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Mechanical (de-)coupling of the lithosphere in the Valencia Trough (NW Mediterranean): what does it mean?

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Abstract

We study the mechanics of lithospheric decoupling in continental extensional basins in relation to the distribution of (non-)competent mechanical layers within the lithosphere and the position of the isostatic compensation level. We specifically address the different modes of deformation taking place in crustal levels according to a self-consistent formulation of the concept of mechanical decoupling. Subsequently, we investigate the style of lithospheric decoupling in the Valencia Trough (NW Mediterranean), a prime example of a young continental rift basin. During its evolution, the lower crust (or at least part of it) acted as a weak, non-competent layer that eventually flowed laterally to accommodate deformation in the subcrustal lithosphere and overlying crust, which became mechanically decoupled. We use a numerical model to discern whether these two layers deformed fully independently (vertical decoupling), or maintaining a mechanical link (horizontal and partial decoupling). Results of our study, constrained by a high-quality database, exclude fully decoupled mode and favor isostatic compensation level in the asthenosphere. Interpretation of our results in light of geological and geophysical data suggests that the present Valencia Trough is best described by partial lithospheric decoupling.

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1. Introduction

Since the late seventies, several mechanisms have been proposed for intraplate deformation and the formation of sedimentary basins. Following initial models of pure shear [1] and simple

shear deformation [2], conceptual ideas have been developed and refined to address the complexities of the mechanisms involved in the final configuration of basins. These include regional flexure of the lithosphere [3], effects of horizontal stress fields [4], thermal processes and different rheological approaches [5] and process-oriented dynamical models [6].

Relatively simple models of basin formation (one-layer elastic, flexural approach, coupled with thermal field effects) have successfully explained the evolution of oceanic basins (e.g.,

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[3,7]. In contrast, these models face difficulties in the case of continental lithosphere, requiring more assumptions (e.g., [5]).

The process of rifted basin formation implies consecutive stages of weakening of initially (moderately) strong lithosphere and lithospheric hardening after the cessation of the rifting cycle, due to the attenuation of the associated thermal anomaly. Also within the lithosphere, the upper crust, lower crust, lithospheric mantle, or any other individual, independent layer might respond in a different manner to the rifting process, depending on the initial geometry and rheological characteristics of the lithosphere and the rate of deformation (e.g., [8–10]). Such a mutually independent response of the various lithospheric layers is often referred as *decoupling* (e.g., [11]). However, different authors have significantly different views on the implications of this concept.

In this paper the issue of (de-)coupling of the lithosphere is addressed. In order to distinguish various interpretations of this concept, a more refined definition of it is proposed. To this aim, the mechanisms of formation of the Valencia Trough extensional basin (NW Mediterranean) are investigated using a numerical modeling approach.

2. Lithospheric coupling and decoupling

2.1. Ductile lower crustal zone

The traditional mechanical subdivision of the external part of the Earth distinguishes a rigid layer (*lithosphere*) overlying a weak, partially melted and low-viscosity layer (*asthenosphere*). For a given mineralogical structure of the upper mantle, the transition zone between these two layers is controlled by the position of the isotherms [12]. On tectonic time scales, the asthenosphere, subject to the weight of the overlying lithosphere (plate), deforms plastically, ‘flowing’ aside, in order to attain isostatic compensation of the lithospheric column.

In an analog rheological reasoning for continental crust, the occurrence of a mid-crustal transition zone in which crustal rocks change from a

brittle to a ductile deformation mode has been proposed (e.g., [13]). At temperatures higher than 300–450°C (depending on the composition, strain rate and geotherm), normally corresponding to lower crustal levels, rocks attain such a low viscosity that can actually creep-flow [12]. According to the conventional rheological models, the uppermost mantle constitutes a strong, more rigid zone confining this flow to lower crustal levels. The general lack of deep crustal seismicity [14], the low seismic velocities found [15] and the bending of faults that flatten out at the top of this level, have been usually pointed as indicators of such a low-viscosity zone [16]. Additionally, some authors argue that the semi-horizontal, lower crustal strong seismic reflectors observed in many areas represent ductile shear bands, which denote actual horizontal mass-flow expected within the low-viscosity zone [17].

The three-layered (strong upper crust, soft lower crust and strong upper mantle) rheological model described above has been questioned recently (e.g., [18,19]). Based on analyses of the seismogenic layer in continents [20] and using gravity and topography data as constraints, these authors propose a rheological model in which the whole crust is the strongest lithospheric layer and the lithospheric mantle is relatively weak.

