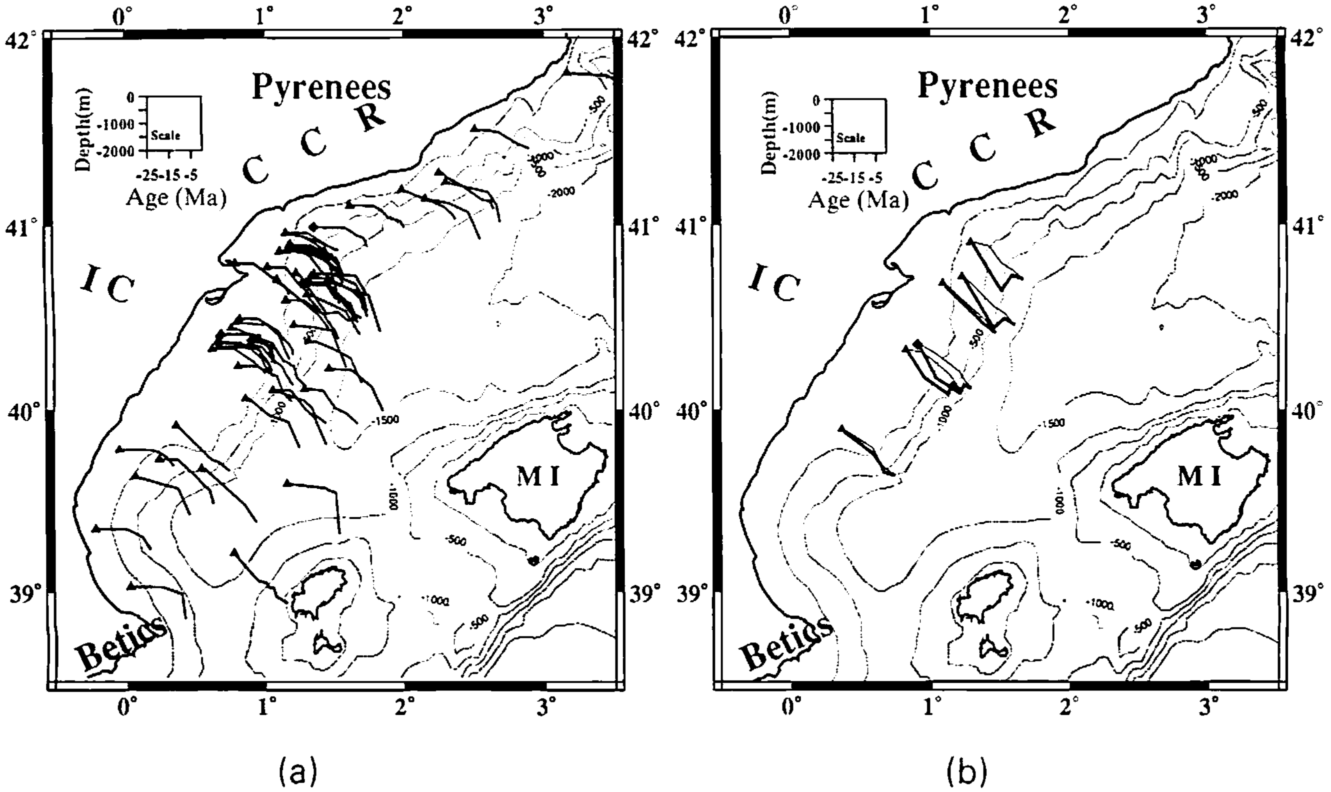

Fig. 4. Tectonic subsidence obtained by backstripping the stratigraphic data at the wells located in Fig. 1b. (a) No water depth corrections, (b) with water depth correction (heavy line, with water depth correction; fine line, without water depth correction). The stratigraphical data are from Lanaja (1987) and the water depth correction is based on unpublished commercial data.

Backstripping (e.g. Watts & Ryan 1976) is a technique that attempts to correct the basement depth for the effects of sediment and water loading. A number of studies have shown that when account is made of compaction, the water depth of deposition and sea-level, this technique provides a reliable estimate of the tectonic subsidence/uplift, even in sediment-filled basins. The different types of backstripping may be distinguished according to the way they treat the response of the lithosphere to loading.

1-D Airy-type

Airy-type backstripping is the most commonly used technique at wells where 'down-hole' information is available on porosity, lithology, biostratigraphy and water depths of deposition. Several studies (e.g. Sclater & Christie 1980) have shown that it is a satisfactory technique to use in order to estimate the amount of thinning, to determine the duration of initial rifting in young margins and to discriminate between the different rift models (e.g. instantaneous or finite rifting).

One problem with an Airy approach is that the estimate of tectonic subsidence/uplift is limited to the immediate vicinity of a well site and, in the case of commercial wells, they are often located on structural highs. In the Valencia trough (Fig. 4a) the general pattern of the tectonic subsidence/uplift seems to be similar at the different wells suggesting that local effects have probably been smoothed out. Unfortunately, there are only a few wells (Fig. 4b) where there is adequate control on the compaction and water depth. Studies of these wells (e.g. Watts *et al.* 1990) have shown that there is a gentle exponential decrease in the tectonic subsidence/uplift with time since rifting. The gentle decrease can be explained by a model (Cochran 1983) in which there was an initial rifting period, about 6 Ma in duration, followed by a thermal subsidence.

A more difficult problem, from the viewpoint of crustal structure, is that the wells are limited in their distribution to the shelf and upper slope of the Ebro platform and Catalan margin (Fig. 4). They therefore do not provide any information on the tectonic subsidence/uplift of the main part of the trough.

2-D flexural type

A better approach, if the tectonic subsidence/uplift is required over a broad area, is to use flexural, rather than Airy-type, backstripping techniques. Flexural backstripping, which has been applied to a number of basins, including the North Sea (Thorne & Watts 1989), east coast, US (Watts 1988) and Gulf of Lyons (Bessis 1986), is particularly well suited to seismic reflection profiles which can be progressively unloaded layer by layer.

The problem with flexural backstripping is that it requires knowledge of the long-term mechanical properties of the lithosphere and how these properties vary with plate and load age. While it has been shown that a simple relationship exists between the age of the lithosphere at the time of loading and the effective elastic thickness, T_e, of oceanic lithosphere it is not clear at present whether such a simple relationship also holds for the continents. Studies of the basement geometry beneath foreland basins (e.g. McNutt *et al.* 1988) suggests that the continental lithosphere is capable of showing both great strength ($80 < T_e < 100$ km), where the cold cratonic interiors are involved and great weakness ($0 < T_e < 25$ km) where extended continental crust is present.

Watts & Torné (in press) flexurally backstripped a seismic reflection profile across the Valencia trough (Line 821, Fig. 2a) and showed that irrespective of T_e, the tectonic subsidence/uplift is associated with a broad region of subsidence that is flanked by a flexural uplift. The backstripped profile resembles quite closely the topography of present day rifts in East Africa, the main differences being that in the trough the width of the extension is much greater and the amplitude of the tectonic subsidence is higher.

3-D flexural type

In basins where the sediment distribution has been well mapped, the flexural backstripping technique may be extended to 3-dimensions.

One of the first applications of this technique was by Thorne & Watts (1989) to sediment thickness data from the North Sea basin. They used the data to map the tectonic subsidence/uplift for different stages of basin development. The maps documented the regions of relative subsidence and uplift for particular time intervals as well as those parts of the basin which had undergone subsidence followed by uplift, due to inversion tectonics.

The Valencia trough, like the North Sea basin, is characterized by a large amount of high-quality well and seismic data. We have therefore backstripped the sediment thickness data for the trough (Fig. 5) using a similar technique as used by Thorne & Watts (1989). In particular, the tectonic subsidence/uplift, TSU, was computed by taking the inverse Fourier Transform of:

$$\mathrm{TSU}\ (k) = \left[\frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)}\right] \mathrm{S}\ (k)\phi_e(k)$$

where S (k) is the discrete Fourier Transform of a sediment thickness grid, $k = (k_x + k_y)^{1/2}$ where k_x is the wavenumber along the x axis and k_y the wavenumber along the y axis, ρ_m, ρ_s and ρ_w are the densities of the mantle, sediment and water respectively. The wavenumber parameter, $\phi_e(k)$, is given by:

$$\phi_e(k) = \left[\frac{Dk^4}{(\rho_m - \rho_w)g} + 1\right]^{-1}$$

where g is the average acceleration due to gravity and D the flexural rigidity of the plate. D is determined from the elastic thickness, T_e, by:

$$D = \left[\frac{ET_e^3}{12(1 - \sigma^2)}\right]$$

where σ = Poissons ratio and E = Young's Modulus.