2.2. Implications of a crustal decoupling level on the tectonic evolution of the lithosphere

Given the thermal and rheological conditions sustaining a lower crustal low-viscosity zone, the implications for the tectonic evolution of the continental lithosphere are important (e.g., [5,21]). Basically, the existence of such a zone between the stronger bounding layers enables them to react in a different manner to deformation, and implies an overall lithospheric strength reduction. This is what causes ‘*decoupling*’ of the lithosphere, and the lower crustal low-viscosity zone is the ‘*decoupling level*’. However, depending on the authors and the approach they have chosen the term *decoupling* has been given different meanings.

Some authors consider *decoupling* as the possibility of two (lithospheric) layers to move only laterally with respect to each other, interacting

mechanically by shearing in the intermediate, weak layer (Fig. 1b; e.g. [22]), facilitating differential stretching. The lithosphere presents a rather low strength (or equivalently, low values of effective elastic thickness, T_e) and the isostatic compensation level remains at the base of the lithosphere. This is what we will call in this paper *horizontal decoupling*.

Other authors have studied *decoupling* as the possibility of two layers to move independently of each other in the vertical sense (e.g. [23], see Fig. 1c). In this view, stresses exerted by the bounding brittle layers would not be transmitted vertically. In contrast, they would be dissipated in the low-viscosity layer, which would flow horizontally in order to compensate isostatic disequilibrium. We will refer to this mode as *vertical decoupling*. For this extreme situation for a continental extensional basin, it could be the case that, as lithosphere stretches, the upper crust could deform independently from the upper mantle.

More recently, ter Voorde et al. [24] and Waltham et al. [25] have proposed a *partially decoupled* mode of lithospheric deformation, which could be considered as a combination of horizontal and vertical decoupling. At first order approximation, the lithosphere is isostatically coupled. In this case, the viscosity of the lower crust is not low enough to produce significant lateral flow to accommodate all load unbalances. Only short-wavelength loads can be compensated in lower crustal levels, whereas long-wavelength loads are compensated in the asthenosphere.

The alternative rheological model with a weak mantle and a strong crust, with no distinction in terms of strength between upper and lower crust as proposed by Maggi et al. [18] and Jackson [19] explicitly considers that the lithosphere is not mechanically decoupled at lower crustal levels (Fig. 1a).

2.3. Isostatic coupling versus isostatic decoupling

Estimates on how the occurrence of a decoupling layer affects isostatic movements are obviously different depending on the concept that is used: each decoupling model locates the isostatic compensation level in a different place (astheno-

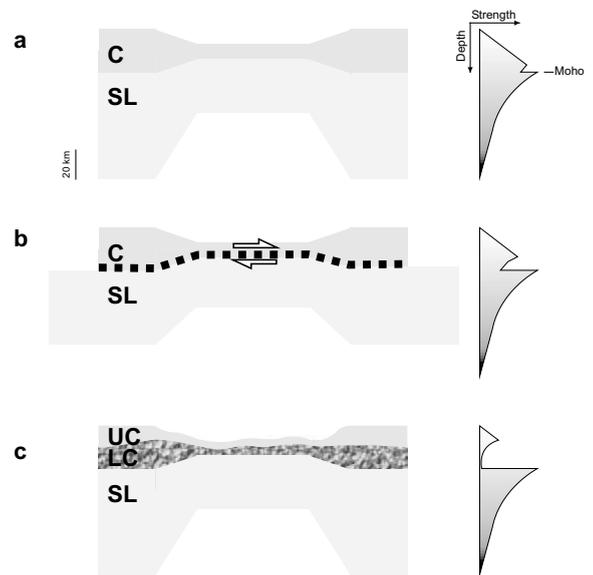


Fig. 1. Cartoon illustrating concepts of horizontal and vertical decoupling. Left: Geometry of the lithosphere. Dark and light gray represent crust and subcrustal lithosphere respectively. Right: Yield strength as a function of depth for brittle–ductile rheology of the lithosphere. C, crust; SL, subcrustal lithosphere; UC, upper crust; LC, lower crust. (a) Uniform stretching (fully lithospheric coupling). (b) Horizontal decoupling [22]. Thick dotted line and white arrows represent the level of relative shearing between the decoupled layers. (c) Vertical decoupling (after [23]).

sphere for horizontal decoupling and lower crust for total vertical decoupling). In this respect, when isostatic analysis is conducted, it can be appropriate to use the term *isostatic coupling* of the lithosphere when referring to a sublithospheric compensation level, and *isostatic decoupling* of the lithosphere when referring to an intralithospheric compensation level.

In an attempt to express with one parameter how the lithosphere would evolve in an extensional tectonic regime given a strength pattern, some researchers have introduced the concept of *lithospheric necking*. This is defined as the level that would remain horizontal as extension progresses in absence of isostatic forces [26,27], and it is related to the position of the strongest lithospheric layer(s) (more difficult to deform) at the beginning of the extension [28]. Thus, the level of necking is the bulk expression of a more detailed lithospher-

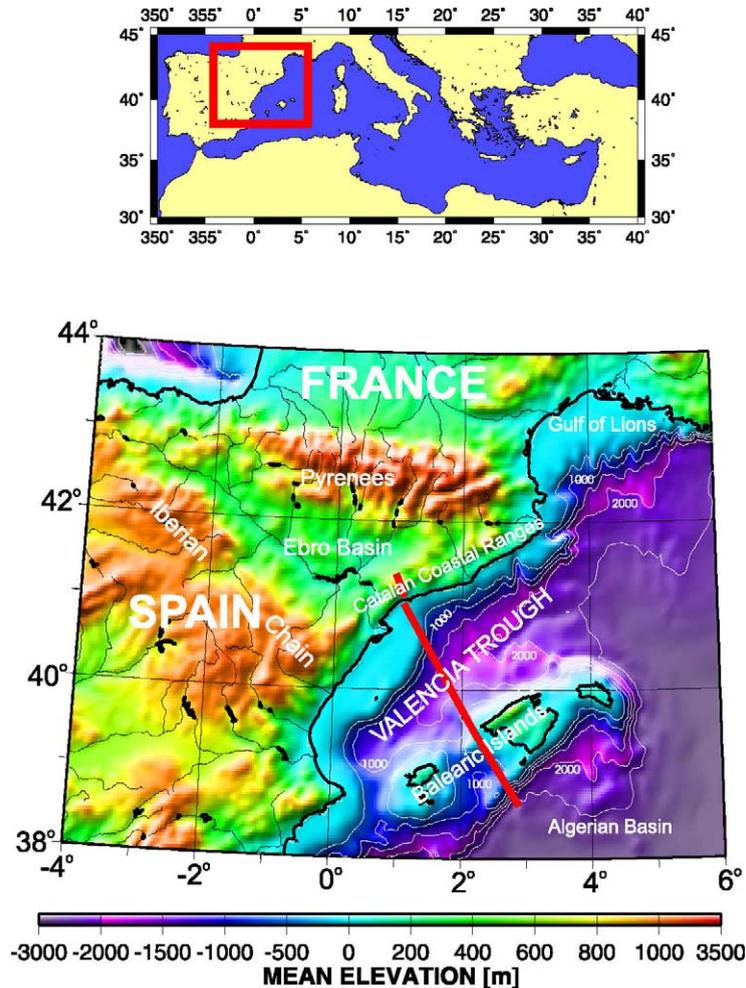


Fig. 2. Upper panel: Location of the study area offshore NE Spain, in the northwestern Mediterranean Sea. Lower panel: Topography and bathymetry map indicating location of ESCI-Catalan Valencia Trough deep seismic reflection profile.

ic mechanical structure, in which rheological properties (prior to and during rifting), relative position of detachment levels and strong/weak layers in the lithosphere, can lead to decoupling. So far, no one-to-one relationship between lithospheric decoupling and necking has been established, and both quantitative and qualitative valuations lack accuracy (e.g., [28]). However, from a modeling point of view, the depth of necking is a very useful parameter and illustrative when interpreted in light of geological and geophysical data.

In the following, the state of decoupling of the

lithosphere in the Valencia Trough is investigated using a numerical model that accounts for isostasy, necking of the lithosphere, differential stretching and thermal effects. The high-quality geological and geophysical data set available in this area offers the framework to test the concepts discussed above.

3. The Valencia Trough

The Valencia Trough is a Neogene extensional basin located offshore northeastern Spain, in the

northwestern Mediterranean (Fig. 2). Its axis is SW–NE oriented, and it is wider to the northeast (where it connects to the Liguro-Provençal Basin) than to the southwest. The Catalan Coastal Ranges and the Balearic Islands bound this basin to the northwest and southeast respectively. Its potential as an oil-producing area triggered many geophysical surveys that yielded a high-quality database. Significant parts of these studies have been summarized and reviewed in [29–32].

The rift basin started to open during Late Oligocene over an inverted Mesozoic basin that was subjected to non-deposition and erosion during most of the Paleogene [33]. It appears that the opening was not synchronous all over the basin [34], but that rifting commenced earlier in the northeast (Late Oligocene; 28 Ma) than in the southwest (Early Miocene; 22 Ma). In our study area, the syn-rift stage, in which fault activity was important, can be dated as Late Oligocene–Early Miocene (25–16 Ma) and the subsequent post-rift stage, with generalized thermal subsidence, lasted until the present day [32].