The first step in the data preparation was to smooth the depth to base of the Neogene and the bathymetry in 2.5 × 2.5 minute 'blocks' and then transform each smoothed block on to a 5 × 5 minute grid using a minimum curvature technique (Smith & Wessel 1990). The two data sets were then subtracted from each other to obtain the sediment thickness grid. Finally, the *total* TSU was obtained by adding the backstrip obtained from the sediment thickness grid to the present-day water depth (i.e. the unfilled part) of the basin.

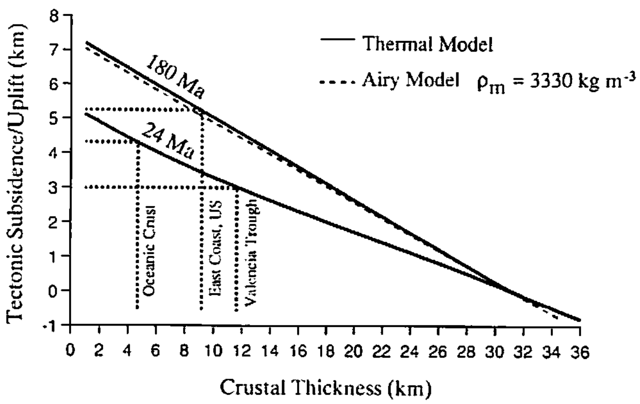

Fig. 6. Relationship between crustal thickness, H and tectonic subsidence/uplift, TSU, for an Airy (dashed line) and thermal model (solid line) of isostasy. The thick dashed lines show the crustal thickness and tectonic subsidence/uplift expected in different tectonic settings. *Oceanic Crust*, the tectonic subsidence/uplift expected for 5 km thick 24 Ma old oceanic crust; *East Coast, US*, the tectonic subsidence/uplift expected for 9 km thick 180 Ma old stretched continental crust; *Valencia Trough*, the tectonic subsidence/uplift expected for 11.5 km thick 24 Ma old stretched continental crust. The crustal thickness in these different tectonic settings is based on available seismic refraction data.

Figure 5 shows a map of the total TSU for the Valencia trough based on T_e = 5 km. The map reveals a broad region of subsidence associated with the Valencia trough that is flanked by an uplift in the Catalan Coastal Ranges and Mallorca. The tectonic subsidence is a maximum in the NE part of the trough where it reaches depths of about 4.2–4.4 km. These depths are similar to those that would be expected for a 5 km thickness of 24 Ma old oceanic crust (Fig. 6). An oceanic origin for this part of the trough is supported by Pascal *et al.* (1992) who showed that the velocity–depth profile at an expanding spread profile (ESP) in the region (ESP 2, Pascal *et al.* 1992) closely resembles that of normal oceanic crust. Elsewhere, the tectonic subsidence reaches values of 3.0–4.0 km which are similar to that expected if the continental crust beneath the trough had been stretched by up to a factor of 3.

Superimposed on the broad pattern of tectonic subsidence/uplift (Fig. 5) are a number of 'local' features which include narrow NNE–SSW-trending zones of relative subsidence and uplift between Barcelona and Mallorca. These features may represent a transform fault complex that developed in response to the opening of the Valencia trough as described by Maillard *et al.* (1992). Other short-wavelength features in Fig. 5 are probably also related to the process of rifting. These include the development of tilted blocks (relative highs) and constructional volcanic centres that flexurally load the crust (relative highs and lows).

Backstrip and flexed Moho

The backstrip Moho

In this paper, the backstrip Moho is defined as the base of the restored crustal structure in the absence of sediment loading. Because sediment loading is one of the main processes that

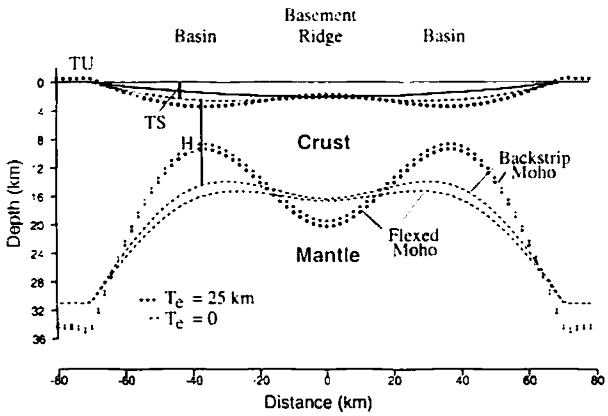

Fig. 7. Schematic diagram showing the backstrip Moho and the flexed Moho for a theoretical basin with two depocentres separated by a basement ridge. The backstrip and flexed Moho were obtained by 2-D flexural backstripping, crustal restoration and thermal modelling. H, thickness of stretched crust; TS, tectonic subsidence; TU, tectonic uplift. The backstrip profiles are based on $T_e = 0$ (dashed lines) and 25 km (solid dots), $\rho_s = 2400$ kg m^{-3}, $\rho_m = 3220$ kg m^{-3} and $\rho_w = 1030$ kg m^{-3}. Note that as T_e is increased, the predicted Moho depths decrease beneath the basins, but increase in the region of the basement ridge.

modifies the post-rift crustal structure in extensional basins, we believe that the backstrip Moho reflects, in the absence of other modifying effects, the crustal structure *at the time of rifting*.

The depth to the backstrip Moho, BM, can be computed directly from TSU provided some assumption is made about isostasy. If an Airy model is assumed, for example, then it is easy to show that:

$$H = T_c - \mathrm{TSU}\left[\frac{(\rho_m - \rho_w)}{(\rho_m - \rho_c)}\right]$$

where T_c = initial crustal thickness, H = crustal thickness and ρ_w, ρ_c and ρ_m are the densities of the water, crust and mantle respectively. Also, since BM = H + TSU we can write,

$$\mathrm{BM} = T_c - \mathrm{TSU}\left[\frac{(\rho_m - \rho_w)}{(\rho_m - \rho_c)}\right]$$

One problem with an Airy model is that it does not take into account that ρ_m is changing as the mantle cools following rifting. To incorporate this effect we need to consider a thermal model of isostasy.

Figure 6 shows the relationship between H and TSU based on the McKenzie (1978) thermal model. The model has been evaluated for two ages since rifting: one corresponding to the Valencia trough (age of initial rifting 24 Ma) and the other to the east coast of the US (180 Ma). Also shown for comparison is the Airy model with a 'normal' density mantle of $\rho_m = 3330$ kg m^{-3}. The two models assume similar densities for the water and crust (i.e. $\rho_w = 1030$ kg m^{-3}, $\rho_c = 2800$ kg m^{-3}), but the thermal model incorporates a mantle density that continually changes as the lithosphere cools following rifting. Both models show a decrease in H with an increase in TSU. The Airy model describes the relationship between H and TSU at the relatively old east coast (USA) margin well, but is a poor description for the young Valencia trough. For this reason, we have assumed a thermal model for the purpose of computing the backstrip Moho in the Valencia trough.

An example of the backstrip Moho is illustrated in Fig. 7 for a basin model with two depocentres separated by a basement ridge. The sediments were flexurally backstripped and the resulting TSU used to compute the crustal thickness and, hence, the backstrip Moho. A basin age of 24 Ma was assumed in the calculations. The depth to the backstrip Moho decreases from 31.2 km beneath the flanks (i.e. the thickness of unstretched crust at sea-level) to about 14 km beneath the centre of each basin. Figure 7 contrasts the depth to the backstrip Moho for a high (T_e = 25 km) and low (T_e = 0 km) strength lithosphere. In the centre of the basin the backstrip Moho based on T_e = 0 km is at a greater depth than the Moho with T_e = 25 km but, at the basement ridge and the depocentre margins the reverse case applies.

The flexed Moho

The flexed Moho (Fig. 7) is the backstrip Moho modified for the effects of sediment loading. For purposes of the calculations, we consider that an individual sedimentary unit is made up of two parts: a load and a flexure. The load is given by the difference between TSU and the present-day water depth (the load is positive if TSU exceeds the present-day water depth and negative otherwise) while the flexure is the deformation of the crust and lithosphere caused by the load (the flexure is downward in the immediate vicinity of a positive load and upward for a negative load).

We point out here that there is no contradiction in our assumption that local compensation models such as Airy or Pratt are used to infer the geometry of the crust at the time of rifting, while a regional scheme of isostasy (i.e. flexure) may apply during the *post-rift* development of a basin. This is because the response of the lithosphere to sediment loading depends on the thermal–mechanical properties of the lithosphere and not its structure (or composition). The use of local models of compensation imply that the lithosphere has little or no strength at the time of rifting, yet the flexural strength of extended lithosphere is presently quite controversial. For example, some authors (e.g. Fowler & McKenzie 1989) argue from sediment-starved passive margins, that T_e during rifting is quite small (about 5 km) while others (e.g. Weissel & Karner 1989) from studies of gravity data from the East African rift suggest quite high values (> 25 km).

The flexed Moho is deeper than the backstrip Moho (e.g. Fig. 7) because of sediment loading. The difference depends on the long-term mechanical properties of the lithosphere with a weak lithosphere giving rise to a large difference and a strong lithosphere a small one.

Comparison of the seismic and flexed Moho

If no other processes are operating, then the flexed Moho should be comparable to the Moho as defined by seismic refraction studies.

A comparison between the seismic and flexed Moho in the Valencia trough is shown in Fig. 8. Before the two crustal structures can be compared in detail, however, it is first necessary to examine some of the uncertainties that might arise in the flexed Moho calculations.

The most important factor to affect the calculated Moho depths are T_e and the densities of the sediment, water and

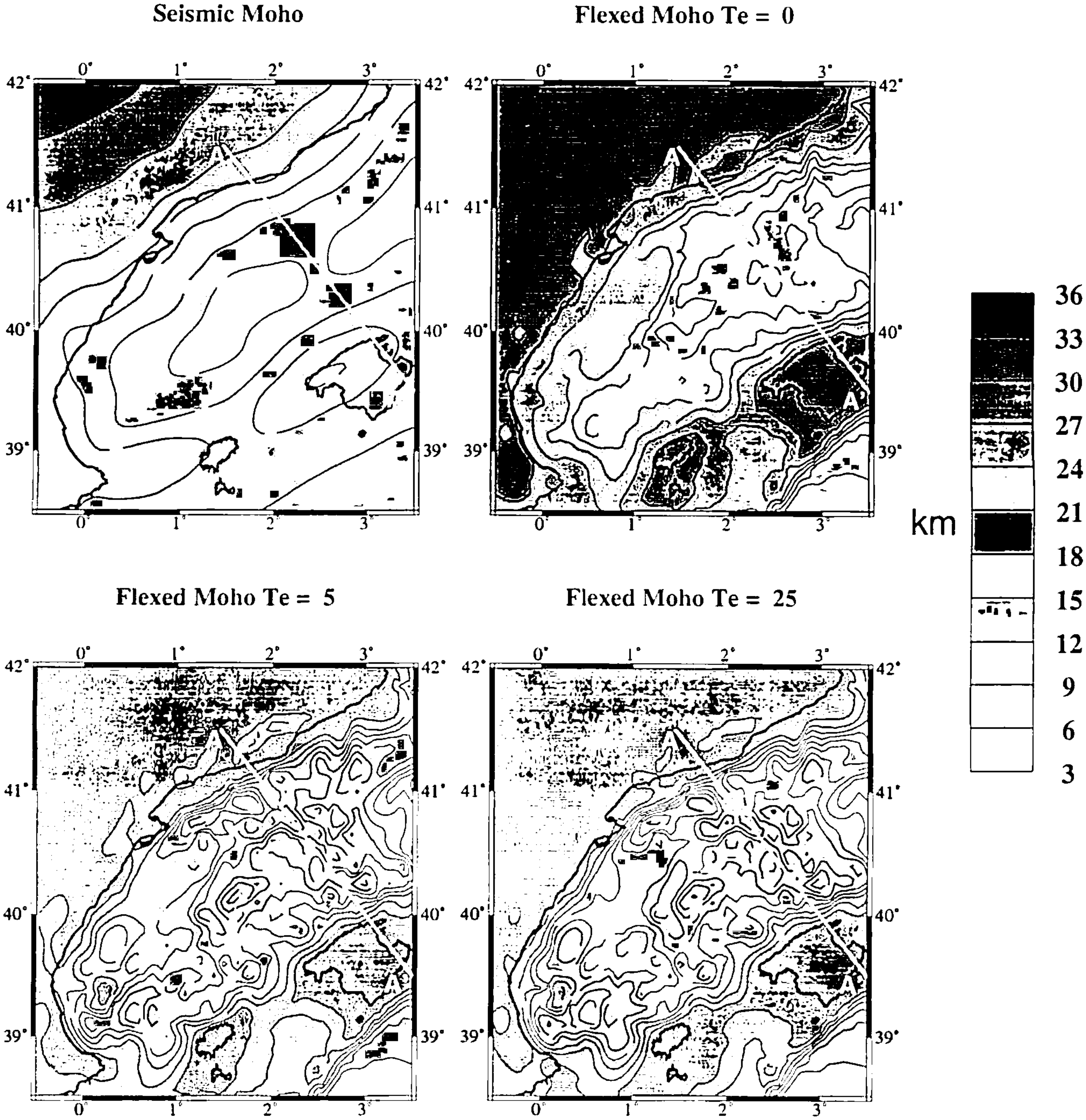

Fig. 8. Comparison of the Seismic Moho and Flexed Moho for different values of T_e. (a) Seismic Moho, (b) flexed Moho with T_e = 0 km, (c) flexed Moho with T_e = 5 km and (d) flexed Moho with T_e = 25 km. The best overall fit between the seismic and flexed Moho is in (b).

mantle. We have therefore carried out a sensitivity analysis (Fig. 9) so as to access the effects of varying these parameters within permissible ranges.

The centre of Fig. 9 shows a 'reference' crustal cross-section that was obtained using T_e = 5 km, ρ_m = 3220 kg m^{-3} and ρ_w = 1030 kg m^{-3} (i.e the same parameters as were used to construct Fig. 5). The flexed Moho was calculated using a load density of ρ_s = 2400 kg m^{-3}. The likely ranges of ρ_m are 3180 to 3330 kg m^{-3} which correspond to the 0 to 1330 °C temperature of mantle materials respectively. Figure 9 shows that the back-strip and flexed Moho depths are not very sensitive to ρ_m and uncertainties that arise from this factor are generally < 1 km. The largest uncertainties are associated with changes in ρ_s and T_e. In particular, increasing T_e and ρ_s has similar but, opposing effects. This is because increasing T_e decreases the capacity of the sediments to load while increasing ρ_s increases it. If ρ_s and T_e are varied within permissible limits then differences in flexed Moho depths of up to 5 km can arise (Fig. 9).

Figure 9 shows that despite a 'trade-off' between ρ_s and T_e, the best fit between the seismically deduced crustal structure and the predicted one, is for the model with T_e = 0 km, ρ_m = 3220 kg m^{-3} and ρ_s = 2400 kg m^{-3}. All other models produce