The present-day Valencia Trough is largely asymmetrical. Bathymetry increases from southwest to northeast. The sedimentary infill of the basin is Late Oligocene to Quaternary in age [35,36]. Syn-rift sediments are preserved mainly in the troughs of the basin and constitute a moderate amount of the total infill. Although post-rift sequences are abundant, the basin is partly sediment-starved [31], being constantly infilled by material eroded mainly from the flanking ranges (Catalan Coastal Ranges and Iberian Chain) and transported by bounding river networks (mainly the Ebro River). In the southeastern margin, Late Oligocene–Middle Miocene compressional tectonics led to the development of a WNW-verging fold-and-thrust belt which involved Paleozoic basement, Mesozoic cover and Paleogene to Middle Miocene sediments [33].

The rifting process developed in continental lithosphere and did not lead to the formation of oceanic lithosphere. However, the crust as well as the whole lithosphere were extremely thinned [30,37] and the thermal anomaly still remains significant, as demonstrated by the relatively high surface heat flow values measured [38].

4. Lithospheric (de-)coupling in the Valencia Trough

4.1. ESCI-Catalan and Valencia Trough seismic line

Our modeling is based on a deep seismic reflection profile that crosses the Valencia Trough from the Iberian Peninsula to the South Balearic Basin (ESCI-Catalan and Valencia Trough profiles, Fig. 3; [32]). This section is constrained by wide-angle data [39], boreholes [29], surface data and industrial seismic profiles [32,40,41]. It provides an excellent image of sedimentary packages filling the basin and of the crustal structure, with the exception of the axis of the basin.

A number of features from the seismic section (Fig. 3) are listed below. They are consistent with older data in other areas of the Valencia Trough [42–46], and therefore can be regarded as representative of the basin:

1. The crust dramatically thins from the Iberian margin towards the axis of the Valencia Trough, getting thicker again underneath the thrust sequences that form the Balearic Promontory. In some parts of the axial domain of the Valencia Trough, thinning is so severe that the lower crust might have disappeared [47] or be intruded by mantle rocks [48] as revealed by anomalous low mantle velocities and gravity models [45,49]. Thickening in the Balearic Promontory is due to thrust tectonics coeval with rifting [32,33].
2. The lower crust is thinned in a higher degree than the upper crust (including the Mesozoic cover). This feature would confirm some degree of mechanical decoupling (horizontal following the convention established in Section 2) between upper and lower crust [32,40,50,51].
3. Upper and lower crusts have markedly different reflectivity patterns. The upper crust is almost transparent whilst the lower crust is more reflective, showing manifest reflectivity variations along the profile. These observations would indicate a major rheological change between both crustal layers. Strong reflectors display lower crustal thinning from the Iberian mainland to the Valencia Trough [39]. Lower