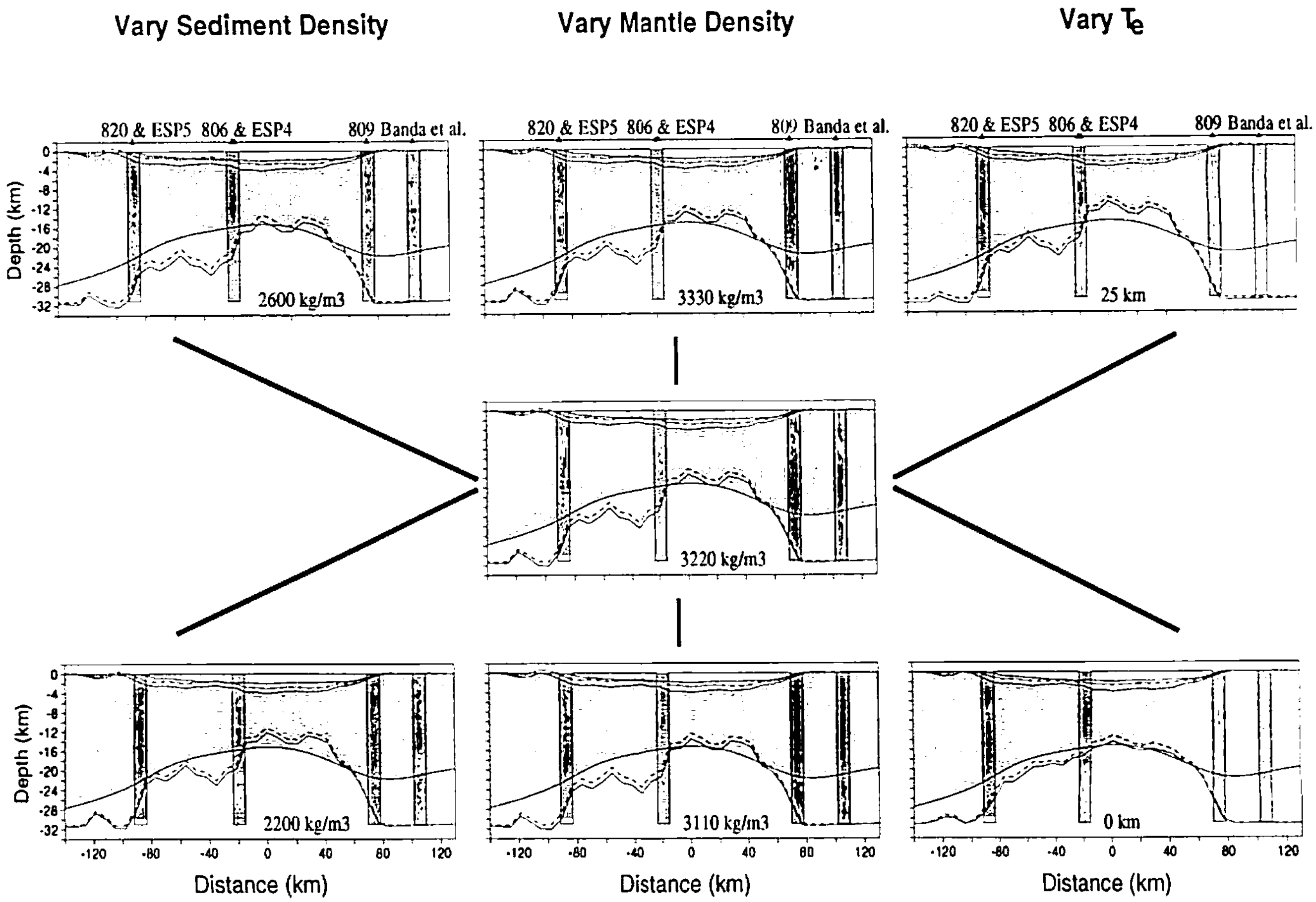

Fig. 9. Comparison of the seismic Moho (base of shaded crustal region), backstrip Moho (dashed line) and flexed Moho (fine solid line) along Profile AA (Fig. 8) for different values of T_e, ρ_s and, ρ_m. The figure shows the sensitivity of the Moho depths to changes in the density when T_e is varied within permissible limits. The central panel shows a 'reference' cross-section which is based on the same parameters as were used in Fig. 5 and $\rho_s = 2400$ kg m^{-3}. The vertical columns with heavy shading show the seismic refraction 'mid-points'. Lightly shaded region represents sediments. The dashed line within the sediments is the tectonic subsidence/uplift obtained by backstripping.

variations in the depth to the Moho that are not observed in the seismic data. We caution, however, that even though the seismic constraints are shown as a smoothed line, the data is actually limited to refraction 'mid-points' and therefore only samples the long-wavelength components of the Moho relief. For example, models with $T_e = 0$ and $T_e = 25$ km produce similar fits to the seismic data at the ESP 4 and 5 mid-points (shaded columns in Fig. 9), but they differ in between.

We prefer the model with $T_e = 0$ because it produces the best overall fit to the seismic Moho gridded data set, but if we had indications from seismic refraction data say that between expanding spread profile mid-points the crust was actually quite thick, then the best fit model would need to be revised. For example, if the crust was thicker beneath the high relief feature between expanding spread profiles 4 and 5, then a model with $T_e = 25$ km might be preferred. This is because, as Fig. 7 shows, the region is located over a relative basement high. By way of contrast to the basin depocentres, a higher strength in this region implies a smaller TSU than the Airy case and hence, locally, a thicker crust.

In order to constrain T_e better, we have also examined the gravity anomaly since this parameter has the potential to provide the best information on the continuity of the Moho *between* individual seismic refraction mid-points. Figure 10 shows that there is a close agreement between the computed anomaly of the best fit model in Fig. 9 and the observed sediment-corrected Bouguer anomaly. Residual gravity anomalies (i.e. observed minus calculated) are small and are confined to (a) a large anomaly off the NW coast of Mallorca, which may be due to an intra-crustal density high, (b) a broad high that is well developed in the NE part of the trough and (c) a small high over the outer part of the Ebro delta.

Discussion

We have shown in this paper that sediment loading of a lithosphere that acquires little or no flexural strength following rifting satisfactorily explains the overall crustal structure of the Valencia trough region as deduced from the available seismic refraction, backstripping, thermal modelling and gravity

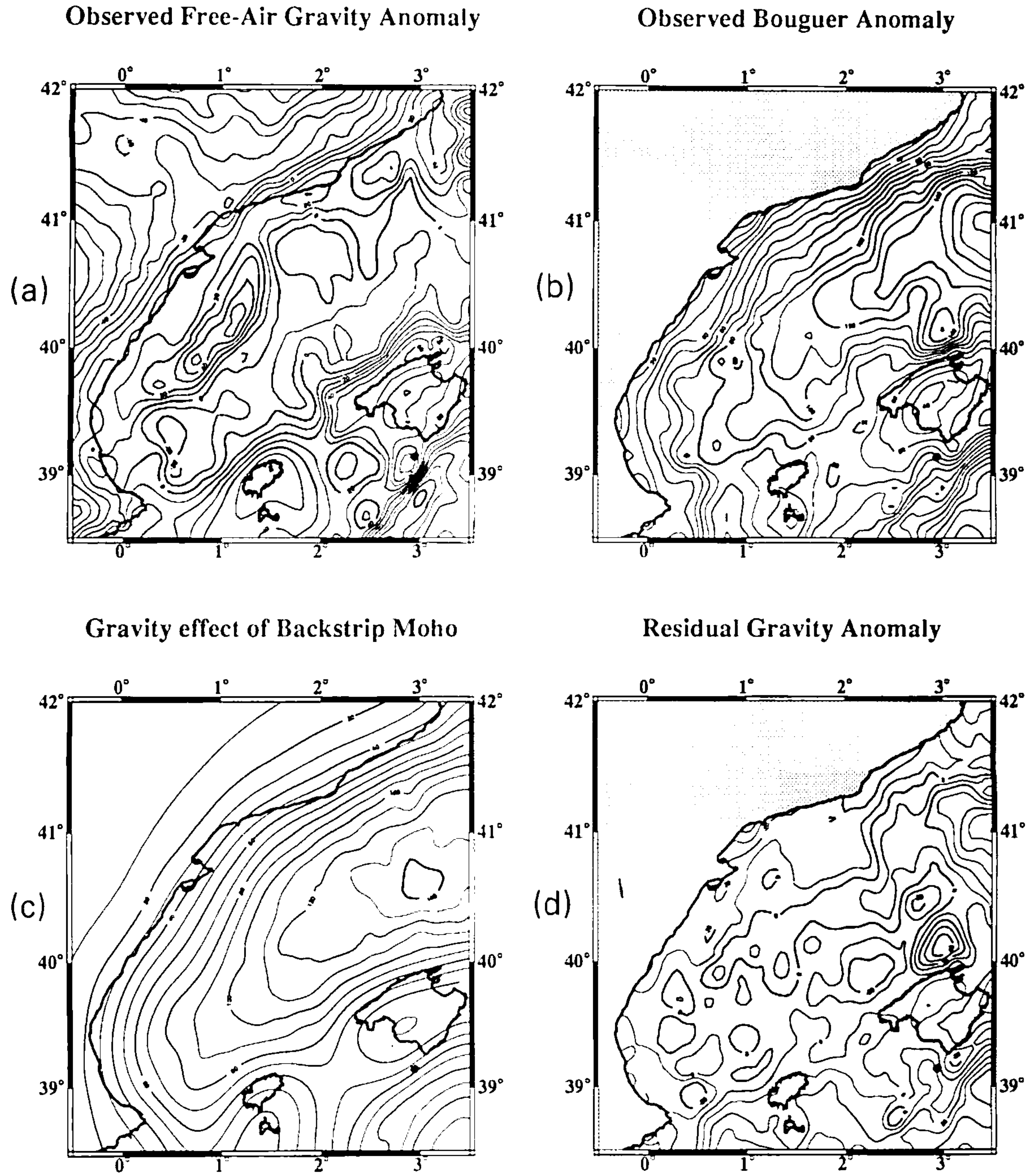

Fig. 10. Comparison of the observed gravity anomaly over the Valencia trough to the calculated anomaly based on backstripping, crustal restoration and sediment loading. (a) Observed free-air gravity anomaly (Bouguer anomaly on land, Free-air anomaly at sea), (b) the Bouguer anomaly at sea corrected for the gravity effect of the sediments, (c) the calculated gravity anomaly with $T_e = 0$, $\rho_s = 2400$ kg m^{-3}, $\rho_m = 3220$ kg m^{-3} and, (d) the residual gravity anomaly obtained by subtracting (b) from (c). The sources for the gravity data are given in Watts & Torné (in press).

anomaly data. The model therefore provides a useful 'reference' with which we can compare the actual crustal structure of the region.

Figure 11 shows a 'residual' Moho depth anomaly map of the Valencia trough that was obtained by subtracting the flexed Moho from the seismic Moho. The flexed Moho in this case is the model that best fits the seismic and gravity anomaly data in Figs 8 and 10. Figure 11 shows that the residual depths are generally small ($\pm$ 2 km) over the central part of the Valencia trough indicating the general applicability of the model to this region.

The largest depth anomalies occur beneath the Catalan Coastal Ranges and the Balearic promontory where the depth of the flexed Moho exceeds the seismic Moho by as much as 8

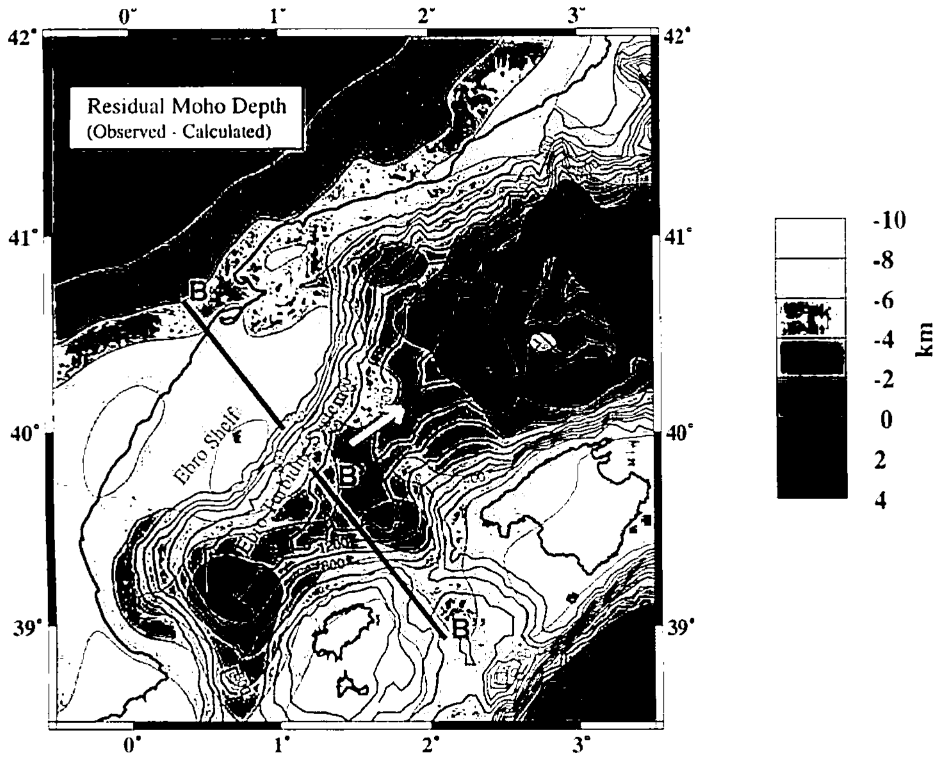

Fig. 11. Residual Moho depth obtained by subtracting the Seismic Moho from the 'best fit' predicted Moho based on backstripping, crustal restoration and thermal modelling. A negative residual indicates a greater than expected depth while a positive one a shallower one. Note that a broad region of negative Moho residual depths is associated with the flanks of the Valencia trough and an intense negative anomaly correlates with the outer part of the Ebro shelf. The location of the Ebro Turbidite System is based on Nelson (1990). The white arrow indicates the net sediment transport direction between the Ebro delta and the Valencia fan.

km (Fig. 11). Watts & Torné (in press) have already noted the 'anomaly' of Mallorca and have attributed the unusually thin crust there to orogenic loading of crust that had already been thinned by a Neogene or earlier heating/rifting event. The depth anomaly over the Catalan Coastal Ranges is more difficult to explain. Watts & Torné (in press) found that the seismic depths were actually greater than the predicted depths (i.e. a positive residual anomaly). This is because they had assumed a crustal thickness of 35 km for the peninsula which was based on extrapolation of data from Banda *et al.* (1983). We have assumed in Fig. 11, however, that Moho depths of this order are only reached in the interior of the Iberian peninsula, and that the Catalan Coastal Ranges are a region that is transitional in crustal thickness between unstretched and stretched continental crust. If this is the case, then the Catalan margin is an example where material has been subtracted from the crust prior to, during, or following rifting. We do not know the mechanism responsible for this, but it may have involved either a previous extension event or excessive heating of the crust and lithosphere.

Positive depth anomalies, indicating a *deeper* Moho than is predicted, correlate with the NE part of the trough. This could be the result (e.g. Fig. 7) of a higher T_e. The associated gravity residual anomaly, however, is also positive which indicates a *shallower* Moho than the predicted one. These apparently divergent results can be explained if magmatic material has been added to the base of the crust, thereby increasing the depth to the Moho, while at the same time a 'dense' component is intruded into the crust which would increase the gravity effect.

The most prominent feature of the depth anomaly map (Fig. 11), however, is an intense negative anomaly (> 10 km) associated with the Ebro delta. The Ebro delta, which is one of the most prominent features of the Catalan margin, has prograded basinward by more than 60 km since the Pliocene. The close correlation between the delta and the depth anomaly is, as Fig. 12 shows, a strong argument that our assumption that stretched crust has a low T_e may not be a correct description of the state of isostasy in this part of the trough.

To examine this question in more detail, we have constructed a profile of the flexed Moho and seismic Moho depth across the Ebro delta and compared it to predicted depths based on a variable rigidity elastic plate model (Fig. 13). A 2-D variable plate model (e.g. Bodine *et al.* 1981) was used since we found no significant difference between the flexure obtained using a uniform rigidity 2-D and a 3-D approach along profile BB". Figure 13 shows that with $T_e = 0$ km the predicted Moho is up to 8 km deeper beneath the delta than is observed. The discrepancy could be reduced by increasing T_e since this would, as Fig. 12 shows, decrease the depth to the flexed Moho beneath the delta. Fig. 13c shows that a model in which the rigidity is low beneath Iberia and the Valencia trough but high beneath the delta can explain the depth to the seismic Moho. The best overall fit beneath the outer part of the shelf is for $T_e = 80$ km which, if correct, is among the highest of values that have been obtained to date from continental flexure studies.

Brunet (1986) and Zoetemeijer *et al.* (1990) have concluded that the Pyrenees foreland basin in France and Spain have a low flexural rigidity ($0 < T_e < 5$ km). The high values beneath the Ebro delta shelf and slope regions cannot therefore be attributed to the fact that the rigidity of the Iberian lithosphere is fundamentally high. We also do not believe that the high

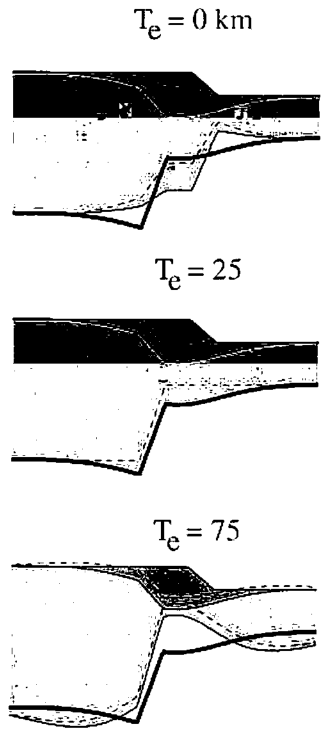

Fig. 12. Schematic diagram comparing the expected position of the Moho (heavy solid line) based on a simple model of crustal thinning and sediment loading with (**middle section**) to the Moho position expected for different assumptions of T_e. The middle section is the same simple model used in Fig. 1 and is based on T_e = 25 km. **Upper Section:** T_e = 0 km. In this case, the depths to Moho beneath the delta would *exceed* the predicted depths based on a simple model and a positive residual Moho depth (using the definitions in Fig. 11) would result. **Lower Section:** T_e = 75 km. In this case, negative residual Moho depths would result. Note that in both cases, the residual depth anomaly maxima will correlate with the region immediately beneath the sediment load.