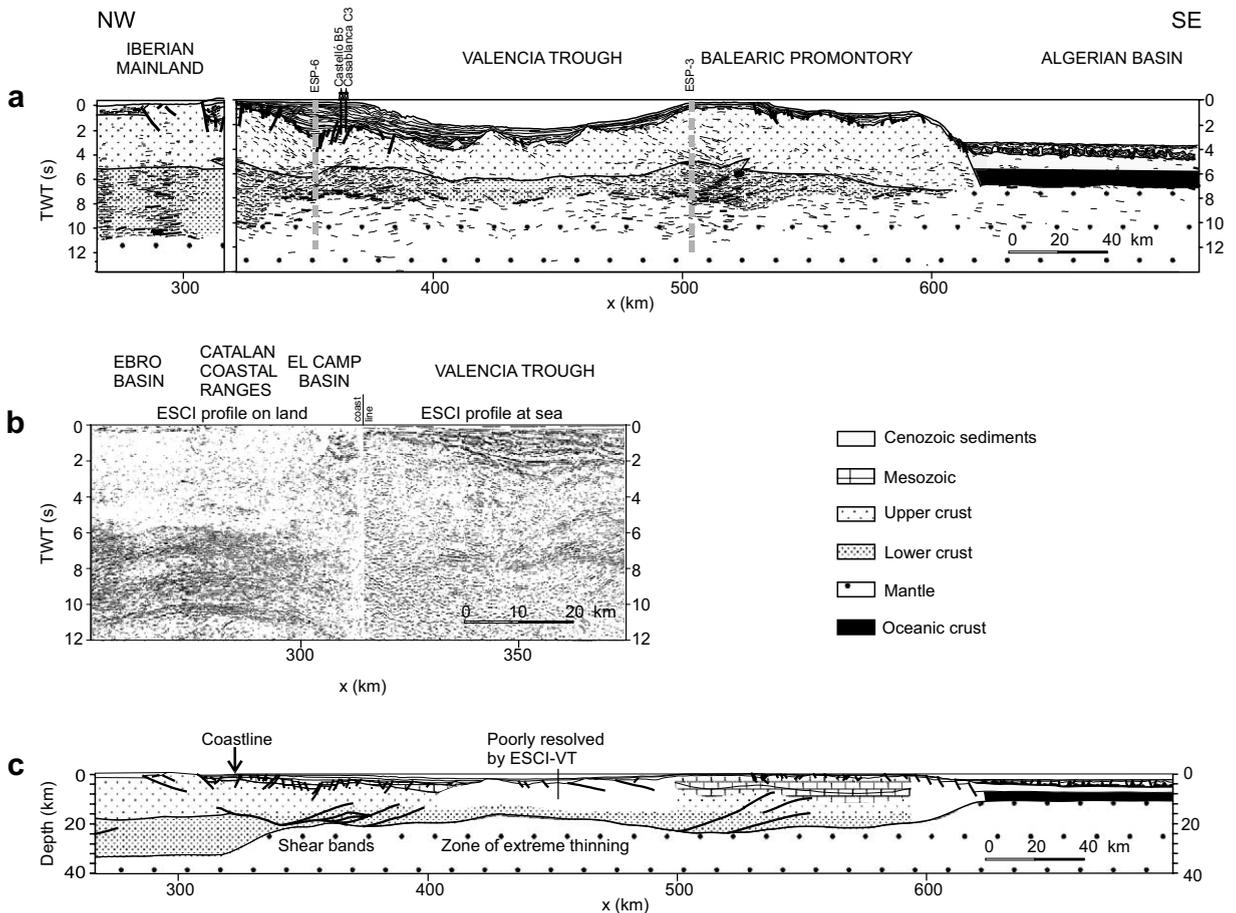


Fig. 3. ESCI-Catalan Valencia Trough seismic reflection profile [32]. (a) Two-way travel time interpreted line drawing and interpretation (after [40]). (b) Composite stacked section of the ESCI vertical reflection profiles on land and at sea [41]. (c) Depth conversion of the section according to Vidal et al. [39] and modified in the shallowest levels to precisely locate the base of Cenozoic sediments, as constrained from well data [29] and ESP data [44,45]. In this figure Mesozoic sediments and upper crust are differentiated. See text for further explanation.

crustal reflectivity diminishes (and eventually disappears) towards the center of the basin as a consequence of Late Oligocene–Early Miocene extension [46]. Beneath the Balearic Promontory, the lower crust partially recovers its reflective character.

4. Extensional faults are observed in the upper crust and not in the lower crust, indicating that the requirements (rheological properties) for brittle failure do not occur in the middle–lower crust. This is also supported by the (moderate) seismicity in the area, which is re-

stricted to upper crustal levels (focal depths ≤ 8 km; [52]). These upper crustal faults are responsible for the formation of secondary sub-basins (El Camp Basin, Tarragona Basin), but cannot account for the opening of the whole Valencia Trough [53]. There is no evidence of fault continuation towards the center of the basin in deep levels [39]. Therefore considering a simple shear mechanism for the development of this rift basin [47] seems to be unrealistic [31,54].

5. The central part of the profile and the upper

layers underneath the Balearic Promontory are not well constrained by the seismics. Comparison with studies carried out in adjacent domains within the Valencia Trough suggests that volcanic edifices could have been sited in crustal levels in the center of the basin, and perturbed the lithospheric rheological layering [50].

According to Sàbat et al. [40] the configuration of lower crustal strong reflectors in the transition from the stable Iberian crust to the Valencia Trough reveals the existence of shear bands that accommodated deformation during rifting. In the Balearic margin some (reverse) shear bands are observed in the lower crust as well [40]. These bands could be vestiges of lower crustal material flow.

4.2. Additional data: vertical decoupling in the Valencia Trough as working hypothesis

All the points described above have been used as evidence for crustal *horizontal decoupling* in the Valencia Trough by many authors [32,50,51,53]. Also, basin crustal-scale restorations [32,53] and evolutionary numerical models of the Valencia Trough [51,55,56] implicitly assume a *horizontal decoupling*. In the models that incorporate isostasy the compensation level is located in the asthenosphere (i.e., *isostatic coupling*). *Vertical decoupling* has not been considered in this area, though.

The Valencia Trough consists of a thermally

young lithosphere. Surface heat flow measurements give intermediate to relatively high values ($78 \pm 13 \text{ mW m}^{-2}$ in average, increasing locally up to 105 mW m^{-2} ; [38]), which are probably underestimated by 10–15% because they were not corrected by effects of recent sedimentation [49]. Numerical studies addressing the thermal structure of the lithosphere in the Valencia Trough constrain the range of expected lower crustal temperatures at present. Thermal modeling results by Fernández et al. [57] show Moho temperatures increasing from 400°C in the axis of the Valencia Trough to 600°C in the Catalan margin. Assuming a rheological stratification with a relatively weak diabase-diorite lower crust, Zeyen and Fernández [49] predict Moho temperatures varying from 350°C in the axis of the basin to more than 600°C in its flanks. Alternatively, Negredo et al. [51], whose model implicitly assumes horizontal decoupling between upper and lower crust, predict Moho temperatures for the present ranging between 450°C to more than 600°C in the flanks. These values were higher at the end of the rifting phase: over 550°C all along the Valencia Trough [51]. Strain rate values in the order of 10^{-15} s^{-1} have been estimated for this area [28,55]. At these relatively high temperatures and surface heat flow, the lower crust (or at least the hottest part of it) will deform in a ductile mode [13] and eventually will be able to flow; during the syn-rift phase all over the basin and in the post-rift phase in the margins. These surface heat flow measurements