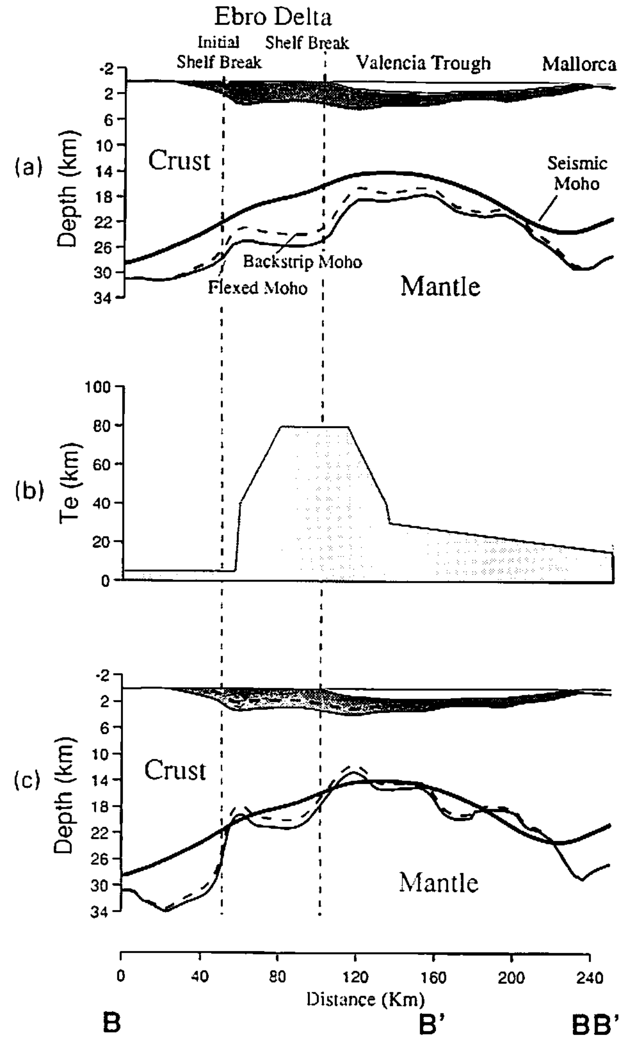

Fig. 13. Profile BB" (Fig. 11) of the Ebro delta. (**a**) Comparison of the Seismic Moho and the Backstrip Moho and Flexed Moho based on the 'best fit' model with T_e = 0, ρ_s = 2400 kg m^{-3} and ρ_m = 3220 kg m^{-3}, (**b**) the distribution of T_e needed to bring the Flexed Moho and the Seismic Moho into agreement and, (**c**) Comparison of the Seismic Moho and the Backstrip Moho and Flexed Moho based on the variation of T_e shown in (b). Note that in order to fit the depth to the Moho a T_e of as high as 80 km is required beneath the delta.

values reflect the fact that extended continental crust has high flexural strength as proposed by Weissel & Karner (1989) for the East Africa rift system, since the Catalan margin and adjacent regions to the Ebro delta are associated with a low T_e.

We believe the high T_e arises because the Ebro delta is a relatively young feature and so there has been insufficient time for the lithosphere beneath the delta to achieve isostatic equilibrium. The extrapolation of data from experimental rock mechanics (e.g. Goetze & Evans 1979) imply that on loading, the lithosphere should 'relax' from its short-term seismic (?) thickness to its long-term elastic thickness. Flexure studies at Hawaii (e.g. Watts & ten Brink 1989) suggest that the relaxation of oceanic lithosphere is essentially complete *within* 1–2 Ma of loading. Although the time over which the continental lithosphere relaxes has been widely debated, with some authors (e.g. Quinlan & Beaumont 1984) favouring a model in which adjustment occurs over tens of millions of years, most authors now agree that the time is short and of the order of a few million years. Since the Ebro delta is a relatively recent feature it is therefore possible that the extended continental lithosphere that underlies it has not had sufficient time to relax.

As would be expected, the lack of adjustment to the Ebro delta is also reflected in the observed free-air gravity anomaly. Figure 10 shows that the delta is associated with a single 'peak' gravity anomaly with an amplitude of 60 mGal and a wavelength of 100 km. The high is strikingly different from the gravity anomaly 'couple' that is typically observed at the world's passive continental margins (e.g. Karner & Watts 1982).

Figure 14 compares the observed gravity anomaly to a calculated anomaly based on flexural backstripping of a variable rigidity plate along profile BB. The gravity anomaly was computed using a similar procedure to that described by Watts (1988) at the US east coast margin in which the effects of rifting and sedimentation are evaluated separately and then summed. The 'rifting anomaly' was calculated using TSU as the upper surface of the crust and the backstrip Moho as the base. In this

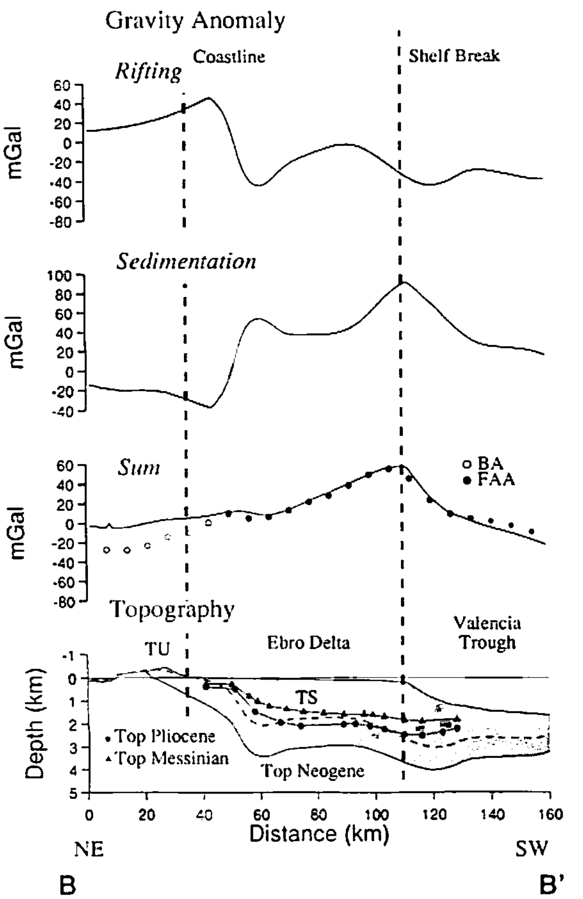

Fig. 14. Comparison of the observed free-air (sea, solid dots) and Bouguer gravity anomaly (land, open circles) with the calculated anomaly based on a high T_e model (Fig. 11). The calculated anomaly (solid line) can be considered (e.g. Watts 1988) as the sum of two parts: the Rifting and the Sedimentation anomaly. The rifting anomaly was calculated from the tectonic subsidence/uplift and the Backstrip Moho obtained by crustal restoration. The sedimentation anomaly was calculated as the combined gravity effect of the sediment load and the flexure at the sediment/basement interface and the Backstrip Moho. The depths to the top of the Pliocene and Messinian are based on data from Nelson (1990).