Table 1
Model parameters

Initial thickness of the crust [62]	32 km
Initial thickness of the upper crust [32,62]	16 km
Initial thickness of the lithosphere [63]	100 km
Initiation of rifting [32]	25 Ma
Duration of rift phase [32]	9 Myr
Sediment density (grain density) (adapted from [55])	2600 kg m^{-3}
Crustal density (average) [55]	2825 kg m^{-3}
Upper crustal density (only for vertical decoupling) [55]	2700 kg m^{-3}
Compensating material density (idem) [55]	2950 kg m^{-3}
Density of the mantle [55]	3270 kg m^{-3}
Surface porosity (derived from [35,60])	0.75
Depth–porosity relation constant (derived from [35,60])	0.55 m^{-1}
Temperature at the base of the lithosphere [24]	1330°C
Thermal diffusivity [24]	$7.8 \times 10^{-7} \text{ m}^{-2} \text{ s}^{-1}$
Thermal expansion coefficient [24]	$3.4 \times 10^{-5} \text{ }^\circ\text{C}^{-1}$

and thermal structure estimates support the idea of vertical decoupling in lower crustal depths. The moderate value of the strain rate prevents any a priori consideration in this respect.

Below, we investigate the role of (*isostatic*) *vertical decoupling* in this region. Provided that all the evidence points to the existence of a crustal decoupling level, the question is: would the lower crust be able to flow aside and become the isostatic compensation level for flexural calculations?

5. Numerical modeling

5.1. Modeling approach

A numerical model that takes into account flexure of the lithosphere, lateral heat transport and lithospheric necking [58] is used. Basin subsidence and flank uplift are simulated by setting stretching factors and corrected for sea level oscillations (sea level curve from [59]) and sediment compaction effects (based on [60]). Other parameters are given in Table 1. Effective elastic thickness (T_e) and depth of lithospheric necking (z_n) are free parameters, adjusted to fit the observations.

Three cases are investigated:

1. *Uniform stretching* of the lithosphere. The lithosphere is fully coupled, deforming as a whole with no intermediate layering over a creep-flowing asthenospheric compensating material. This was the conventional approach followed by Janssen et al. [56].
2. *Horizontal decoupling* between crust and lithospheric mantle (i.e., different stretching factors for crust and lithospheric mantle), and *isostatic coupling* with the compensation level located in the asthenosphere.
3. Fully *vertical decoupling* between crustal layers (uniform stretching of upper crust) and *isostatic decoupling* with a lower crustal compensation level.

The case of partial decoupling has not been tested because this would require an extra parameter (the degree of decoupling) that we are not able to constrain. Moreover, the predicted rift basin geometry resulting from a partially decoupled scenario is hard to distinguish from the

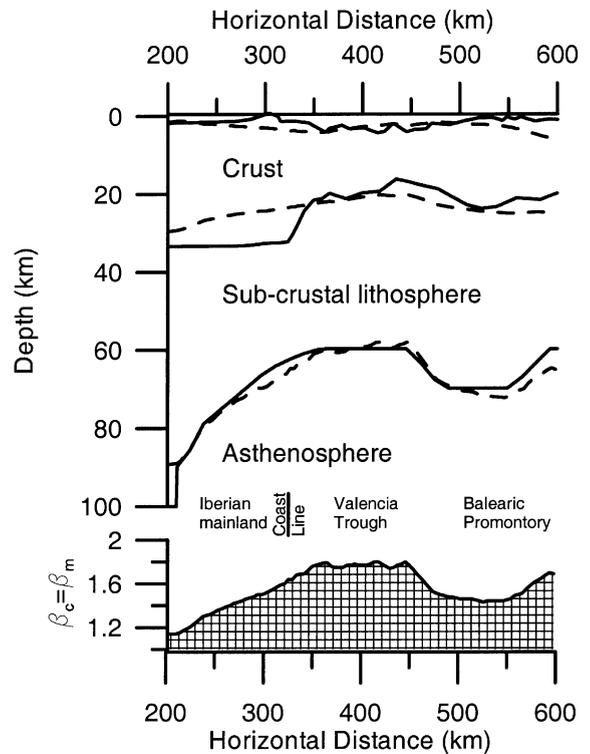


Fig. 4. Results for fully coupled case (uniform stretching; $\beta_c = \beta_m$; as in Fig. 1a). Upper panel shows lithospheric structure. Continuous lines represent observed horizons and dashed lines calculated horizons. Lower panel shows lithospheric thinning factors (β_c and β_m), derived by dividing pre-rift average lithospheric thickness (see Table 1) by present-day lithospheric thickness (taken from Fig. 3c and [37]).

one resulting from the isostatically coupled (horizontal decoupling) case [24].