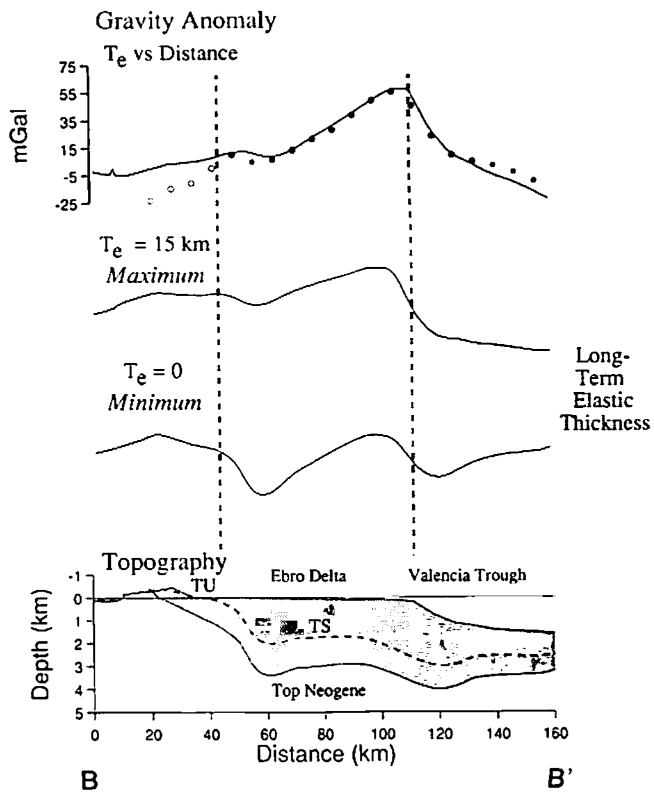

Fig. 15. Comparison of the observed (sea, solid dots; land, open circles) and calculated gravity anomaly expected for the high T_e model (upper profile) to more typical values expected at rift-type basins. Maximum is gravity anomaly for $T_e = 15$ km which is the maximum value expected for 24 Ma old stretched continental crust if T_e follows the depth to the 450 °C oceanic isotherm. Minimum is gravity anomaly for $T_e = 0$ km. This corresponds to the lowest value that might be expected for rift-type basins – which generally seem to correlate with small values. Lower profile: topography, bathymetry and sediment distribution. Thin dashed line is the tectonic subsidence/uplift based on a high T_e model.

case, TSU was based on a variable T_e model (Fig. 13). The 'sedimentation anomaly' was calculated by combining the positive gravity effect of the sediment load (which is calculated from the difference between the TSU and the present-day water depth) and the negative effect of the flexure of the basement and the Moho. Figure 14 shows there is an excellent agreement between the observed and the predicted gravity anomaly based on the high T_e model.

In order to examine the sensitivity of the gravity anomaly to changes in T_e, we have compared the predicted gravity for the high T_e model to two other cases: one where the lithosphere has no flexural strength ($T_e = 0$) and the other where it has acquired a strength similar to that of 24 Ma old oceanic crust ($T_e = 15$ km). Figure 15 shows that neither of these models adequately explains the amplitude and wavelength of the observed gravity anomaly. Furthermore, only a high T_e model explains the amplitude and the steep gradients of the gravity high that are associated with the Ebro delta sediment load.

Figure 13 implies that the mechanically supportive layer beneath the Ebro delta is up to 80 km thick which, while close to the seismic thickness of 24 Ma oceanic lithosphere (e.g. Leeds 1975), is only about half what some authors consider to be the seismic thickness of continental lithosphere as defined by the depth to the low velocity zone (e.g. Anderson 1979). If the seismic thickness of the Iberian peninsula lithosphere exceeds 80 km then these results suggest *some* stress relaxation beneath the delta, but not enough for the thickness of the mechanically supportive layer to reach its long-term elastic thickness. As Fig. 16 shows, the long-term elastic thickness of extended continental lithosphere is generally small. This suggests that a significant amount of adjustment may yet take place before isostatic compensation can be considered complete beneath the Ebro delta.

An important question is the loading history of the Ebro delta since this information might constrain the characteristic time over which stress relaxation in extended continental lithosphere actually occurs. Unfortunately, despite much progress in recent years in better understanding the post-Pliocene development of the Ebro delta, there is still a good deal of uncertainty regarding the age of the delta beneath the outer

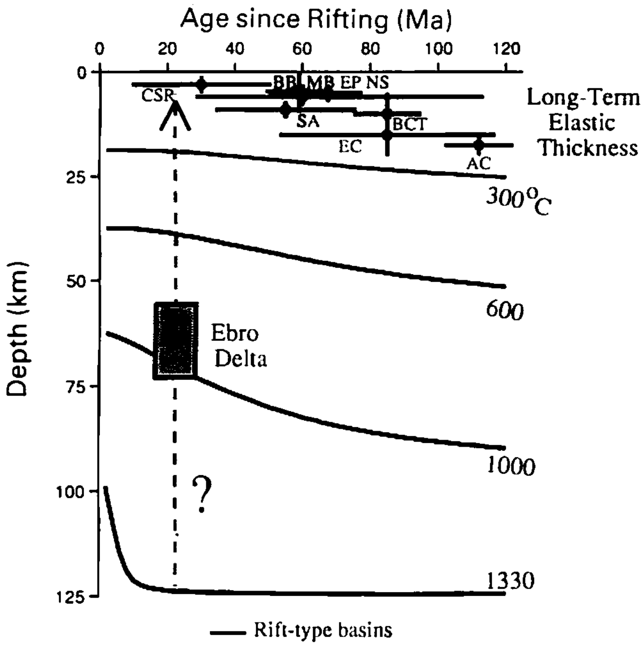

Fig. 16. Plot of T_e v. age of the lithosphere since rifting showing the result from the Ebro delta together the results of studies from other rift-type basins around the world. The age is an estimate of the average age of the basin since rifting and bars reflect the age range that the estimates of T_e *most likely* apply to. The heavy lines show the 300, 600, 1000 and 1330 °C isotherms that would be expected (McKenzie 1978) for continental lithosphere that has been stretched by a factor of 2. We show only the isotherms for comparison purposes and no implication is made that they represent the thermal structure of all the rift-type basins plotted in the figure even though some of them have been stretched by a factor of 2 in part. The Ebro basin value is much higher than previous estimates at rift-type basins (solid dots), and this suggests that an insufficient time has elapsed for the lithosphere beneath the basin to 'relax' from its short-term (?seismic) to its long-term elastic thickness. CSR, Coral Sea Rise (Karner & Watts 1982); SA, South Africa (Karner & Watts 1982); EC, east coast, US (Karner & Watts 1982); AC, Amazon Cone (Nunn & Aires 1988); BCT, Baltimore Canyon Trough (Watts 1988); BB, Bay of Biscay (Diament *et al.* 1986); MB, Michigan basin (Nunn & Sleep 1984); NS, North Sea (Barton & Wood 1984); EP, Exmouth Plateau (Fowler & McKenzie 1989).

part of the shelf and the slope. The data that are available (e.g. Nelson 1990; Nelson & Maldonado 1990) suggest that the depth to the top of the Pliocene corresponds quite well to the backstripped depth based on the variable rigidity model (Fig. 14) indicating that there has been little or no contribution to the tectonic subsidence/uplift since then. This suggests, depending on the time-scale used, that *at least* 1.6 Ma is required for extended continental lithosphere to achieve isostatic equilibrium.

We have not been able therefore in this study to constrain the actual relaxation time of the continental lithosphere. However, the approach that has been followed of comparing the seismically constrained crustal structure to the predicted structure based on simple sediment loading models will, we believe, *eventually* lead to a better understanding of the thermal and mechanical properties of extended continental lithosphere. The acquisition of further seismic reflection and refraction data to define the precise geometry of the transition between stretched and unstretched crust together with the development of more refined backstripping, crustal restoration and sediment loading models offer, we believe, the most promise of addressing this problem in the next decade.

References

Anderson, D. L. 1979. The deep structure of the continents. *Journal of Geophysical Research*, 84, 7555–7560.

Artemjev, M. E. & Artyushkov, E. V. 1971. Structure and isostasy of the Baikal rift and the mechanism of rifting. *Journal of Geophysical Research*, 76, 1197–1211.

Banda, E., Ansorge, J., Boloix, M. & Córdoba, D. 1980. Structure of the Crust and Upper Mantle beneath the Balearic Islands (Western Mediterranean). *Earth and Planetary Science Letters*, 49, 219–230.

——, Udias, A., Mueller, St., Mezcua, J., Bloix, M., Gallart, J. & Aparicio, A. 1983. Crustal structure beneath Spain from deep seismic sounding experiments. *Physics of the Earth and Planetary Interiors*, 31, 277–280.

Barton, P. J. & Wood, R. J. 1984. Tectonic evolution of the North Sea Basin: Crustal stretching and subsidence. *Geophysical Journal of the Royal Astronomical Society*, 79, 98–1022.

Bessis, J. 1986. Some remarks on the study of subsidence of sedimentary basins. Application to the Gulf of Lions margin (Western Mediterranean). *Marine and Petroleum Geology*, 3, 37–63.

Bodine, J. H., Steckler, M. S. & Watts, A. B. 1981. Observations of flexure and the rheology of the oceanic lithosphere. *Journal of Geophysical Research*, 86, 3695–3707.

Bott, M. H. P. 1971. Evolution of young continental margins. *Tectonophysics*, 11, 319–327.

Bois, C., Gariel, O. & Pinet, B. 1990. ECORS Deep Seismic Surveys across Paleozoic Basins in France and Adjacent areas: a Review. *In*: Pinet, B. & Bois, C. (eds) *The Potential of Deep Seismic Profiling for Hydrocarbon Exploration*. Technip, Paris. 381–402.