5.2. Modeling results

The simplest case of complete lithospheric coupling (Fig. 4) gives a reasonable fit in those areas where the calculated horizons (i.e., base of Cenozoic sediments, base of the crust and base of the lithosphere) are smooth. In contrast, it fails to reproduce the shape of those horizons in areas of high gradient, as the transition from the Iberian crust to the Valencia Trough beneath the Catalan Coastal Ranges and the areas of extreme thinning in the center of the basin.

Fig. 5 displays results for the *horizontal decoupling* scenario. Despite some minor short-wave-

length deviations, there is a good fit with observations. Best fit is obtained for low T_e values (5 km) and intermediate depths of necking ($z_n = 15$ km). These values are in accordance with previous modeling studies predicting low rigidities [55,56], but situate the level of necking at shallower depths than Janssen et al. [56] (who proposed z_n values between 17 and 33 km). These authors, who assumed uniform stretching of the whole lithosphere and ignored differential stretching of the subcrustal lithosphere, needed a deeper necking level in order to obtain substantial flank uplift fitting the topography of the CCR. In turn, as part of the topography of the CCR predicted in our model results from the onshore thinning of

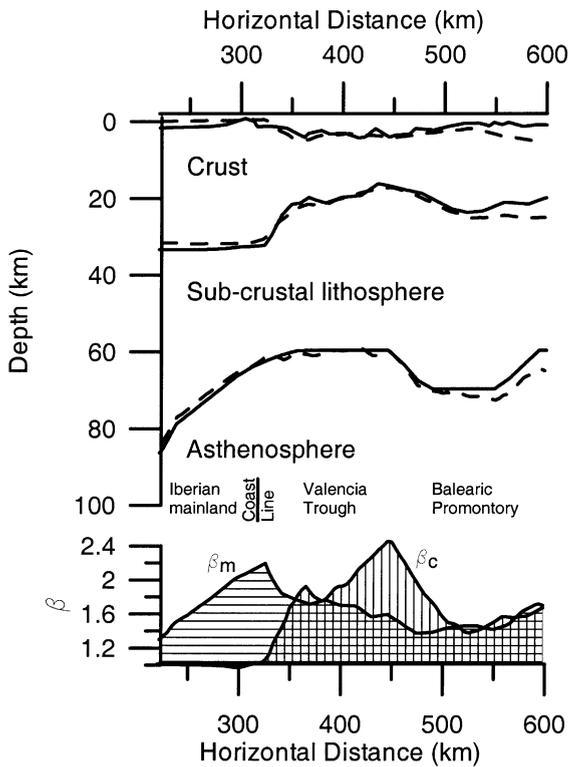


Fig. 5. Results for vertically coupled and horizontally decoupled case ($\beta_c \neq \beta_m$; as in Fig. 1b). Upper panel shows lithospheric structure. Continuous lines represent observed horizons and dashed lines calculated horizons. Lower panel shows thinning factors (β_c for the crust and β_m for the sub-crustal lithosphere), derived by dividing pre-rift average crustal and subcrustal lithospheric thicknesses (see Table 1) by present-day crustal and subcrustal lithospheric thicknesses (taken from Fig. 3c and [37]).