Brunet, M. F. 1986. The influence of the evolution of the Pyrenees on adjacent basins. *Tectonophysics*, 129, 343–354.

Cochran, J. R. 1981. Simple models of diffuse extension and the pre-seafloor spreading development of the continental margin of the northwestern Gulf of Aden. *Proceedings of the 26th International Cong. Symposium on Continental Margins*. Oceanologica Acta, 154–165.

——1983. Effects of finite extension times on the development of sedimentary basins. *Earth and Planetary Science Letters*, 66, 289–302.

Cox, K. G. 1980. A model for flood basalt volcanism. *Journal of Petrology*, 21, 629–650.

Dewey, J. F., Helman, M. L., Turco, E., Hutton, D. H. W. & Knott, S. D. 1989. Kinematics of the Western Mediterranean. *In*: Coward, M. P., Dietrich, D. & Park R. G. (eds) Geological Society, London, Special Publication, 45, 265–283.

Diament, M., Sibuet, J. C. & Hadaoui, A. 1986. Isostasy of the Northern Bay of Biscay continental margin. *Geophysical Journal of the Royal Astronomical Society*, 86, 893–907.

Dyment, J. 1990. Some complementary approaches to improve the deep seismic reflection studies in sedimentary basin environment: The Celtic sea basin. *In*: Pinet, B. & Bois, C. (eds) *The Potential of Deep Seismic Profiling for Hydrocarbon Exploration*. Technip, Paris, 403–423.

Foucher, J. P., Mauffret, A., Steckler, M., Brunet, M. F., Maillard, A., Rehault, J. P., Alsono, B., Desegaulx, P., Murillas, J. & Ouillon, G. 1992. Heat Flow in the Valencia Trough: Geodynamic Implications. *Tectonophysics*, 203.

Fowler, S. & McKenzie, D. 1989. Gravity studies of the Rockall and Exmouth Plateaux using SEASAT altimetry. *Basin Research*, 2, 27–34.

Gallart, J., Rojas, H., Diaz, J. & Dañobeitia, J. J. 1990. Features of Deep Crustal Structure and the Onshore-Offshore Transition at the Iberian Flank of the Valencia Trough (Western Mediterranean). *In*: Dañobeitia, J. J. & Pinet, B. (eds) *Geophysics of the Mediterranean Basin. Journal of Geodynamics*, 12, 233–252.

Goetze, C. & Evans, B. 1979. Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics. *Geophysical Journal of the Royal Astronomical Society*, 59, 463–478.

Intergovernmental Oceanographic Commission (IOC) 1988. *International Bathymetric Chart of the Mediterranean (IBCM)*. 1:1 million (at 38°N).

Karner, G. D. & Watts, A. B. 1982. On isostasy at Atlantic-type Continental Margins. *Journal of Geophysical Research*, 87, 2923–2948.

Lanaja, M. 1987. *Contribución de la exploración petrolífera al conocimiento de la geología de España*. IGME, Madrid.

LEEDS, A. R. 1975. Lithospheric thickness in the western Pacific. *Physics of the Earth and Planetary Interiors*, 11, 61–64.

MAILLARD, A., MAUFFRET, A., WATTS, A. B., TORNÉ, M., PASCAL, G., BUHL, P. & PINET, B. 1992. Tertiary sedimentary history and structure of the Valencia Trough (Western Mediterranean). *In*: BANDA, E. & SANTANACH, P. (eds) The Valencia Trough: Geology and Geophysics. *Tectonophysics*, 203, 57–76.

MCKENZIE, D. P. 1978. Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters*, 40, 25–32.

MCNUTT, M. M., DIAMENT, M. & KOGAN, M. G. 1988. Variations of elastic plate thickness at continental thrust belts. *Journal of Geophysical Research*, 93, 8825–8838.

NELSON, C. H. 1990. Estimated post-Messinian sediment supply and sedimentation rates on the Ebro continental margin, Spain. *Marine Geology*, 95, 395–418.

——& MALDONADO, A. 1990. Factors controlling late Cenozoic continental margin growth from the Ebro delta to the western Mediterranean deep sea. *Marine Geology*, 95, 419–440.

NUNN, J. A. & AIRES, J. R. 1988. Gravity anomalies and Flexure of the Lithosphere at the middle Amazon Basin, Brazil. *Journal of Geophysical Research*, 93, 415–428.

——& SLEEP, N. H. 1984. Thermal Contraction and flexure of the lithosphere of intra-cratonic basins: A three-dimensional study of the Michigan basin. *Geophysical Journal of the Royal Astronomical Society*, 76, 587–635.

PASCAL, G., TORNÉ, M., BUHL, P., WATTS, A. B. & MAUFFRET, A. 1992. Crustal and Velocity Structure of the Valencia Trough (Western Mediterranean) Part II: Detailed interpretation of 5 Expanded Spread Profiles. *In*: BANDA, E. & SANTANACH, P. (eds) *The Valencia Trough: Geology and Geophysics*. *Tectonophysics*, 203, 21–36.

QUINLAN, G. M. & BEAUMONT, C. 1984. Appalachian thrusting, lithospheric flexure and the Paleozoic stratigraphy of the eastern Interior of North America. *Canadian Journal of Earth Sciences*, 21, 973–996.

ROWLEY, D. B. & SAHAGIAN, D. 1986. Depth-dependent stretching: a different approach. *Geology*, 14, 32–35.

ROYDEN, L. H. & HORVATH, F. (eds) 1988. *The Pannonian Basin: A Study in basin evolution*. American Association of Petroleum Geologists, Memoir 45.

SCLATER, J. G. & CHRISTIE, P. A. F. 1980. Continental Stretching: An explanation of the post-mid-Cretaceous subsidence of the central North Sea basin. *Journal of Geophysical Research*, 85, 3711–3739.

SMITH, W. H. F. & WESSEL, P. 1990. Gridding with a Continuous Curvature Surface in Tension. *Geophysics*, 55, 293–305.

THORNE, J. & WATTS, A. B. 1989. A Quantitative analysis of North Sea Subsidence. *American Association of Petroleum Geologists Bulletin*, 73, 88–116.

TORNÉ, M., PASCAL, G., BUHL, P., WATTS, A. B. & MAUFFRET, A. 1992. Crustal and Velocity Structure of the Valencia Trough (Western Mediterranean). Part I: A combined refraction/wide angle reflection and near vertical reflection study. *In*: BANDA, E. & SANTANACH, P. (eds) *The Valencia Trough: Geology and Geophysics*. *Tectonophysics*, 203, 1–20.

WATTS, A. B. 1988. Gravity anomalies, crustal structure and flexure of the lithosphere at the Baltimore Canyon Trough. *Earth and Planetary Science Letters*, 89, 221–238.

——& RYAN, W. B. F. 1976. Flexure of the lithosphere at continental margin basins. *Tectonophysics*, 36, 25–44.

——& TEN BRINK, U. S. 1989. Crustal Structure, Flexure and Subsidence history of the Hawaiian Islands. *Journal of Geophysical Research*, 94, 10473–10500.

——& TORNÉ, M. In press. Subsidence History, Crustal Structure and Thermal evolution of the Valencia trough: A young extensional basin in the western Mediterranean. *Journal of Geophysical Research*.

——, ——, BUHL, P., MAUFFRET, A., PASCAL, G. & PINET, B. 1990. Evidence for reflectors in the lower continental crust before rifting in the Valencia trough. *Nature*, 348, 631–635.

WEISSEL, J. K. & KARNER, G. D. 1989. Flexural uplift of Rift Flanks due to Mechanical Unloading of the Lithosphere during Extension. *Journal of Geophysical Research*, 94, 13919–13950.

WHITE, N. & MCKENZIE, D. P. 1988. Formation of the 'Steer's Head' Geometry of Sedimentary Basins by Differential Stretching of the Crust and Mantle. *Geology*, 16, 250–253.

ZEYEN, H. J., BANDA, E., GALLART, J. & ANSORGE, J. 1985. A wide angle seismic reconnaissance survey of the crust and upper mantle in the Celtiberian chain of eastern Spain. *Earth and Planetary Science Letters*, 75, 393–402.

ZOETEMEIJER, R., DESEGAULX, P., CLOETINGH, S., ROURE, F. & MORETTI, I. 1990. Lithospheric dynamics and Tectono-Stratigraphic Evolution of the Ebro Basin. *Journal of Geophysical Research*, 95, 2701–2711.

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