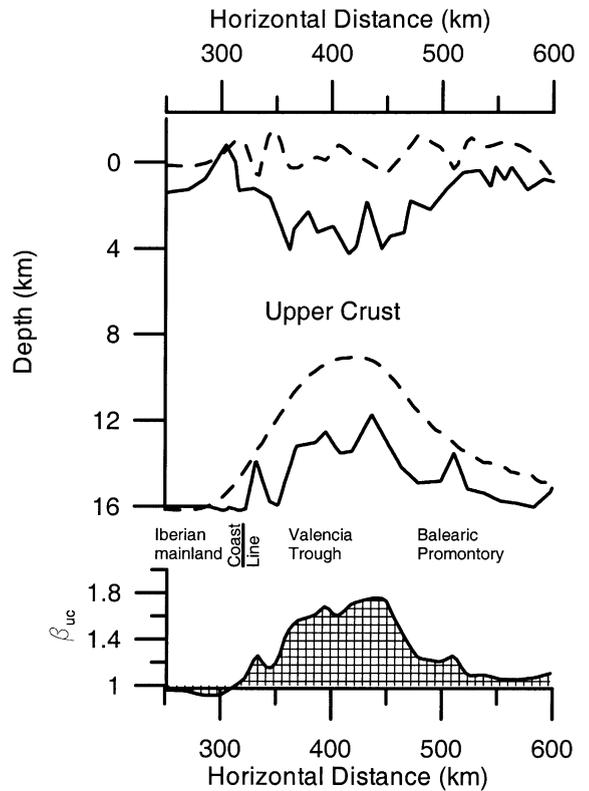


Fig. 6. Results for vertically and horizontally decoupled case (as in Fig. 1c). Upper panel shows lithospheric structure. Continuous lines represent observed horizons and dashed lines calculated horizons. Lower panel shows upper crustal thinning factors (β_{uc}), derived by dividing pre-rift average upper crustal thickness (see Table 1) by present-day upper crustal thickness (taken from Fig. 3c).

the subcrustal lithosphere, we do not require such a deep level of necking. Summarizing, our modeling predictions favor intermediate to shallow levels of necking instead of deep ones [61].

Here it is worth nothing that a similar horizontal decoupling model, but with the decoupling level located in mid-crustal levels (i.e., decoupling between upper crust and the rest of the lithosphere) was tested as well, yielding unsatisfactory results.

Results for *vertical decoupling*, focussing on upper crustal levels, are shown in Fig. 6. It is found that any reasonable combination of modeling parameters does not give a satisfactory result. There is always too little subsidence predicted, and in general upward buoyancy forces are

more significant than downward loading forces. Due to the low density of the lower crust compared to the asthenosphere, (isostatic) total vertical decoupling does not cause enough flexural subsidence of the upper crust.

5.3. Discussion

The best fitting T_e value (5 km) found in our modeling corresponds to a low flexural rigidity ($D \approx 8 \times 10^{20}$ N m), that is associated to a weak lithosphere. Furthermore, the best fitting intermediate level of necking ($z_n = 15$ km) denotes that the main contribution to the overall lithospheric strength is not rooted in the mantle. We relate this value with a strength maximum located in the base of the upper crust that controlled lithospheric deformation at the beginning of the rifting. This idea is supported by the fact that this level coincides with the depth of detachment in which master faults, controlling the opening of marginal sub-basins (Barcelona, Vallès-Penedès, El Camp, Tarragona), flatten out. That the depth of necking (indicative of the strength of the lithosphere at the initiation of the rifting) is lower than the T_e value (representative for the present-day strength of the lithosphere) reveals lithospheric weakening as a consequence of the rifting process. As lithospheric stretching progressed, the temperatures in lower crustal levels significantly increased [51]. This might have triggered lower crust flow leading to a vertical decoupling scenario. Post-rift thermal relaxation and intense thinning of the flow channel in the lower crust (both more pronounced in the axis of the basin) progressively destroyed this mechanism. This means that, at present, the viscosity of the lower crust is not low enough to prevent stress transmission between upper crust and subcrustal lithosphere, and the isostatic compensation level is located in the asthenosphere. This is confirmed by the fact that the model that best explains the present-day geometries is horizontal decoupling (and not vertical decoupling as could be interpreted from the low T_e value predicted).

Taking into consideration the geological and geophysical data as well as previous modeling studies, we interpret our modeling results as re-

vealing that the Valencia Trough lithosphere responds to a partial decoupling deformation mode. In such a scenario, crust and subcrustal lithosphere are stretched to a different degree, the level of isostatic compensation for long-wavelength features is located in the asthenosphere and only short-wavelength features are compensated by (very limited) lower crustal flow. This last point could resolve the failure of the horizontal decoupling model to fit short-wavelength geometrical features (Fig. 5 and Section 5.2).

In contrast with recent studies contesting the ‘jelly sandwich’ rheological model [18,19], this study supports the occurrence of mechanical decoupling in the Valencia Trough.

6. Conclusion

The example of the Valencia Trough shows the necessity for unifying criteria when using the term (*de-*)*coupling*, specially in cases in which analyses of data and modeling results can be easily misinterpreted. At first sight, a number of observations suggest that the lithosphere in the Valencia Trough could have developed according to a fully isostatic vertical decoupling mode of deformation. Some predicted values for modeling parameters, such as the low T_e and the intermediate depth of necking, can be interpreted as consistent with this hypothesis. However, the model considering *horizontal decoupling* best fits the observed geometries. In order to integrate these observations and modeling results in a consistent manner, we suggest that the *partially decoupled* model, combining low flexural rigidities and sublithospheric isostatic compensation level, constitutes the best mechanical description of the Valencia Trough lithosphere at present.

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